MAGNETITE DEPOSITS OF EL ROMERAL, CHILE: 
PHYSICAL GEOLOGY, SEQUENCE OF EVENTS, AND 
PROCESSES OF FORMATION

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FOR THE DEGREE OF 
DOCTOR OF PHILOSOPHY

By
Arthur Albin Bookstrom

March 1975
I certify that I have read this thesis and that in my opinion it is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

[Signature]
(Principal Advisor)

I certify that I have read this thesis and that in my opinion it is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

[Signature]

I certify that I have read this thesis and that in my opinion it is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

[Signature]

Approved for the University Committee on Graduate Studies:

[Signature]
Dean of Graduate Studies
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Arthur Albin Bookstrom 
Stanford University, 1975

El Romeral iron mine is in the coastal iron province of 
northern Chile, a north-trending belt of magnetite deposits 
about 30 km wide and 600 km long.

Iron at El Romeral is mined from two orebodies in which 
magnetite is microscopically intergrown with actinolite. The 
Main orebody is a lenticular, steeply dipping magnetite-rich 
mass in actinolitized andesite porphyry. The North orebody 
consists of conformable pods of magnetite in actinolitized 
biotite schists with steeply dipping foliation. Both ore-
rebodies and many smaller pods of magnetite trend northerly, 
adjacent and subparallel to the eastern border of a diorite 
pluton, and within a north-trending network of anastomosing 
strike-slip faults

Ore-zone boundaries are gradational, and altered rocks 
within and around the orebodies contain magnetite, actino-
lite, oligoclase, diopside, clinozoisite, sphene, chlorapa-
tite, scapolite, tourmaline, chlorite, pyrite, calcite, micas, 
and clays. Diorite west of the orebodies has been altered 
to aplitic quartz-feldspathic rock containing various pro-
portions of oligoclase, quartz, albite, and microcline.

The schists, phyllites, and quartzites of El Romeral 
probably are late Paleozoic in age, whereas the andesite 
porphyry, diorite, and magnetite deposits are early
Cretaceous in age. Ore deposition followed diorite crystallization and was contemporaneous with strike-slip faulting and emplacement of intramineral dioritite dikes.

It has been variously suggested that Chilean magnetite deposits were formed by magmatic injection, by deuteric release of iron from granodiorites, by remobilization of iron from metasedimentary rocks, and by hydrothermal replacement. It is unlikely that the Romeral deposits crystallized from ore magmas, as intimate association of low-Ti magnetite with actinolite and secondary oligoclase suggests transitional greenschist- to amphibolite-facies metasomatism at submagmatic temperatures in the range of 550-475°C. It also is unlikely that the Romeral deposits were formed by deuteric release of iron during quartzo-feldspathic alteration of diorite. The quartzo-feldspathic rocks are largely post-ore in age, and the amount of iron that could have been derived by such alteration is insufficient, probably by a factor of at least 10, to account for the iron added to rocks of the ore and actinolitic zones. Nor is it considered likely that iron was leached from local hematitic quartzites and then redeposited to form the orebodies.

Iron probably was initially mobilized as a constituent of dioritic magma generated at 80-100 km depth by partial melting of basaltic and/or peridotitic materials subducted beneath the western edge of the South American continent. Relatively unaltered rocks of the Romeral diorite pluton, which probably crystallized from such a magma, are less iron-
rich than average andesites. This suggests that the parent materials of the diorite magma were not uncommonly iron-rich, and that little iron was assimilated during rise of the magma to the hypabyssal environment.

During magmatic crystallization, iron, water, and other ore-forming constituents such as magnesium, phosphorous, chlorine, sulfur, copper, and boron, evidently were concentrated in the residual liquid by progressive crystallization of early plagioclase, later augite, and minor ilmenite. Intramineral dioritite dikes, which contain abundant primary and secondary magnetite and amphibole, probably tapped iron- and water-enriched residual magma that is inferred to have existed below El Romeral during the time of ore deposition.

Hydrothermal concentration of iron began when pressure relief, probably caused by strike-slip faulting, allowed exsolution of aqueous fluids from the residual magma at temperatures near 1000°C. Iron partitioned into these fluids and probably diffused upward through a fluid-filled network of fractures and pores to sites of deposition at temperatures mainly in the range 550-475°C.

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By ________________________________ Dean of Graduate Studies
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INTRODUCTION

romeral (rōm ě rál), n., m.
(Span., Chile) A hilly, rocky
semi-desert countryside with
scattered low bushes and cacti.

Location

El Romeral iron mine, in the coastal foothills of the
Andes of Chile, lies north of the fertile Central Valley and
south of the Atacama desert at 29°43' S, 71°15' W (Fig. 1).
It is about 9 km east of the Pacific coast and is separated
from the coast by two rugged north-trending ridges, Cerro
Juan Soldado (1200 m) and Cordón La Liga (900 m). The
deposit extends along the east slope of Cordón La Liga. Its
pre-mining outcrop, at altitudes of 350 to 450 m, lay about
150 m above the valley floor of Quebrada Romeral, an inter-
mittent braided stream that drains southeastward about 14 km
to Coquimbo Bay (Figs. 2 and 3).

The city of La Serena, 20 km south of El Romeral by
road, is near the mouth of the Elqui River, which flows from
the high Andes to Coquimbo Bay. The port city of Coquimbo
is about 9 km southeast of La Serena, and the iron-ore load-
ing facilities at Puerto Guayacán are in a small, protected
bay just south of Coquimbo, about 38 km from the mine by
railroad.

Climate

Temperatures at El Romeral vary between 5° and 25°C,
and average annual rainfall is about 100 mm (Melo, 1965).
During fall and winter, morning coastal fog is common. Most of the moisture that supports the scant vegetation comes in the form of mist or light drizzle. Spring and summer weather is generally warm and dry, but every few years a cloudburst causes flash floods.

Mining History

The first mining claims at El Romeral were staked in 1903 for copper. Most of the claims that cover the present mine were located in 1911 by Rodolfo Wedeles and Juan Monreal, and adjacent ground subsequently was located by others (L. F. Kuhnhenn, personal communication, 1972).

In 1912 Bethlehem Steel Co. obtained a 30-year lease on the high-grade El Tofo iron mines, about 30 km north of El Romeral, and a new company, Bethlehem Chile Iron Mines Co., was formed to mine the ore (Opec, 1968). In 1936, when El Tofo was in full production, rights to the similar but apparently smaller El Romeral deposit were acquired from a German firm, Muller and Guttenoffnungshitte (L. F. Kuhnhenn, personal communication, 1972). At that time, trenching and tunneling had been done to sample the near-surface ore, but no further exploratory work was attempted until the main Tofo orebody was approaching depletion.

In 1959 several deep drill holes showed that the Romeral magnetite deposits have considerable vertical extent. Magnetometer surveys made by Antonio Pena during the following year indicated that these deposits, then partly covered
Figure 1. Index map.
Figure 2. El Romeral iron mine, as seen by looking eastward from the top of Cordón La Liga, in 1970. Geological contacts are traced in ink.
Figure 3. The Main pit and Cordón La Liga, as seen from the east rim of the Main pit, in 1970. Geological contacts are traced in ink.
by fanglomerate, are more extensive laterally than previously had been thought. From 1960 to 1968 diamond drilling supervised by Pena disclosed about 200 million tons of ore averaging about 50 percent iron. The drill cores were geologically logged by Antonio Pena and Oscar Melo.

Production from the martite zone of the Main orebody began in 1955. Production from the North orebody began in 1960 and continued through 1969, when the deposit was worked out to the 400 m level. A beneficiating plant with a capacity of 1000 tons per day was completed in 1963. It includes a crusher, magnetic separator, and conveyor system (Fig. 4).

By 1970 the Main orebody had been mined to the 310 m level. The stripping rate had been increased through installation of a new power plant and use of new drills, shovels, and trucks, and it was planned to increase the mining rate upon completion of a new and larger mill then under construction. The intended production rate at that time was 4.5 million tons of concentrates per year over a projected mine life of 18 years. In 1971, after the Chilean national elections of 1970, Bethlehem Chile Iron Mines Co. was obliged to sell El Romeral mine to the Chilean Steel Co. (CAP).

Previous Geologic Work

During 1964 and 1965 a 1:5000 topographic and geologic map of the Romeral deposit and a surrounding 13.4 sq km area was made by Oscar Melo, mine engineer, and a 27-page geologic
Figure 4. Oblique aerial photograph of the Romeral beneficiating plant, 1970.
report was prepared to accompany the completed map. In 1968 Erik Klohn made a systematic study of fractures for purposes of pit design, and in 1970 this study was updated by Francisco Franquesa.

Three 15-minute quadrangles near El Romeral have been mapped by geologists of the Instituto de Investigaciones Geológicas de Chile. Cuadrángulo Lambert (La Serena) (Aguirre and Egert, 1970), Cuadrángulo Quebrada Marquesa (Aguirre and Egert, 1965), and Cuadrángulo Vicuña (Dedíós, 1967) were mapped at 1:50,000 scale. The larger area of Hoja Ovalle (Thomás, 1967) was mapped at 1:250,000 scale. These maps and the accompanying descriptions, along with the 1:1,000,000 Geologic Map of Chile (Ruiz and Corvalán, 1968), and the book Geología y Yacimientos Metalíferos de Chile (Ruiz and others, 1965), have provided a regional framework for the geology of El Romeral.

Objectives and Scope

The present work began with detailed field mapping of the rocks in the district, and with study of their structural arrangements and age relationships. It continued with mineralogical, petrologic, and geochemical investigations in the laboratory. Products of igneous differentiation were studied by comparing mineralogical and chemical compositions of sequentially emplaced rocks. Products of thermal metamorphism and hydrothermal alteration were studied by comparing mineralogical and chemical compositions of fresh and
altered rocks. To assess the results of chemical redistribution semi-quantitatively, an approximate accounting of chemical gains and losses was made for the rocks in each of the alteration zones.

Physical boundary conditions for processes of igneous differentiation and ore formation were estimated on the basis of stratigraphic considerations and experimentally determined stability ranges for observed mineral assemblages. Unsuccessful attempts were made experimentally to simulate magnetite-apatite ore magmas at temperatures suggested by the Romeral mineral assemblage, as well as vapor-phase transport of magnetite and apatite down a temperature gradient under subsolidus conditions.

Of the several hypotheses previously offered to explain the genesis of Chilean magnetite deposits like El Romeral, none was found to be completely consistent with the observations made at El Romeral. A new model, involving magmatic and hydrothermal concentration of iron, was then proposed.

Summary of Pertinent Literature

Iron deposits: Ruiz and Erickson (1962) classified the iron deposits of Chile into three general types: (1) banded metasedimentary quartz-magnetite deposits (Ruiz, 1965; González-Boronino, 1970), (2) contact metasomatic deposits (Ruiz and others, 1965), and (3) magnetite flows (Park, 1961).

Deposits of the contact metasomatic type have been economically the most important to date. Ruiz and others
(1965) listed and briefly described more than 40 iron deposits with known reserves ranging from half a million to one hundred million tons of high grade ore. Fourteen apatite-magnetite deposits found in the same region, commonly near the magnetite deposits, also were listed and briefly described.

Ruiz and others (1965) pointed out that the magnetite and apatite-magnetite deposits define a mineralized belt, about 30 km wide, that extends for 600 km between latitudes 26°S and 32°S (Fig. 5). They also gave rather detailed descriptions of six major deposits, El Algarrobo, El Romeral, Cerro Imán, Adrianitas, Huantéme, and Boquerón Chañar.

Geijer (1931) visited El Algarrobo and El Tofo, classified them as "Kiruna type," and inferred that they had been formed by magmatic segregation and filter pressing. He supposed that the andesitic host rocks of the iron deposits at Algarrobo are younger than the surrounding granodiorite into which they were intruded as the lower part of a dike-like volcanic feeder. He suggested that a residual magma of iron oxides, developed at greater depths, was later intruded into the andesite to form the ore, and that this intrusion was accompanied by amphibolitization of the andesites (Ruiz, 1967).

Brüggen (1934) recognized that the andesitic host rocks at El Algarrobo are older than the granodiorite, and that they constitute a roof pendant now in fault contact with the adjacent granodiorite. He also noted that the iron ore is
Figure 5. The Chilean iron province (after Ruiz and others, 1965, parte II, Fig. 4).
concentrated on the western side of this pendant, adjacent to an iron-poor "aplitic" marginal facies of the granodiorite west of the orebody. Brüggen suggested that the "aplite" had been derived from granodioritic magma by gravitational settling of its ferromagnesian constituents. He further inferred that these iron-bearing constituents had been concentrated at depth and later injected into the andesitic roof pendant to form the orebodies (Ruiz, 1967).

Ruiz (1943, 1965) recognized that the magnetite and associated minerals at El Algarrobo were emplaced predominantly by fracture filling and metasomatic replacement. He attributed the replacement and vein filling to a pneumatolytic or in part hydrothermal phase derived from the granodiorite magma and composed of H₂O, CaCl₂, Cl₂, NaCl, KCl, FeCl₃, SiCl₄, PCl₄, B₂O₃, H₂S, and CO₂. He further suggested that this fluid probably separated when the magma was completely liquid or had only a small percentage of crystals. This idea was based on the observed concentration of iron in intruded and metamorphosed volcanic rocks, rather than in the intrusive rocks.

Ruiz and others (1968) reported that "aplitic" rocks of the kind noted by Brüggen are present in most of the important Chilean iron deposits, but that gradational contacts and microscopic textural relationships indicate that these rocks are in fact altered and bleached derivatives of fine-grained granodiorite. They suggested that deuteric alteration produced the bleached rocks, that replacement of
ferromagnesian minerals by quartz and feldspar released large quantities of iron, magnesium, and calcium, and that metallic ions dissolved in the deuteric volatiles migrated toward the intruded rocks, where magnetite was deposited to form the orebodies. The inferred process is similar to that proposed by Mackin and Ingerson (1960) to explain the origin of the iron deposits of Iron Springs, Utah.

In an excellent overview of iron deposits of the Pacific Basin, Park (1972) observed that around the Atlantic and Indian Oceans, the borders of large, stable continental shields are underlain by sedimentary itabirites or taconites of Precambrian age, whereas deposits usually attributed to replacement processes predominate around the Pacific "ring of fire." Although exposures of Precambrian rocks around the Pacific are few, Park pointed out that some of them contain showings of iron formation, and he suggested that buried Precambrian iron formation may be more common around the Pacific than is generally supposed. He further observed that most of the replacement-type deposits are associated with faults, many of which parallel the borders of the Pacific basin, and that insofar as the peripheral fractures and deformed zones around the continents are expressions of plate tectonics, the associated ores also may be considered as products of sea-floor spreading.

Park envisioned a process of partial mobilization of iron in buried Precambrian iron formations and other iron-rich rocks, and indicated that "Such mobilization would be
effective along deforming continental margins where intrusives and their companion mobile fluids are working their way up through iron-rich materials." The molten magnetite or "ore magma" thus formed, in places charged with large amounts of volatile and other highly mobile constituents contributed or activated by the magma, would rise along faults and along the borders of the intrusions until an environment favorable to deposition was reached. Park also pointed out that the present character of the resulting deposit would depend not only on the nature of the transporting fluids, but also on the environment of deposition.

"Injection of magnetite magma appears to have been especially common where movement to the surface has been rapid, as in areas of explosive volcanic activities. And where skarns abound, temperatures and pressures were probably high, and communication with the surface was impeded."

Sillitoe (1972) noted that in the orogenic belts of North and South America, metal zones are aligned approximately parallel to continental margins, generally in the following sequence from west to east: Fe; Cu (with some Au and Mo); Pb, Zn, and Ag: and in some regions Sn or Mo (Fig. 6). He attributed this pattern to the release of metals from basaltic oceanic crust and pelagic sediments during partial melting at progressively deeper levels on subduction zones dipping eastward beneath the continent. The metals, he suggested, ascended as components of calc-alkaline magma.
Figure 6. A diagrammatic representation of the generalized sequence of metal provinces in western America in the context of a plate tectonics-subduction model (after Sillitoe, 1972, Fig. 1, p. 814).
Theoretical and experimental studies: Because many metasomatic magnetite deposits are closely associated with igneous rocks of intermediate composition, it is important to review previous theoretical and experimental studies on the origins of intermediate igneous rocks; on the behavior of iron in magmas and in hydrothermal fluids; and on stability fields of iron-bearing minerals in magmatic and hydrothermal environments.

Dioritic and/or andesitic magmas have variously been regarded as hybrid magmas resulting from mixing of basaltic and rhyolitic magmas, as basaltic magmas contaminated by assimilated sialic crustal material, as products of fractional crystallization of basaltic magmas, and as products of partial melting of lower-crustal and/or upper-mantle materials.

Because the hybrid-magma hypothesis is inadequate to explain the wide range of known igneous mineral assemblages, mingling of magmas is no longer considered to be a principal factor in magmatic evolution (Turner and Verhoogen, 1960, p. 88). Furthermore the great volumes of andesitic material in island arcs where earlier sialic material is lacking preclude sialic crustal contamination of basalt as a general explanation for the origin of andesites.

From previous experimental work, Bowen (1928) explained the origin of andesites by a hypothesis of fractional crystallization of basaltic magma at crustal pressures. He emphasized early crystallization of ferromagnesiaminerals,
resulting in residual concentration of alkalies and silica. However, Fenner (1929) pointed out that basaltic magmas more commonly differentiate toward iron enrichment and silica depletion than toward andesitic or granitic compositions.

Osborn (1959, 1962) showed experimentally that oxygen pressures could have a profound effect on the behavior of iron and on differentiation trends of basaltic magmas. If oxygen pressure remains at a high level during fractional crystallization, iron-bearing oxides begin to form at high temperatures and continue to crystallize throughout the consolidation of the magma. Thus the residual magma becomes progressively depleted in iron and enriched in silica - the Bowen or calc-alkaline trend. But if initial oxygen pressure is lower, and decreases with falling temperature, then ferrous silicates crystallize rather than iron-bearing oxides, and the residual magma becomes enriched in iron and depleted in silica - the Fenner or tholeiitic trend. Osborn (1959) further pointed out that calc-alkaline magmas generally have higher water contents than tholeiitic magmas, and he suggested that high water content may be the principal factor in causing high pO₂, which in turn should cause iron depletion and silica enrichment. This model seems to explain how water-rich, iron-poor residual magmas might form, but it seems to preclude the evolution of water-rich, yet iron-rich residual magmas, such as those from which the intramineral dikes of El Romeral are inferred to have formed.

Hamilton and Anderson (1967) calculated pO₂ curves for
crystallizing basaltic magmas containing water and concluded that unless hydrogen diffusion could be shown to be important, the probability of constant or increasing $pO_2$ is small. Without buffering, $pO_2$ should drop at least two orders of magnitude during crystallization, largely because of temperature decrease. Thus the conditions required to produce iron depletion and silica enrichment by the Osborn mechanism are probably not commonly attained in nature.

This conclusion was substantiated by Eggler and Burnham (1973), who did water-unsaturated melting experiments on andesites, and thereby determined melting temperatures for constituent minerals at various water fugacities, oxygen fugacities, and total pressures. They found that neither ilmenite nor magnetite is stable near the silicate liquidus under theoretically reasonable conditions, and therefore they suggested that neither ilmenite nor magnetite can play a major role in fractionation.

On the basis of rock melting experiments at high pressures, Green and Ringwood (1969) proposed two complementary hypotheses for derivation of the calc-alkaline rock suite from the mantle by progressive partial melting or fractional crystallization of hydrous basalt or amphibolite at 15-80 km depth to produce dioritic magmas and/or by partial melting of quartz eclogite at 80-150 km depth to produce granodioritic magmas.

Eggler and Burnham (1973) concluded that partial melting in the upper mantle, either of subducted basaltic ocean
floor or of peridotite, is a more reasonable mechanism for generation of andesites than is fractionation of basalts by plagioclase-pyroxene-oxide phase subtraction.

From an experimental study, Fischer (1950) determined the melting temperature for pure magnetite at about 1620°C, and the melting temperature for pure apatite at 1650°C. He also reported lower melting temperatures for magnetite-apatite mixtures, as well as fields of immiscibility involving iron oxide-, phosphate-, and silica-rich liquids.

Philpotts (1967) found that mixtures of magnetite, diorite, and fluorapatite, containing apatite in excess of 30 percent, form three immiscible liquids on melting: (1) apatite-rich, (2) magnetite-apatite, (3) silicate. He approximately located a eutectic at 25 ± 3 weight percent of apatite and at a temperature of 1410 ± 10°C.

Gibbon and Tuttle (1967), working in the system FeO-Fe₂O₃-SiO₂-H₂O, found that two liquids exist over much of the compositional range studied: (1) a silica liquid that quenches to a low-iron silica glass, and (2) an iron-rich liquid that quenches at high fO₂ to a mixture of fayalite and magnetite or cristobalite. At 1120°C and 2000 bars, the iron-rich liquid may approach a composition of 80 percent magnetite, 20 percent silica, and this liquid persists to temperatures as low as 1060°C. An argon:hydrogen gas mixture of 99:1 was used to control oxygen fugacity.

Weidner (1968), in his study of the system Fe-C-O, found that iron-oxide liquids persist in the presence of graphite to temperatures as low as 815°C at 350 bars.
Krauskopf (1964) showed by means of thermodynamic calculations that volatilities of iron chlorides may become important where HCl is exceptionally abundant, but that volatility is an unlikely explanation for iron transport in general.

The experiments of Tuttle and Bowen (1958) led them to state that aqueous "vapor . . . can selectively dissolve feldspars and quartz, and leave behind lime, magnesia, P_2O_5, and perhaps iron." However, Ovchinikov (1958, 1963) reported that Si, Al, Na, and Fe were transferred via an aqueous vapor phase from molten granite to cool limestone contacts in experiments done at atmospheric pressure. In addition, Burnham (1967) showed that potassium, silicon, and iron are leached from felsic rock at high temperatures in the presence of an aqueous vapor phase, and that these elements are redeposited at cooler temperatures. Furthermore, Martin and Piwinski (1969) noted evidence of iron fractionation into a coexisting aqueous vapor phase during vapor-saturated melting experiments with granites, quartz monzonites, granodiorites, tonalites, and diorites. Hematite crystals were found on the walls of the gold capsules and at the peripheries of the melted charges, suggesting that the hematite euhedra were a quench product from the vapor and not a stable phase in equilibrium with the crystals and melt under the conditions of the run. In subsolidus temperature-gradient experiments involving the interaction of an aqueous vapor phase with andesite and diabase, Martin and Piwinski
put powdered rock in a perforated gold inner capsule at one end of an 11.4 mm-long gold outer capsule, filled the outer capsule with distilled water, sealed it, put it into a pressure vessel at 5 kb, and superimposed a temperature gradient, with the rock powder at the hot end of the gradient (700°C), and the water-filled end of the outer capsule at the cool end of the gradient (400°C). The starting rock became depleted in iron, silica, sodium, potassium, aluminum, and silica. Hematite crystallized in the 515° to 470°C temperature range, and quartz and feldspar crystallized from 515° to 400°C.

Holser and Schneer (1961) measured solubility of magnetite in dilute HCl. They found that 300 ppm ferrous iron can be dissolved from magnetite in .0002 molar HCl at 390°C and 440 bars.

Helz (1971) determined equilibrium solubility curves for magnetite at temperatures from 150°C to 500°C. "Solutions containing NaHCO₃, NH₃, NaOH, or pure water revealed negligible magnetite solubility up to 300°C, and even at 500°C a maximum of only about 20 ppm Fe. In contrast, dilute HCl (10⁻²-10⁻³m) produced solubilities ranging from about 30 ppm below 275°C to over 1000 ppm at 500°C." Chloride complexes are thought to be unimportant because the concentration of NaCl had no measurable effect on solubility. "Changing the oxidation state of the solution had no effect below 350°C, but at higher temperatures the solubility was enhanced in the more oxidized solutions."
Helz concluded that, "In acidic solutions up to 275°, solubility relationships are consistent with Fe$^{2+}$ being the predominant iron species in solution. However above about 350° in acidic runs, the principal species must contain ferric iron and have a net charge no greater than +0.7 per atom of iron." He proposed Fe$_2$(OH)$_5^+$ as the most probable formula for the species soluble in acidic solutions, and Fe$_2$(OH)$_6^0$ for the species soluble in neutral to basic solutions at high temperature.

Mineral stabilities: Liou (1973, in press) has experimentally studied the phase relations between greenschist and amphibolite facies assemblages. His work indicates that the assemblage actinolite + oligoclase + epidote, which is typical of altered zones around Romeral, is stable between 475° and 550°C at 2 kb.

Holdaway (1972) studied the thermal stability of epidote, clinozoisite, and zoisite as a function of fO$_2$ and iron content, and found that in the presence of abundant iron, clinozoisite is stable at low fO$_2$, whereas epidote is stable at higher fO$_2$.

Buddington, Fahey, and Vlisidis (1955) studied variation in the titanium content of magnetite relative to that of its host rock. They found wide variations in titanium contents of magnetites which did not correspond with titanium contents of the whole rocks, but which did correlate with inferred temperatures of formation of the rocks. They
concluded that where ilmenite is present, the percent of TiO$_2$ in magnetite is largely a function of temperature, and they suggested the use of titaniferous magnetite for relative geological thermometry.

Heier (1956) criticized Buddington for largely ignoring the possible effects of pressure and chemical composition of the system. He pointed out that substitution by Ti in the magnetite lattice is accompanied by reduction of some of the trivalent iron, and that an oxidizing environment will tend to oppose this.

Lindsley (1961-62) experimentally determined the compositions of titaniferous magnetites in equilibrium with ilmenite at different temperatures for the oxygen buffer assemblages nickel-nickel oxide, quartz-fayalite-magnetite, and wüstite-magnetite. He put ulvöspinel-rich cubic spinels, or magnetite-rich cubic spinels plus ilmenite into silver-palladium capsules with water. He then placed the sealed capsules into pressure vessels containing the buffer assemblages, filled the pressure vessels with water, and brought them up to temperature (550-1000°C) and pressure (2 kb or less). Because Ag$_{70}$Pd$_{30}$ capsules are permeable to hydrogen, water inside the capsules equilibrates with the external buffer assemblage, thus controlling the fO$_2$ of the experiment.
The following reactions took place during the buffered experimental runs: (1) ulvöspinel-rich cubic spinel + O₂ ⇄ Fe-Ti spinel + ilmenite, (2) magnetite-rich cubic spinel + ilmenite ⇄ Fe-Ti spinel + ilmenite ss + O₂.

Lindsley (1961-62, p. 102) found that unit-cell dimensions of titaniferous magnetites vary directly with Ti content and that "for each temperature and oxygen fugacity, the unit-cell edges of the Fe-Ti spinels formed in reactions 1 and 2 are nearly identical and are intermediate between those of the initial cubic spinels, suggesting that the Fe-Ti spinels have approached the composition of the spinel that is in equilibrium with ilmenite ss."

His experiments strongly supported the hypothesis that magnetite-ilmenite intergrowths result from oxidation of magnetite-ulvöspinel solid solutions and established that the composition of the cubic spinel in equilibrium with ilmenite ss is strongly dependent on oxygen fugacity as well as on temperature. By the same methods, Lindsley (1962-63) determined the compositions of the hematite-ilmenite solid solutions in equilibrium with magnetite-ulvöspinel solid solutions at different temperatures, oxygen fugacities, and total pressures. Then, combining the data on magnetite-ulvöspinel ss with those on ilmenite-hematite ss, he constructed a diagram that permits estimation of both fO₂ and temperature of formation for coexisting analyzed magnetite-ilmenite pairs.

An attempt was made in the present study to use...
Lindsley's diagrams to determine temperatures and oxygen fugacities at which igneous and hydrothermal Fe-Ti oxides formed at El Romeral. Application of the experimental data to natural minerals involves the assumption that the Fe-Ti oxide pairs formed in equilibrium during crystallization of the rock containing them (Lindsley, 1962-63). The effects of impurities are unknown, but rough estimates can be made as follows:

1. MgO, MnO₂, and CaO replace FeO, and, by diluting FeO, they extend the fO₂-T stability range of the Fe-Ti oxides containing them.

2. Cr₂O₃, Al₂O₃, V₂O₃ occur in solid solution with magnetite as the spinels FeCr₂O₄, FeAl₂O₄, and FeV₂O₄. Their effect on the fO₂ stability of magnetite is roughly similar to that of ulvöspinel.

Verhoogen (1962) studied the distribution of Ti between silicates and oxides in igneous rocks and concluded that diopside-ilmenite is the likely assemblage at high temperature and low pO₂, whereas magnetite-sphene is favored by lower temperature and/or higher pO₂.

Methods of Investigation

All field work was done during late May through early September of 1970. In the two Romeral open pits, all accessible exposures were mapped at a 1:1000 scale on progress maps provided monthly by the mine topographers. The geology of the 13.4 sq km area around the mine was mapped.
by outcrop isolation at a scale of 1:5000 on the excellent topographic base constructed by Melo (1965).

Brief reconnaissance traverses northward to the apatite-magnetite deposits of La Escoba, westward to the coast, northeastward, up Quebrada Chacay, and southward to the La Serena quadrangle were plotted at 1:50,000 scale on topographic maps from the Instituto Geográfico Militar de Chile.

About 1700 hand samples were collected, and diamond drill cores from several holes on two cross sections were examined and sampled.

Brief trips were made with Francisco Franquesa to El Tofo, El Algarrobo, and Cerro Imán, near Copiapó, to see other major iron deposits; to Cemento Melón, to see an apatite-magnetite deposit; to Cerrillos, south of La Serena, to see Cretaceous ocoite andesites interbedded with limestones and cherts; to Los Choros to see schists of Paleozoic age and volcanic andesites of Jurassic age; and to the Instituto de Investigaciones Geológicas de Chile, in Santiago, to discuss the iron deposits of Chile with Carlos Ruiz F. and the geology of the La Serena quadrangle with Ernesto Egert R.

During 1970-71 geologic maps and sections were compiled, and assay-contour maps were made to show iron grade and phosphorous distribution at the 310 level of the Main pit. Sample location maps were made, hand samples were sorted by rock type, and representative samples of each major rock
type were chosen for thin sectioning. Thin sections and rough-polished slabs were prepared for about 170 samples. One side of each slab was stained for plagioclase and K-spar by the method of Laniz and others (1964), and 100 slabs were polished for microscopic examination in reflected light.

During 1971-72, most of the minerals of El Romeral were identified under the direction of Dr. C. O. Hutton. About 600 plagioclase determinations were made, using the universal stage. The Michelle-Levy and Ritmann methods were routinely used to determine compositions of polysynthetically twinned plagioclases. The Federov method, in conjunction with determinative curves by Dr. C. O. Hutton (1971), was used to determine both the composition and structural state of at least one plagioclase crystal from each rock type. Untwinned plagioclases were identified by measuring the $X'$ index of refraction on (001), using the determinative curves of Heinrich (1965). Optical properties of pyroxenes and amphiboles were determined in index oils, with sodium light, and with the crystals oriented properly by use of a spindle stage. Optical properties of other minerals of variable composition, such as epidotes, scapolite, and apatite were determined in index oils with sodium light.

Chemical compositions for minerals were estimated by matching optical properties and petrologic associations of the minerals to those of chemically analyzed minerals reported in the literature (Appendix C). Chemical composition of one amphibole sample was determined by microprobe analysis.
by Jane E. Pike (Appendix G). Clay minerals, pyrophyllite, and fine-grained muscovite were identified by X-ray diffraction. Opaque minerals were identified by reflected light microscopy, and microtextures were studied in high-power oil immersion mounts. Titanium contents of Fe-Ti oxides were estimated on the basis of unit cell parameters determined by X-ray diffraction and were checked by emission spectrographic analysis.

In order to study trends of igneous differentiation, and to assess changes that occurred during hydrothermal alteration, the mineralogical and chemical compositions of many rocks were determined.

In determining modes, 600 points were initially counted per sample, but it was found that a larger 1500-point sample was required for opaque minerals present in small percentages. In addition, it was found necessary to measure grain sizes of opaque and dark-colored minerals, to apply the Holmes correction factor for opaque minerals (Chayes, 1956), and to apply 1/2 the Holmes correction factor for fine-grained, dark-colored minerals. Because the igneous plagioclases of the Romeral rocks are generally zoned, it was necessary to make separate An determinations on cores, inner rims, rims, and unzoned groundmass feldspars, and to estimate the volume percentages of each. Because opaque grains are generally complex intergrowths of Fe-Ti oxides that are identifiable in polished section but not in thin section, volumetric percentages of various opaque minerals in the
intergrowths were visually estimated in each of ten or more randomly located grains in the polished billet from which the corresponding thin section had been cut.

Chemical compositions of 92 rock samples were calculated from modal analyses as suggested by Friedman (1960) and Seyfert (1964). An ALGOL W computer program written by James Krumhansl was used to perform the calculations on the Stanford IBM computer (Appendix D).

Chemical analyses of 10 rock samples provided by the University of Mexico provided essential checks for the calculated compositions. The analyses were performed by Alberto Obregón, Chief Chemist of the Laboratorios de Química, Universidad Autónoma de México by methods summarized in Appendix B.

Through comparison of the chemical analyses with tentative calculated compositions, it was found that major-element calculations were much more sensitive to changes in volume percentages of minerals than to minor changes in estimated mineral compositions.

To determine iron content of ferruginous quartzites near El Romeral, eleven samples were analyzed for iron by titration (Appendix B).

Two melting experiments were done in an attempt to produce artificial water-rich magnetite-apatite magma at temperatures below 900°C. A series of temperature-gradient, vapor-phase transport experiments were also done in an effort to simulate hydrothermal mobilization and transport
of magnetite and apatite at subsolidus temperatures. These experiments were carried out at the Stanford Earth Science Laboratory under the direction of Dr. W. C. Luth, and with the help of James Blencoe.

In the melting experiments granular mixtures of magnetite and apatite were put above granular quartz in a gold capsule with water. If melting were to occur, the heavy iron-oxide liquid would move downward and be found between the quartz grains at the end of the experiment. However, no melting occurred at 800° or 900° C.

In the vapor-phase-transport experiments, pulverized magnetite and apatite from El Romeral were put at the bottoms of 11 mm-long Ag-Pd capsules, and the upper parts of the capsules were filled with pulverized La Liga andesite. About 10 weight-percent water, or .02 molal HCl, or 2 molal NaCl solution was added, and the capsules were sealed. They were then put into nickel-nickel oxide pressure vessels and put under 640-380°C and 770-450°C temperature gradients at 2 kb pressure for about two weeks, with the magnetite-apatite mixtures at the hot, bottom end of the furnace. The hope was that iron would be taken into solution at the hot end and would be transferred by diffusion to the cooler end, where it would be redeposited as magnetite. No such transfer was observed in any of the runs, but in the 770-450°C runs, the capsules leaked, and there was insufficient time to repeat the experiment.
Regional Geologic Setting

The major geological features of northern Coquimbo Province are the Peru-Chile trench, the Benioff seismic zone, the Paleozoic crystalline basement complex, the calc-alkaline batholiths intrusive into the andesites, and the Atacama fault, a north-trending strike-slip fault system that parallels the coast for about 1000 km (Fig. 7).

The Peru-Chile trench, about 4 to 6.5 km deep, runs parallel to the coast of Chile about 80 to 150 km off shore (Scholl and others, 1968). From the trench, the present Benioff seismic zone dips eastward beneath the western edge of the South American continent. At the latitude of northern Coquimbo Province, the dip of the seismic zone is about 20 to 25°E (James, 1971) (Fig. 7).

Along the coast and near it, Paleozoic metasedimentary basement rocks are exposed in eroded anticlines and upthrown fault blocks (Figs. 7 and 8). These rocks are tightly folded phyllites, schists, and quartzites. Foliations and axial planes of folds in these rocks generally strike northerly to northwesterly and dip steeply. South of Punta Lengua de Vaca, the metasedimentary rocks are intruded by Permo-Triassic granites.

Paleozoic metasedimentary rocks and granites are overlain unconformably by thin, discontinuous Triassic continental sedimentary rocks and rhyolitic volcanic rocks (Ruiz and others, 1965), and by a thick section of
Jurassic and Cretaceous andesites (Fig. 7).

Volcaniclastic island-arc andesites of Jurassic age are exposed in the cores of eroded anticlines near the coast, whereas Jurassic miogeosynclinal limestones are exposed east of La Serena, near the head of the Elqui River (Ruiz and others, 1965).

Lower Cretaceous submarine andesite flows and sills overlie the Jurassic volcaniclastic andesites. They are interlayered with thin marine limestones that thicken easterly, and they are overlain by Upper Cretaceous andesite flows and breccias, which are interlayered with continental conglomerates (Aguirre and Egert, 1965).

The entire Mesozoic andesite section probably was about 10-12 km thick in northwestern Coquimbo Province, and the pre-ore Mesozoic section probably was about 7 ± 3.5 km thick. In general, the andesites are metamorphosed to low-grade meta-andesites, and they are warped into broad, north-trending folds. At the western edge of the continent these folds have gentle dips (less than 20°), wave lengths of about 50 km, and amplitudes of about 0.7 km. Eastward they become progressively shorter in wave length (Fig. 7).

A succession of elongate, north-trending batholiths of calc-alkaline composition intrudes the Mesozoic volcanic andesites. The relative ages of the batholiths generally decrease from west to east, with Jurassic batholiths along the coast, Cretaceous batholiths 5 to 80 km inland, and Cretaceous-Tertiary batholiths 75-100 km east of the coast
Figure 7. Generalized geology of northern Coquimbo province (after Ruiz and Corvalán, 1968). The section was constructed from the map, from a contour map of the Peru-Chile trench (Scholl and others, 1968), and from a cross section of the inclined seismic zone (James, 1971).
EXPLANATION

QUATERNARY
- Qel  Alluvium

TERTIARY
- T  Rhyolitic volcanic rocks
- Tg  Granitic rocks

CRETACEOUS-TERTIARY
- Tg  Granitic rocks

CRETACEOUS
- Kg  Andesites and continental sedimentary rocks (Vinita Formation)
- Kf  Andesites and continental sedimentary rocks (Quebrada Marquesa Formation)
- Kf  Granodioritic batholiths
- Km  Magnetite deposits
- Kg  Romeral diorite
- Ka  Andesite sills and marine limestones
- Kap  Intrusive andesite porphyry

JURASSIC-CRETACEOUS
- Km  Quartz monzonite

JURASSIC
- Jd  Granodiorite and quartz diorite
- Jm  Marine limestones
- Jv  Volcaniclastic andesites

TRIASSIC
- T  Volcanic rhyolites and continental sedimentary rocks

PALEOZOIC
- Pa  Granite

- Schists, phyllites, and quartzites

Contact
Fault
Thrust fault
(Fig. 7).

North of Coquimbo Province, the extensive north-trending strike-slip Atacama fault system parallels the coast for about 1000 km. In northern Chile the fault is active, with right-lateral recent stream offsets of as much as 1.5 km. Discontinuous breaks delineate its trend for 300 km southward, from Copiapo to El Romeral (Saint Amand and Allen, 1965).

Local Geologic Setting

The magnetite deposits of El Romeral are in a composite sliver (4 km by at least 8 km) of metasedimentary rocks of Paleozoic age and andesitic and dioritic rocks of lower Cretaceous (Neocomian) age, which is bounded on the west by a pre-ore batholith of quartz monzonite, and on the east by a post-ore batholith of granodiorite (Fig. 8). North of El Romeral the metasedimentary basement rocks are overlain with angular unconformity by altered andesites of probable Jurassic age. West of El Romeral the andesites are intruded by a large body of andesite porphyry, which is in turn intruded by two large masses of diorite. The eastern margin of the largest diorite mass is cut by a north-trending strike-slip fault system, and the Romeral magnetite deposits are in altered andesite porphyries and schists near the fault system and near the eastern contact of the largest diorite mass (Fig. 8). Reconstruction of the Jurassic and early Cretaceous sections suggests that the Romeral magnetite deposits formed under about 7 ± 3.5 km of andesitic rocks.
Figure 8. Generalized geology of the Romeral vicinity.
EXPLANATION

QUATERNARY

Qol
Alluvium and fanglomerate, undivided

CRETACEOUS

Kga
Granite aplite

Kgs
Punta de Piedra granodiorite

La Totorra quartz-bearing diorite

Actinolite-aplite-magnetite vein

Quartz-feldspathic rock

Km
Intramineral diorite dikes

Kaf
Magnetite ore

Kdf
Actinolite rock

Kap
Romeral diorite

La Liga intrusive andesite porphyry

JURASSIC - CRETACEOUS

Trondjemite

Quartz monzonite

JURASSIC

La Escoba volcaniclastic andesite

PALEOZOIC

Ps
Schist, phyllite, quartzite

Fault

definite

indeterminate

concealed

Contact

definite

indeterminate or gradational

concealed or inferred
The Main orebody at El Romeral is a large, lenticular, steeply dipping mass of magnetite-rich rock surrounded by actinolitized phyllites and andesite porphyry. The phyllite lies north of the orebody, and the porphyry forms a re-entrant between a large west mass and a smaller east mass of Romeral diorite. The orebody is immediately east of and subparallel to the strike-slip Romeral fault zone, which is generally along the eastern margin of the west mass of Romeral diorite (Fig. 10). The ore consists of magnetite, microscopically intergrown with actinolite, and it is cut by actinolite and calcite veinlets. Ore-zone boundaries are gradational, and andesite porphyry southeast of the Main orebody is cut by magnetite veinlets and is altered to actinolitic rock containing oligoclase, diopside, clinozoisite, sphene, apatite, scapolite, tourmaline, chlorite, pyrite, and calcite.

Several northwest-trending, steeply dipping intramineral diorite dikes cut the Main orebody. Four generations of these dikes can be recognized. In some places the first- and second-generation dikes are substantially actinolitized and replaced by magnetite, but in other places the same dikes cut the ore and are no more than slightly mineralized. The third- and fourth-generation dikes are predominantly post-ore in age, but they are cut by magnetite-bearing actinolite and chlorite veinlets.

At the north end of the Main orebody, actinolitized andesite porphyry is in contact with altered Paleozoic
phyllites that have been thermally metamorphosed to oligoclase-biotite-cordierite hornfelses, metasomatized to actinolitic and chloritic hornfelses, and hydrothermally altered to argillites. The andesite-phyllite contact strikes about east-west and dips steeply. A few pods of magnetite ore lie north of the contact, along the northwest-trending, steeply dipping foliation of the phyllites.
Figure 9. Topography, surficial deposits, and El Romeral mine workings.
Figure 10. Generalized bedrock geology of the Romeral area.
The North orebody consists of conformable lenses of magnetite in steeply foliated actinolitized schists that lie structurally and stratigraphically below the phyllites. The schists and phyllites surrounding the ore contain disseminated actinolite and magnetite, accessory apatite and pyrite, and minor tourmaline and chalcopyrite.

North and west of the North pit, fault-bounded slivers of biotite schist have been replaced by apatite and magnetite, and have been converted by shearing deformation to cataclasites with flow banding suggestive of igneous emplacement. Isoclinallly folded hematite-banded quartzites are present east of the North orebody.

West of both orebodies, the Romeral diorite has been altered to aplite-like quartzo-feldspathic rock within an irregularly shaped zone that ranges from 300 to 800 m wide (east-west), and from 1500 to 3000 m long (north-south), as shown in Fig. 10. Contacts of the quartzo-feldspathic zone with actinolitized diorite and phyllite are gradational, and transitional zones range in width from a few millimeters to tens of meters. Plagioclase in rocks of the quartzo-feldspathic zone ranges from oligoclase to albite, and potassium feldspar, found near the outer edges of the zone, is microcline.

Actinolite veins containing magnetite and apatite are scattered within a 4 km radius of El Romeral. Most dip steeply and strike at high angles to the Romeral orebodies. They cut all major igneous rock units older than the Romeral
orebodies, and they also cut quartzo-feldspathic rocks west of these orebodies. The veins range in thickness from about one centimeter to several meters, and they are characterized by actinolite prisms that are perpendicular to vein walls and exhibit comb structure. Magnetite and apatite are present interstitially among the actinolite prisms, in crustified central portions of the veins, and in veinlets that cut the larger veins. At several localities, including La Escoba, the veins have been mined for apatite.

About 3.5 km northwest of El Romeral, near La Escoba, Paleozoic biotite schists are unconformably overlain by sheared and altered Jurassic (?) andesites that contain abundant secondary biotite and actinolite. About 1.5 km south of La Escoba, the black Jurassic (?) andesites are intruded by gray-green La Liga andesite porphyry. Such intrusive porphyry crops out over the entire ridge and west side of Cordón La Liga. It is intruded by Romeral diorite, and probably is correlative with the andesite porphyry that is the host rock of the Main orebody. Both porphyries are devoid of layering, flow banding, pillow structure, or vesicles. They probably represent a magma chamber that fed lower Cretaceous submarine sills and flows of the Pm Arqueros.

About 4 km west of El Romeral is the Cerro Juan Soldado quartz monzonite batholith. Marginal parts of this batholith are cut by magnetite-apatite veins; the batholith therefore is presumed to be pre-ore in age. The post-ore Punta de
Piedra granodiorite batholith cuts quartzites, schists, and magnetite pods in schist about 150 to 1500 m east of the Romeral orebodies. This batholith has a discontinuous marginal facies of quartz-bearing diorite. Granodiorite aplites cut both the quartz-bearing diorite and the granodiorite facies of the batholith.

A tabular summary of the rocks of northern Coquimbo Province, and of the sequence of geological events they may represent, is given in Figure 11.
Figure 11. Tabular summary of rocks, rock ages, and sequence of geologic events, northern Coquimbo Province, Chile.
PRE-ORE ROCKS

Metasedimentary Rocks of Paleozoic Age

Quartzites, schists, and phyllites north of the Main orebody are the oldest rocks exposed at El Romeral. They are present in the deeply eroded core of a broad anticline of Tertiary (?) age, where they are overlain with angular unconformity by Jurassic (?) volcanic andesites and are intruded by early Cretaceous andesite porphyry and diorite.

The metasedimentary rocks are tightly folded into nearly isoclinal, chevronlike folds, which are themselves gently cross-folded and crenulated. Hematitic quartzites in the cores of large isoclinal folds grade outward and stratigraphically upward into schists, which in turn grade into phyllites (Fig. 12).

Quartzites: Hematitic quartzites are present east of the North orebody, northeast of the Main orebody, and north of the northern margin of the east mass of Romeral diorite (Fig. 10). Where unaltered, these quartzites are white- and pink-layered, and the layers are nearly isoclinally folded. However, hematite has been leached from the rocks in many places, so that only faint pink remnants of the hematite layers remain. In other places, the hematite bands are obscured by introduced hematite.

The quartzites typically contain 74 to 99.9 volume percent of quartz and 0.1 to 24 volume percent of hematite. In ten quartzite samples analyzed by wet chemical methods,
Figure 12. Generalized composite section, showing the stratigraphy and structure of the metasedimentary rocks of El Romeral.
the iron content averaged 5.6 weight percent and ranged from 0.05 to 16.5 weight percent.

The iron content of quartzites near bodies of Romeral diorite is very low, and such quartzites contain poikiloblastic plagioclase, diopside, and actinolitic hornblende. However, the iron content of quartzites 80 to 145 m from the diorite contact is higher than average. This suggests that hematite was leached from quartzites near the diorite and was redeposited in the quartzites farther from the gradational diorite-quartzite contact (Fig. 56, p. 151).

In many samples dust-sized hematite particles (.005-.1 mm) are evenly distributed along boundaries of quartz grains, but the quartz grains themselves contain no hematite. In others hematite occurs in amoeboid clots (.01-.3 mm) concentrated along diffuse bands that outline isoclinal folds. These ferruginous clots cross quartz grain boundaries and include isolated remnants of quartz that are in mutual optical continuity. Thus they may well have partially replaced some quartz grains.

In quartzites near the Romeral orebodies, actinolite, calcite, apatite, and sphene are present. Microscopic needles of actinolite are enclosed within quartz grains, but the crystals of calcite, apatite, and sphene are at grain boundaries.

Southeast of the North orebody, sericite and muscovite are present in the quartzite in a zone parallel to the post-ore East fault. They probably were added during post-ore
hydrothermal alteration.

Axial portions of the isoclinal folds in quartzite are several times thicker than the limbs, which suggests gross migration of silica from the limbs toward the crests and troughs. Quartz grains in the axial regions generally are larger (.5 to 2.5 mm), more equant, and more lobate than the smaller (.05 to .15 mm), elongate, and relatively straight-sided grains along the fold limbs. Some of these differences in grain size and shape probably reflect differences in strain rate, the smaller, elongate grains having formed in limb areas of rapid strain, and larger, more equant grains in axial areas of slower, more isotropic strain. Other differences between the larger and smaller grains may have resulted from differences in degree of post-deformational recrystallization. The larger grains show undulatory extinction and are preferentially oriented with their C crystallographic axes about parallel with the axes of the respective folds. In contrast, the smaller grains of the limbs show sharp extinction and do not have preferred crystallographic orientation. These observations suggest that the larger grains have retained the orientation and strain induced during folding, whereas the smaller grains, which probably were originally more strained, may have recrystallized in response to later thermal metamorphism.

The scarcity of feldspar and the lack of detrital heavy minerals such as zircon and rutile suggests that the hematitic quartzites were derived from chemically precipitated
Figure 13. Boulder of banded hematitic quartzite from El Tofo. Hematite banding is less evident in the quartzites of El Romeral.

Figure 14. Outcrops of hematite-depleted quartzite about 1 km east of the Romeral orebodies. Discontinuous, faint pink remnants of hematite banding have been accentuated by drawing on the rocks with a red marking pen.
ferruginous cherts. Hematite banding may have been present in the original sediments, but it probably was enhanced by metamorphic segregation during regional deformation and metamorphism, and then obscured during thermal metamorphism and hydrothermal alteration.

Schists: Actinolitized biotite schists are the host rocks of the North orebody and of several smaller conformable pods of massive magnetite. The essential minerals of these schists are brown biotite, plagioclase (An\textsubscript{22} to An\textsubscript{37}), and quartz. Rock types include plagioclase-biotite schists, biotite-quartz schists, and plagioclase-biotite-quartz schists with interlayered quartz-plagioclase quartzites. Common secondary constituents are actinolite, magnetite, and cordierite. Minor secondary constituents include chlorapatite, clinozoisite, microcline, tremolite, montmorillonite, pyrite, and chalcopyrite (Table 1).
A schistosity is defined by mutually parallel biotite plates, and a foliation by alternating biotite-rich layers (.1 to 2 mm thick) and quartzo-feldspathic layers (.1 to 5 mm thick). The biotite crystals are .05 to .2 mm across and about .02 to .04 mm thick. Fold axes are marked by sharp, acute angles, but the biotite flakes along these axes are not bent or broken (Fig. 15). The biotite apparently recrystallized after the deformation that produced the chevron folds.

Quartz and oligoclase-andesine grains range from .02 to .1 mm across. They are relatively equant and strain free, and their boundaries are generally smooth. There is little interlocking between grains, and 120° triple junctions are
Figure 15. Microtexture of schist from the North pit. Biotite, quartz, and plagioclase are recrystallized. Actinolite and anhedral magnetite preferentially replace biotite-rich layers. Apatite is mostly in quartzo-feldspathic layers.
common, suggesting a close approach to textural equilibrium. Quartz and plagioclase generally are confined to concordant layers, but in axial portions of some folds, migmatitic quartz-plagioclase pods and stringers cut across biotite-rich layers.

Actinolite grains are concentrated in the biotite-rich layers as prisms that lie in the plane of foliation but show no preferred lineation (Fig. 15). Magnetite tends to be concentrated in the biotite-actinolite layers. Single and composite magnetite grains generally are lobate, skeletal, and very irregular in shape. Along the western side of the North pit, near the Romeral fault, skeletal magnetite grains are commonly rod-shaped, with their long axes plunging steeply southward (Fig. 45, p. 132).

Rare tremolite prisms lie in the foliation but have no preferred lineation. Many of them have been partly replaced by actinolite. The replacement was epitaxial and was guided by grain boundaries and cleavage planes (Fig. 16). The cordierite porphyroblasts (about 1 x 2mm) are sievelike, and are randomly oriented. They contain abundant inclusions of feldspar, biotite and quartz (Fig. 17). The grains of chlorapatite, which are equant and subrounded, occur singly or in clusters and are closely associated with actinolite and magnetite.

Fine-grained pyrite is disseminated in the schist along the west side of the North pit. The grains are angular, and some of the pyrite contains blebs of chalcopyrite. Rare
Figure 16. Epitaxial replacement of tremolite by actinolite in schist, North pit.

Figure 17. Cordierite porphyroblast in schist, North pit.
chalcopyrite grains are partly altered to chalcocite and covellite.

Crenulations are superimposed on tight, chevronlike folds as shown in Figure 18. The average trend of axial planes of folds in the North pit is N 15° W, 87° SW, as shown on the equal area net in Figure 19. Fold axes at the north end of the pit plunge northwest, whereas fold axes at the south end plunge southeast. The mean plunge of crenulation axes is 61° S. The double plunge of the isoclinal fold axes, and perhaps the crenulations, may both have resulted from a second episode of folding during which broad, roughly east-west trending folds, and steeply plunging crenulations were superimposed on earlier northeast-trending, nearly isoclinal folds. Alternatively, the crenulations may have formed later, in response to strike-slip faulting.

The sedimentary antecedents of the schists probably were sandy shales and shaly sandstones. These were deeply buried, deformed into nearly isoclinal folds, and metamorphosed, probably to quartz-biotite-albite schists. Somewhat later they may have been thermally metamorphosed to quartz-oligoclase-andesine schists. Later still, they were further deformed into broad, east-northeast trending cross folds. South-plunging crenulations may have formed during this second episode of folding or during later strike-slip faulting.

In Cretaceous time the schists were thermally metamorphosed during intrusion and crystallization of the La Liga andesite porphyry and Romeral diorite. Quartz, oligoclase,
Figure 18. Chevronlike, nearly isoclinal folds in schists and phyllites at the north end of the North pit, at the 400 m elevation.

a. quartz-plagioclase (An30) pods in schist.
b. small-scale chevronlike folds in schist.
c. crenulations superimposed on isoclinal folds in schist.
Figure 19. Orientations of foliations, axial planes of folds, and plunges of crenulation axes in schists of the North pit (lower hemisphere equal-area projections).

Figure 19a. Poles to foliations and crenulation axes of the North pit (lower hemisphere equal-area projections). Figure 19b. Plunges of fold axes in schists in the north and south parts of the North pit.

Figure 19c. Plunges of crenulation axes in schists of the North pit.
and biotite were recrystallized, and cordierite was formed. The schists were then subjected to shear stresses associated with the Romeral strike-slip faulting. While shearing took place, hydrothermal fluids permeated the schists, and magnetite, actinolite, apatite, clinozoisite, tourmaline, axinite, pyrite, and minor chalcopyrite partially replaced the schists.

Phyllites: Phyllites crop out north of the intrusive andesite porphyry in the north part of the Main pit and along all sides of the North pit (Fig. 10). They lie stratigraphically above the schists (Fig. 12) and below the Jurassic angular unconformity (cross section, Fig. 8).

The essential minerals of the phyllites are plagioclase (about An30) and yellow-brown biotite. Porphyroblasts of cordierite are abundant near the Romeral diorite contact. Actinolite is abundant near the Romeral orebodies, where green chloritic knots also are common. At the north end of the Main orebody, phyllites are argillized. Montmorillonite is the most common clay mineral, but dickite, kaolinite, pyrophyllite, and sericite are present in zones of most intense post-ore alteration.
Table 2
MINERALOGICAL COMPOSITION* OF PHYLLITE

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol.%</th>
<th>An</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>47.6</td>
<td>30</td>
<td>34-65</td>
</tr>
<tr>
<td>biotite</td>
<td>23.1</td>
<td></td>
<td>0-51</td>
</tr>
<tr>
<td>chlorite</td>
<td>11.7</td>
<td></td>
<td>0-38</td>
</tr>
<tr>
<td>cordierite</td>
<td>9.0</td>
<td></td>
<td>0-23</td>
</tr>
<tr>
<td>magnetite</td>
<td>4.8</td>
<td></td>
<td>0-10</td>
</tr>
<tr>
<td>nontronitic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>montmorillonite</td>
<td>3.3</td>
<td></td>
<td>0-17</td>
</tr>
<tr>
<td>limonite</td>
<td>0.3</td>
<td></td>
<td>0-2</td>
</tr>
<tr>
<td>actinolite</td>
<td>0.2</td>
<td></td>
<td>0-4</td>
</tr>
<tr>
<td>apatite</td>
<td>0.1</td>
<td></td>
<td>0-2</td>
</tr>
<tr>
<td>clin zoisite</td>
<td>0.1</td>
<td></td>
<td>0-1</td>
</tr>
</tbody>
</table>

* This is a mean composition, based on modal analyses of four samples (Appendix A, Table A-2).

Foliation in the phyllites generally strikes northwest and dips steeply. The largest exposures of phyllites are on the southwest limbs of anticlines, and therefore the predominant observed dip is to the southwest. In well foliated phyllites, biotite flakes are mutually subparallel, and biotite-rich layers about .1 mm thick are partially segregated from plagioclase-rich layers about .1 to .3 mm thick. In poorly foliated phyllites, biotite flakes are more randomly distributed and oriented.

The plagioclase grains are equant and have smooth boundaries. There is little interlocking between adjacent grains, and 120° triple junctions are common. This indicates a close approach to textural equilibrium among the primary constituents of the phyllites.

Cordierite porphyroblasts (.5 to 5 mm) are poikiloblastic
and contain numerous biotite and plagioclase inclusions. The cordierite increases in abundance toward the diorite contact, so that phyllites pass gradationally into porphyroblastic cordierite hornfelses that texturally resemble andesite porphyries and previously have been mapped as meta-andesites (Melo, 1965).

Gray-green, spherical to oblate knots of intergrown chlorite, nontronitic montmorillonite, and magnetite are common in phyllites northeast of the Main orebody and east of the North orebody. In some places chloritic knots constitute 5 to 30 volume percent of the rock and give it a spotted appearance. Within the Romeral fault zone spotted hornfelses are sheared to cataclasites, and biotite and plagioclase grains have been deflected around the nodules.

Phyllite west of the Main orebody, between the Romeral fault and the Romeral diorite, is completely actinolitized to a felty, foliated rock composed of acicular actinolite and disseminated magnetite.

The phyllites were derived from shaly sediments, and were deformed and metamorphosed along with the quartzites and schists. Intrusion and cooling of the Romeral diorite thermally metamorphosed the phyllites to cordierite hornfelses. Pneumatolysis associated with formation of the Romeral orebodies changed the phyllites to felty actinolitic rocks and spotted chloritic hornfelses. Shearing near the Romeral fault transformed the spotted hornfelses
into nodular cataclasites. Post-ore hydrothermal alteration changed some of the phyllites to argillites.

The phyllites contain only small pods of magnetite ore, perhaps because the phyllites were less permeable than the schists, less fractured than the intrusive andesites, and therefore less penetrable by ore-forming hydrothermal fluids than either the schists or the fractured andesites.

**Quartz Phyllonite:**

A 50 meter wide quartz phyllonite zone, trending N 40° W and dipping 75° SW, traverses the quartzite southeast of the North pit and is cut by the post-ore north-northeast fault (Figs. 10 and 12). In outcrop the quartz phyllonite resembles a phyllite in that it is very fine grained and distinctly foliated. Unlike the Romeral phyllite, however, the phyllonite contains about 75 percent of quartz, 20 percent of actinolite, 3 percent of magnetite, only 2 percent of oligoclase, and no biotite. The quartz is very fine grained (.05-.1 mm), its grain boundaries are smooth, and its extinction is moderately oscillatory. The foliation is caused by actinolite needles which are concentrated in thin layers (.05-.1 mm) along the limbs of very tight folds with 5-10 mm wave length and 2-10 cm amplitude. In addition, lenses of actinolite needles and lenticular magnetite grains are subparallel with the plane of foliation.

The quartz phyllonite probably is extremely deformed quartzite. Grain shapes indicate that the rock is
recrystallized, and the oscillatory extinction further indicates that the recrystallized grains are somewhat strained. Most of the deformation probably took place during or after the late stages of isoclinal folding of the quartzite. Recrystallization accompanied later thermal metamorphism, and introduction of actinolite and magnetite accompanied Romeral ore deposition. Re-straining of the phyllonite quartz grains may have occurred during post-ore movement on the nearby East fault.

Correlation and Age of the Metasedimentary Rocks

Previous investigators at El Romeral have observed apparent gradation from sheared, altered andesites to altered schists and have postulated that the schists were derived from Mesozoic andesites by shearing and metamorphism (Melo, 1965; Ruiz and others, 1965; and Ruiz, 1968). However, several lines of evidence make this hypothesis untenable and suggest instead that the quartzites, schists, and phyllites are pre-Permo-Triassic in age. First, it can be seen in thin section that the phyllites contain cordierite porphyroblasts rather than plagioclase phenocrysts, and hence are not compositionally similar to the andesite porphyries. Second, it is unlikely that the Mesozoic rocks, which are overwhelmingly andesitic, could have been transformed into quartzites, quartz-bearing biotite schists, and cordieritic phyllites by mesozonal shearing and contact metamorphism. Third, the structural style of the quartzites, schists, and
phyllites, which are pervasively isoclinally folded, is very different from that of the Mesozoic andesites, which are gently warped into very broad, open folds and are fractured and brecciated along fault zones. Fourth, the angular unconformity near La Escoba, between biotite schist and underlying volcaniclastic andesites of Jurassic (?) age, suggests that the schists are pre-Mesozoic in age. Finally, the schists and phyllites of El Romeral are mineralogically and structurally similar to those of Punta Lengua de Vaca at the south end of Coquimbo Bay about 70 km south of El Romeral, which are intruded by granites that have been radiometrically dated as Permo-Triassic (219 ± 20 m.y.). The Punta Lengua de Vaca schists are older than the granites which intrude them, and hence are pre-Permo-Triassic in age. The Romeral metasedimentary rocks are very similar to those of Punta Lengua de Vaca, and they probably also are pre-Permo-Triassic in age.

The pre-Mesozoic schists of central and northern Chile are strongly deformed and lack fossils, and until recent years they have been regarded as Precambrian. The results of radiometric dating, however, indicate that the schists of central Chile are Paleozoic in age (González-Boronino, 1970, p. 980). No radiometric dates have been reported for those of northern Chile, but the schists of El Romeral are structurally and mineralogically similar to the schists of central Chile, and they appear to have undergone a similar structural and metamorphic history.
Radiometric dating by Munizaga (1967) and structural and petrologic work by González-Boronino (1970) have established the following sequence for development of the metasedimentary rocks of central Chile:

1. Deposition of a thick section of pelitic and psammitic sediments during Carboniferous time. Rb/Sr data fit a 342 ± 5 m.y. isochron.

2. Curepto deformation and metamorphism during Permian time. K/Ar dating gives approximately 245 m.y.

3. Intrusion of granitic batholiths, accompanied by Nirivillo thermal metamorphism, during Permo-Triassic time. K/Ar dating of the intrusions give ages ranging from 217 ± 20 m.y. to 238 ± 20 m.y.

4. Pichilemu deformation and metamorphism during early Triassic time, approximately 210 m.y. ago by K/Ar dating.

The structural trend of the Curepto series is nearly north-south, and the intensity of deformation and metamorphism increases from east to west, as follows:

Zone I: slates (quartz-muscovite-chlorite-albite) in which stratification is well preserved.
Zone II: phyllites (quartz-muscovite-biotite) in which stratification can be detected where alternating sandy and shaly beds are present.

Zone III: schists and quartzites (quartz-muscovite-chlorite-albite-magnetite-garnet) with stratification almost completely obliterated.

Isoclinal folding, north-northwest trending schistosity, and incipient boudinage in the Romeral metasedimentary rocks suggest that they are correlative with Curepto rocks of zones II to III. The hematitic quartzites of El Romeral probably are correlative with magnetite-bearing quartzites of Curepto zone III at Relún, near lake Lieu Lieu (38° to 38°45' S), and at Buchalemu, near Punta Sirena (34°35' S). The magnetite-bearing quartzites at Relún contain about 20 to 40 weight percent of iron and minorapatite. They are about 20 to 30 m thick and are intercalated with mica schists (Ruiz and others, 1965, p. 245; González-Boronino, 1970).

In thermal metamorphic aureoles around Permo-Triassic intrusives, Mirivillo hornfelses contain oligoclase-andesine, andalusite, and/or sillimanite. No Permo-Triassic batholiths are known in the immediate vicinity of El Romeral, but exposures of pre-Mesozoic rocks are limited. Oligoclase-andesine in the Romeral phyllites and schists therefore could have
Figure 20. Samfrau orogenic belts, including the Curepto, Pichilemu, and Andean orogenic belts of Chile (compiled from DuToit, 1937; Craddock, 1970; González-Boronino, 1970; and Ruiz and others, 1965; and assuming that the Precambrian and Paleozoic rocks shown by Ruiz and others, 1965, are equivalent to the Paleozoic metasedimentary rocks of central Chile, as mapped by González-Boronino. 1970. and dated by Munizaga. 1967).
been formed during either Permo-Triassic or Mesozoic thermal metamorphism, or both.

Double plunges of fold axes at Romeral, as well as steeply plunging crenulations, may have both resulted from Pichilemu cross folding during early Triassic time. Alternatively, the crenulations may have formed in response to later strike-slip faulting. Nevertheless, the structural trends and probable late-Paleozoic to early Mesozoic ages of the Chilean metamorphic basement rocks suggest these rocks may have been formed in the Samfrau Orogenic Zone of DuToit (1937). A more recent reconstruction of the Samfrau Zone by Craddock (1970) shows it to comprise several orogenic belts that formed along the southern margin of the Gondwana supercontinent (Fig. 20).

Curepto sedimentation and orogeny, reflected in the rocks of the Chilean basement complex, were roughly contemporaneous with deposition and orogeny during middle to late Paleozoic time in the Borchgrevink-Tasman belt at the far eastern end of the Samfrau Orogenic zone. Pichilemu deformation and metamorphism were roughly contemporaneous with orogeny in the Sierra-Cape-Ellsworth orogen, which extended across Argentina, the tip of South Africa, and part of Antarctica, during early Mesozoic time (Fig. 20).

La Escoba Andesite Porphyry of Jurassic Age

Dark gray volcanlastic andesite porphyry rests with angular unconformity on well foliated, steeply dipping
biotite schists near La Escoba. This rock may be correlative with andesite of Jurassic age found north and south of El Romeral but not previously recognized in the vicinity of El Romeral. La Escoba andesite was found during reconnaissance mapping north and west of El Romeral, near La Escoba, and on the lower eastern slopes of Cerro Juan Soldado.

Textures of the dark, La Escoba andesites range from aphanitic to porphyritic, with as much as 25 percent plagioclase phenocrysts (.5 to 2.5 mm) in an actinolitized, biotitized dark greenish to brownish gray, micro-fragmental groundmass. No layering, pillow structures, or vesicles were observed near El Romeral, but steep cross-bedding was seen in similar greenish gray, micro-fragmental, volcaniclastic andesites at El Tofo.

At La Escoba the dark gray andesites are cut by actinolite-apatite-magnetite veins and are altered to actinolite-scapolite-muscovite-chlorite-bearing rocks near the veins. Farther from the veins, they contain secondary actinolite and biotite. Average mineralogical and chemical compositions of andesites from near La Escoba are given in Table 3.

About 1 km east of La Escoba, near the saddle between Quebrada Chorillos and Quebrada Agua del Mulato, a thick section of dark gray andesite with nearly horizontal joints is exposed on the west side of the gulch. On the opposite side of the gulch are vertically foliated biotite schists. The schist-andesite contact is covered by alluvium in the
Table 3

MINERALOGICAL AND CHEMICAL COMPOSITIONS* OF LA ESCOBRA ANDESITE PORPHYRY

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Mean Vol.%</th>
<th>Mean An</th>
<th>Constituent</th>
<th>Mean Wt.%</th>
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<td>CaO</td>
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</table>

*Mean compositions are based on modal analyses of four samples, 600 points each, and 1500 points each for opaques (see Appendix A, Tables A-3, A-4, and A-5).
gulch, which lies along the northward projection of the Romeral fault. On the east side of the gulch a small mass of dark greenish gray andesite appears to rest unconformably on schist. The schist has pervasive, steeply dipping schistosity, the contact is nearly horizontal, and the overlying porphyritic andesite is not schistose.

On the lower southeastern slopes of Cerro Juan Soldado, the dark gray andesites are cut by trondjemite and quartz monzonite of the Cerro Juan Soldado batholith. On the lower western slopes of Cordón La Liga, they are cut by gray-green La Liga intrusive andesite porphyry.

The dark gray andesites probably are Jurassic in age. They are younger than the Paleozoic schists, older than the Juan Soldado batholith, and older than La Liga intrusive andesite porphyry. Those of La Escoba are similar to Jurassic black andesites of Hoja Ovalle, near Tongoy and about 60 km south of El Romeral. These rocks were briefly described by Thomás (1967, p. 17) and tentatively correlated with the Fm La Negra of early Jurassic age, as named and described by Garcia (1964) for exposures near Antofagasta. The dark gray andesites of La Escoba also are similar in appearance to altered black andesites of El Tofo, which are designated as volcanic andesites of Jurassic age on the geologic map of Chile (Ruiz and others, 1968). Micro-fragmental texture in some of the Tofo black andesites, and differentially welded cross-bedding observed in one outcrop, indicate that these rocks are volcanlastic.
According to paleogeographic maps by Ruiz and others (1965), both El Tofo and El Romeral lie along the projection of a Jurassic eugeosynclinal facies which consists of island-arc extrusive andesites. Jurassic volcanic breccias and turbidites (trench facies) are found to the north between Chañaral and Taltal. Jurassic andesite porphyries interlayered with continental conglomerates of the Fm Algarrobal are present to the southeast in Quadrangulo Vicuña (Dediós, 1967). Marine Jurassic limestones of miogeosynclinal or back-arc facies are found still farther east.

No Jurassic sections have been measured near El Romeral. Thicknesses reported by Ruiz and others (1965) for exposures north of Copiapó range from "several thousand meters" of volcanic andesite in the eugeosynclinal, island-arc facies along the west coast, to about 2000 m of limestone in the eastern miogeosynclinal area. An east-west cross-section (Ruiz and others, 1965) at 26° S shows the full eugeosynclinal Jurassic section to be 9000 m thick. Taking 9000 m as a maximum and 2000 m as a minimum, an intermediate value of 5500 ± 3500 m is suggested as a reasonable estimate for the thickness of the Jurassic section at El Romeral prior to post-ore erosion.

Cerro Juan Soldado Batholith of Late Jurassic or Early Cretaceous age

During a rapid reconnaissance traverse from El Romeral west to the coast, it was noted that the eastern margin of
the Cerro Juan Soldado quartz monzonite batholith, about 4 km west of El Romeral, is cut by magnetite-apatite veins. The batholith is therefore probably pre-ore in age, and it is probably not correlative with the post-ore Punta de Piedra granodiorite batholith east of El Romeral, which cuts actinolitized, magnetite-bearing schists but is not actinolitized and is not cut by magnetite-actinolite-chlorapatite veins, even though it is very close to El Romeral.

The metasedimentary rocks, andesites, diorites, and ore deposits of El Romeral thus are situated between two batholiths of different ages, and hence they do not constitute a large roof-pendant in a single large granodioritic batholith. Furthermore, neither of the two batholiths is spatially, temporally, or compositionally as closely related to the Romeral ore deposits as are the Romeral diorite and the intramineral diorite dikes.

Rocks of the Cerro Juan Soldado batholith crop out west of El Romeral from the eastern slopes of Cerro Juan Soldado to the coast (Figs. 7 and 8). The batholith is about 10 km in exposed width and about 23 km in length. It consists of medium-grained quartz monzonites. An eastern border zone about 250 m wide is somewhat more calcic and less silicic than the major central portion of the batholith. A satellite body of trondjemite east of the batholith may represent a magmatic differentiate of the quartz monzonite. Mineralogical and chemical compositions of typical quartz monzonite and trondjemite samples are given in Appendix A, Tables A-6
The Cerro Juan Soldado quartz monzonites cut La Escoba andesites of Jurassic age but do not cut La Liga andesite porphyries of lower Cretaceous age. Although the pertinent evidence is meager, it would appear that the Cerro Juan Soldado batholith is late Jurassic or earliest Cretaceous in age.

**Fm Arqueros of Early Cretaceous (Neocomian) Age**

Submarine andesite sills and flows of the Fm Arqueros probably were present at El Romeral when the magnetite deposits were formed, but they have since been eroded away (Fig. 81, p. 276). Arqueros andesites are still present southeast of El Romeral, in Quadrángulo Lambert (La Serena), and to the east in Quadrángulo Quebrada Marquesa, where they range in thickness from 850-1230 m (Aguirre and Egert, 1965). Because the Arqueros section generally thickens westward, it is estimated that these andesites were at least 1250 m thick at El Romeral when the Romeral magnetite deposits were formed.

Most of the Arqueros andesites that were briefly examined at Cerrillos, southeast of La Serena, are dark gray-green aphanitic to porphyritic rocks. Some, called ocoites, contain large plagioclase phenocrysts (1-2cm) in a greenish-black aphanitic groundmass that is sparsely pocked with spherical vesicles 1-3 mm in diameter. Such ocoites are thought to represent submarine sills that were intruded into soft marine sediments just below the water-sediment interface.
(Franquesa, personal communication, 1970). At Cerrillos the ocoites are interlayered with thicker sections of less spectacularly porphyritic andesites and with thin lenses of limestone and chert that generally thicken and become more continuous to the east. Marine fossils from the limy units have been dated as Neocomian in age (Aguirre and Egert, 1965).

No detailed petrographic descriptions of Arqueros andesites are available, but 18 chemical analyses of these rocks from Oyarzún and Villalobos (1969) are summarized in Appendix A, Table A-12.

La Liga Andesite Porphyry of Early Cretaceous (Neocomian) Age

La Liga andesite porphyry is the host rock of the Main orebody at El Romeral. It crops out between the east and west masses of Romeral diorite and west of the west mass of Romeral diorite, along the entire ridge and west side of Cordón La Liga.

Outside the intensely actinolitized aureole that surrounds the Main orebody, La Liga andesite porphyry contains about 5 to 25 volume percent of zoned plagioclase phenocrysts (1-3 mm) in a gray-green microcrystalline groundmass composed of primary plagioclase, diopside, hornblende, and magnetite-ilmenite, along with secondary actinolite, sphene, clinozoisite, and penninite (Table 4). The rock is cut in most places by stockworks of actinolite and clinozoisite veinlets, many of which contain magnetite and some of which also contain apatite and sphene.
The primary textural and compositional characteristics of the La Liga andesite porphyry are nearly uniform (Appendix A, Tables A-10 through A-13), and La Liga andesite porphyry has little foliation, few primary mappable internal contacts, no pillow structures, and no vesicles. It is thought to represent a single large andesitic intrusion rather than a series of andesite dikes, sills, or flows. It cuts Paleozoic rocks and Jurassic volcanic andesites, but it is pre-ore in age. Inasmuch as the ore is older than the Punta de Piedra batholith, which probably is late Neocomian in age, the La Liga andesite porphyry probably is early Neocomian in age. The submarine andesite sills and flows of the Fm Arqueros also are early Neocomian in age, and the La Liga intrusion may represent a body of magma that was subjacent to these sills and flows during their formation.
Table 4

MINERALOGICAL AND CHEMICAL COMPOSITIONS*
OF LA LIGA ANDESITE PORPHYRY

<table>
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<tr>
<th>Mineral</th>
<th>Mean Vol.%</th>
<th>Mean An</th>
<th>Chemical Constituent</th>
<th>Mean Wt.%</th>
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* Mean compositions are based on eight analyses (Appendix A, Tables A-10 and A-11).

As shown in Table A-12, Appendix A, the La Liga andesite porphyry contains less iron and more silica than the Arqueros andesites. If the La Liga and Arqueros magmas were initially similar, compositional differences could have been developed during different cooling and crystallization histories. Whereas the Arqueros andesites probably crystallized rapidly and without much differentiation, the La Liga intrusive andesites probably crystallized more slowly and underwent moderate differentiation along the calc-
alkaline trend.

Microdiorite Dikes

Dikes of gray-green, aphanitic augite microdiorite cut the La Liga andesite porphyry and are cut by Romeral diorite near the crest of Cordón La Liga. The microdiorite is composed of primary microcrystalline plagioclase, augite, and minor magnetite-ilmenite, along with secondary actinolite (Tables A 14 through A 16, Appendix A). The dikes range in thickness from less than a meter to nearly 200 meters. Reconnaissance mapping along the ridge suggests that they are arranged in a crudely radial pattern, roughly centered upon the lower western slopes of Cordón La Liga.

Ocoite Dikes

Steeply dipping dikes of ocoite containing large (1-2 cm) plagioclase phenocrysts in a dark gray-green groundmass, crop out near the crest of Cordón La Liga. They range in thickness from less than one meter to about 100 meters. They cut the La Liga andesite porphyry and augite microdiorite dikes, but they are cut by Romeral diorite.

Textural features along with modal analyses and chemical compositions are given in Tables A 17 and A 18, Appendix A. Except for their lack of vesicles, the ocoites of Cordón La Liga are similar to those of Fm Arqueros. It therefore seems possible that ocoite dikes of Cordón La Liga fed
ocoite sills of the Fm Arqueros.

Romeral Diorite
of Early Cretaceous (Neocomian) Age

Romeral diorite is spatially, temporally, and genetically more closely related to the Romeral orebodies than any other large pluton in the vicinity. It forms the largest major pre-ore igneous unit in the area, and two large masses of it crop out near the orebodies. The large west mass of Romeral diorite is about 150 m west of the Main and North orebodies, and the smaller east mass is about 300 to 400 m east of the Main orebody. The two masses probably merge at depth, so that the Main orebody occupies a re-entrant, or roof-pendant of La Liga andesite porphyry between them.

The Romeral diorite is a fine- to medium-grained equigranular rock composed of gray plagioclase crystals and minor augite and magnetite-ilmenite in a dark green actinolitic mesostasis containing some clinozoisite and sphene (Table 5, and Tables A 19, A 20, and A 21, Appendix A). Both east and west masses of the diorite are cut by stockworks of actinolite veinlets and by thick actinolite-magnetite-chlorapatite veins. West of the orebodies, the diorite is altered to a white, aplitic quartzo-feldspathic rock.

The Romeral diorite cuts La Liga andesite porphyry. It is coarser grained and contains somewhat less iron and more
silica than the porphyry, but otherwise the two rock types are almost identical. This suggests that they had a common source, and that the Romeral diorite may represent a resurgence of andesitic magma into already crystalline La Liga andesite porphyry. The minor textural and compositional differences between the two rocks could have resulted from differences in cooling histories. The coarser grain size of Romeral diorite suggests that it cooled more slowly than the La Liga andesite porphyry, perhaps because it was intruded into already-hot wall rocks. The slight compositional differences between Romeral diorite and La Liga andesite porphyry suggest that calc-alkaline differentiation proceeded somewhat further in Romeral diorite than in La Liga andesite porphyry.
**Table 5**

MINERALOGICAL AND CHEMICAL COMPOSITIONS* OF ROMERAL DIORITE

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<th>Mean An</th>
<th>Chemical Constituent</th>
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* Mean compositions are based on five analyses (Appendix A, Tables A 19 and A 20).
Post-diorite, Pre-ore Dikes

Several thin, widely scattered, steeply dipping, hydrothermally altered dikes cut the east and west masses of Romeral diorite (Fig. 21). The fact that these dikes are altered suggests that they were emplaced before hydrothermal alteration related to ore deposition. Thus they were emplaced during a relatively short time span, after diorite crystallization, but before ore deposition. Compositionally these dikes all have dioritic affinities, but some are iron-enriched, and others are iron-depleted with respect to Romeral diorite. Dikes of diorite pegmatite have about the same chemical composition as Romeral diorite, but dikes of porphyritic meladiorite contain more iron, more phosphate, and less silica. By contrast, dikes of aplitic quartz diorite, dacite porphyry, and albite-actinolite porphyry contain less iron, and more silica than Romeral diorite. Complete modal analyses, chemical compositions, and textural descriptions are given in Appendix A, Tables A-22 to A-30. Although the post-diorite, pre-ore dikes constitute only a small total volume of rock, they are petrogenetically important in that they suggest that both iron-enriched and iron-depleted magmas existed below El Romeral just prior to ore deposition.
Figure 21. Post-diorite, pre-ore dikes.
INTRAMINERAL DIKES

Many dikes cut the Main orebody at El Romeral. Some of these are younger than all ore mineralization, but many are cut by magnetite veinlets, and others are substantially altered and replaced by magnetite. These mineralized dikes are here termed intramineral dikes because it is inferred that they were emplaced after ore deposition had begun but before it had ceased. Their rocks are dioritic in composition but aplitic (saccharoidal) in texture, and they are herein called dioritites. According to Johannsen (1939) this term was suggested by Polenov for diorite aplites, and it is used here to prevent confusion with other rocks that have been called aplites at El Romeral.

From a practical point of view it is useful to know the distribution of the intramineral dikes as an aid to the prediction of ore dilution in mining. The dikes are of special geologic interest in that they provide samples of magma that is inferred to have existed below El Romeral during ore formation.

The intramineral dikes are arranged in an en-echelon pattern. They consistently trend N 30-50 W and dip steeply (Fig. 22). There is little or no offset across them, and their textures are undeformed. The dikes commonly have lenticular, subparallel apophyses, and they typically contain unrotated, splinterlike inclusions of magnetite ore, oriented parallel to their walls. These features suggest
that the intramineral dikes filled tensional gash fractures. The orientation of the dikes with respect to the Romeral fault further suggests that such gash fractures might have opened in response to left-lateral movement on the fault.

Four generations of intramineral dikes cut the Main orebody. As shown in Figure 22, they are: IM-1 dioritite porphyry; IM-2 dioritite; IM-3 dioritite; and IM-4 porphyritic dioritite. Modal analyses, textural descriptions, and chemical compositions of samples from the four sets of intramineral dikes are given in Appendix A, Tables A 31 to A 42.

Diorite porphyry dikes (IM-1) are predominantly post-ore in the south-central part of the Main orebody where they are scapolitized and cut by thin magnetite-actinolite veinlets. However, dikes of this same type are substantially replaced by magnetite and actinolite in the northern part of the orebody. These dikes are clearly cut by intramineral (IM-4) dikes, but no direct cross-cutting relationships were observed between IM-1 and IM-2, or IM-3 dikes. The intramineral (IM-2) dioritite dikes are post-ore on the west side of the orebody, but on the southeast side of the orebody they have been partially replaced by magnetite, actinolite, clinozoisite, scapolite, and pyrite. These dikes are absent from the central part of the orebody where they may have been completely replaced by magnetite and actinolite. Intramineral dioritite dikes of the IM-3 type are predominantly post-ore at the outer edges of the
Intramineral dikes in the Main orebody.
orebody, but they are partially replaced by magnetite in the central part of the orebody, where ore grade changes from about 50 percent of Fe to 20 to 35 percent of Fe across the trend of the largest IM-3 dike. The late intramineral, porphyritic dioritite dikes (IM-4) cut IM-1, IM-2, and IM-3 dikes. They are predominantly post-ore in age, but they are cut by magnetite-actinolite veinlets, which in turn are cut by veinlets containing oligoclase, chlorite, calcite, quartz, magnetite, chalcopyrite, chalcocite, and covellite.

The intramineral dioritites are composed of fine-grained plagioclase (An$_{30}$ ± 5), actinolitic hornblende, and disseminated titanomagnetite. Secondary constituents include abundant actinolite and various amounts of chlorite, scapolite, magnetite, pyrite, and chalcopyrite. Most of the amphibole in dikes within the Main orebody is secondary actinolite, but southeast of this orebody the dikes contain about 15 weight percent of actinolitic hornblende. This actinolitic hornblende is thought to be primary because it is phenocrystic and is rimmed by later actinolite and chlorite (Fig. 23).

Most intramineral dikes contain about 2 to 4 volume percent of disseminated, subhedral titanomagnetite grains (Figs. 23 and 26). These grains are pale pinkish to brownish gray in reflected light, and they are slightly anisotropic. Their unit cell dimension, as measured by X-ray diffraction, is 8.408 Å, as compared to 8.395 Å for pure magnetite. Emission spectrographic analysis indicates
that these grains contain 1500 to 2000 ppm Ti, 500 to 1000 ppm V, 500 to 700 ppm Mn, 500 ppm As, 200 ppm Zn, 100 ppm Sb, 1000 to 10,000 ppm Ca, and 2000 to 5000 ppm Mg (Appendix F). The high Ca and Mg values may in part reflect minor plagioclase and actinolite in the analyzed samples. However, the otherwise high impurity contents of these magnetite grains, together with their expanded lattices, suggest that they formed at high, probably magmatic temperatures. Some disseminated titanomagnetite grains contain microscopic intergrowths of titanomaghemite, pseudobrookite, hematite, and ilmenite (Fig. 23), which suggests partial oxidation of the impure magnetite grains at submagmatic temperatures (Lindsley, 1961-62).

In addition to magmatic titanomagnetite grains, intramineral dikes commonly contain inclusions of magnetite. Intricately shaped, skeletal grains of isotropic magnetite, some of which contain wide hematite lamellae, are abundant near the margins of some intramineral dikes. The isotropy of such grains and their concentration near dike margins suggest that they are inclusions of ore magnetite. The delicate skeletal shapes of these grains suggest that they have been partially dissolved, and their hematite lamellae indicate that they have been partially oxidized (Figs. 27, 40 and 41).

Intramineral dikes also are cut by magnetite veinlets and are locally replaced. Magnetite-actinolite veinlets generally have non-matching walls and contain isotropic
hydrothermal magnetite. Peripheral to the veinlets, grains of anisotropic (magmatic) magnetite commonly have overgrowths of isotropic (hydrothermal) magnetite (Figs. 23a and 24). Both the unmatching veinlet walls and the magnetite overgrowths suggest some replacement of dioritite by magnetite. Some intramineral dikes are so substantially replaced by magnetite, actinolite, clinzoisite, scapolite, and pyrite that their original textures have been almost obliterated (Fig. 25).
Figure 23. Generalized sketch showing microtextures observed in intramineral (IM-3) diorite dikes in the southeastern part of the Main pit (310 to 370 m elevations). The enlarged sketches illustrate the following:

a) magmatic titanomagnetite grain, partly oxidized to maghemite, and overgrown by low-Ti, hydrothermal magnetite; (b) equant, magmatic titanomagnetite grains with thin hematite lamellae along octahedral planes; (c) magmatic titanomagnetite grain, partly oxidized to maghemite, and possibly partly resorbed; (d) intergrowth of titanomagnetite, ilmenite, magnetite, pseudobrookite, and maghemite; (e) olivine-green actinolitic hornblende, overgrown by blue-green actinolite, overgrown by gray-green chlorite.
Figure 24. Actinolite-magnetite veinlet cutting scapolitized intramineral (IM-1) diorite porphyry, southwestern Main orebody, 320 m elevation.

Figure 25. Intramineral (IM-2) diorite, largely replaced by magnetite, scapolite, actinolite, apatite, clinozoisite, and chlorite. This sample is from an altered intramineral (IM-2) diorite dike on the southwest side of the Main orebody, at the 320 m elevation.
Figure 26. Equant, magmatic titanomagnetite grains in intramineral (IM-4) porphyritic dioritite, southeastern Main orebody, 310 m elevation.

Figure 27. Contact of intramineral (IM-4) porphyritic dioritite with magnetite ore, northwestern Main orebody, 310 m elevation. The shapes of the magnetite inclusions suggest partial magnetite resorption.
Two explanations can be offered for the veinlets and replacement deposits in the intramineral dikes. One is that dike-emplacement caused remobilization of ore minerals which were then redeposited in the dike rocks. The other is that the dikes were emplaced during ore deposition and were altered and replaced during post-dike hydrothermal alteration and ore deposition. Both processes probably occurred, but post-dike hydrothermal replacement probably was the dominant process. If replacement bodies in the intramineral dikes had formed predominantly by remobilization of iron from the ore, then one might expect that the replacements would have been more evenly distributed through the dike rocks than they are. The replacement bodies presumably would have been just as abundant in the dikes that cut the western part of the orebody as in the dikes that cut the eastern and central parts of the orebody, but they are not. Furthermore, the replacement bodies presumably would have been just as abundant in the later dikes as in the earlier dikes, but they are not. Thus, it would appear that post-dike hydrothermal replacement was the dominant process in the formation of replacement bodies in the intramineral dikes. Assuming this to be true, it would appear that replacement fronts shifted from west to east as the Main orebody grew, and then retreated toward the central part of the orebody as the Romeral thermal gradient decayed, and ore deposition waned. Furthermore, inasmuch as early intramineral dikes on the western side of the Main orebody are
altered to scapolitic rocks, whereas somewhat later intramineral dikes are replaced by magnetite, actinolite, clinozoisite, and pyrite on the western side of the orebody, and late intramineral dikes are typically cut by chlorite-magnetite-chalcopyrite-bearing veinlets, it would appear that the compositions of the Romeral hydrothermal fluids changed with time as the orebodies formed.

The dioritic compositions of the intramineral dikes suggest that they are genetically related to the Romeral diorite. The fact that they were emplaced later than the Romeral diorite suggests that they may represent late, residual diorite magmas. The fact that some intramineral dikes contain over 15 weight percent of primary amphibole suggests that these residual magmas were rich in water (Holland, 1972, p. 282). The fact that intramineral dioritites contain about 7 to 9 percent of iron in primary minerals suggests that these water-rich residual diorite magmas also were iron-enriched with respect to both Romeral diorite and La Liga andesite porphyry. Finally, the fact that steeply dipping intramineral dikes were emplaced at several times during ore deposition suggests that a body of iron-rich, water-rich residual diorite magma existed below the Romeral ore deposits as they formed.
POST-ORE ROCKS

Late Dikes

Dikes of many kinds cut the Romeral magnetite deposits and the altered rocks that surround them (Fig. 28). They include numerous biotite diorite dikes, several albitite dikes, several hornblende and hornblende-biotite lamprophyre dikes, a few quartz-albite pegmatite dikes containing tourmaline, at least one nepheline-bearing oligoclase-arfvedsonite-biotite dioritite dike. Modal analyses and chemical compositions of such late dikes are given in Tables A 43 and A 44, Appendix A.

Most of the late dikes are thin (10 cm to 2 m), discontinuous, and steeply dipping. Their ages relative to one another are not well established. They are absent from the post-ore Punta de Piedra batholith, which suggests that they were all emplaced during a relatively short time span after ore deposition was completed. They constitute only a small amount of rock, but they are petrogenetically important because they represent extreme igneous differentiation in a limited area and over a short time interval. Most of them contain more sodium and less calcium than the earlier igneous rocks, but some are silica-rich and iron-poor, whereas others are silica-poor and iron-rich. Total iron-oxide contents of the late dikes range from only 0.5 weight percent of $\text{Fe}_2\text{O}_3 + \text{FeO}$ in quartz-albite pegmatites to about 9.5 weight percent in biotite dioritites and 21 weight per-
EXPLANATION

1. porphyritic diorite
2. quartz-albite pegmatite
3. quartz-albite aplite
4. quartz-albite granite
5. biotite diorite porphyry
6. porphyritic biotite diorite
7. biotite diorite
8. biotite diorite, hornblende-biotite lamprophyre
9. olivine-biotite diorite, hornblende-lamprophyre

Figure 26. Late dikes. Most dip steeply.
cent in nepheline-bearing diorite. Silica contents range from 76 percent of SiO₂ in quartz-albite pegmatite to about 53 percent in biotite diorite and only 48 percent in nepheline-bearing diorite.

These extreme compositional variations among dikes irrupted in a small area and over a short time span suggest that by the time the Romeral magnetite deposits had formed, extreme compositional gradients had developed in a subjacent body of residual magma.

**Punta de Piedra Batholith**

A granodioritic batholith crops out 150 to 1500 m east of the Romeral orebodies. It is hundreds of kilometers long (N-S) and ranges from 10 to 40 kilometers in exposed width (Fig. 7). Three sub-units of the batholith are present near El Romeral: (1) a discontinuous marginal quartz-bearing diorite, which is tentatively correlated with La Totora gabbro-diorite-granodiorite mapped by Aguirre and Egert (1970) along the eastern margin of the batholith; (2) a granodiorite that is continuous with the Punta de Piedra granodiorite mapped by Aguirre and Egert (1970) southeast of El Romeral, and which makes up the bulk of the batholith; and (3) pods of granite aplite that are scattered within the batholith, particularly along its western margin (Fig. 10).
La Totora quartz-bearing diorite: The La Totora quartz-bearing diorite is texturally and chemically similar to Romeral diorite, but there are several subtle mineralogical differences. The La Totora diorite contains more quartz than the Romeral diorite, and its plagioclase is somewhat more calcic (An$_{54-34}$). It also contains hornblende, salite, fluorapatite, and microcline, whereas the Romeral diorite contains actinolite, augite, chlorapatite, and no potassium feldspar (Tables A 45 to A 47, Appendix A).

Gradational contacts between the La Totora quartz-bearing diorite and the Punta de Piedra granodiorite suggest that the La Totora unit is a marginal facies of the granodioritic batholith.

Punta de Piedra granodiorite: The Punta de Piedra granodiorite is a medium-grained rock composed of zoned plagioclase (An$_{64-23}$), microcline, hornblende, biotite, and quartz, with minor salite, magnetite, ilmenite, and fluorapatite (Tables A 48, A 49, and A 50, Appendix A). Mapping by Aguirre and Egert (1970) in Quadrángulo Lambert (La Serena) southeast of El Romeral, together with reconnaissance in Quebrada Chacay northeast of El Romeral, indicates that the Punta de Piedra granodiorite constitutes the bulk of the batholith for distances of 7 to 10 km east of El Romeral.

Granite aplite: The granite aplite is a saccharoidal rock composed of microcline, quartz, plagioclase (An$_{32-15}$),
biotite, and magnetite (Tables A 51 and A 52, Appendix A). An irregularly shaped body of this rock, 1 km long and 300 m across, cuts the La Totora quartz-bearing diorite and Punta de Piedra granodiorite near the western margin of the batholith, east of the North orebody. Granite aplite also is present along the western margin of the Punta de Piedra granodiorite north of El Romeral. The aplite typically contains more silica and potash, and less calcium, magnesium, sodium, and iron than the Punta de Piedra granodiorite, of which it probably is a late, water-rich differentiate.

Alteration and mineralization: The quartz-bearing diorite, the granodiorite, and the granite aplite crop out within about 150 m of the North orebody, and the granodiorite cuts actinolitized schist that contains magnetite pods. But these batholithic rocks contain no magnetite pods, and they are not cut by magnetite-bearing veins or veinlets. Nor are they actinolitized like the wallrocks of the Romeral orebodies, or silicified and feldspathized like the Romeral diorite west of the orebodies. Instead they are locally argillized where they contain minor disseminated copper sulfides, and in one place they are cut by white quartz veins, some of which have been mined on a small scale for gold.

Some quartzites near the contact of the granodioritic batholith are altered, but their secondary minerals are slightly different from those of the dioritized quartzites
near the Romeral diorite contact or the altered rocks near the Romeral orebodies. For example, the quartzites near the contact of the granodioritic batholith contain secondary andesine rather than oligoclase, epidote rather than clinozoisite, fluorapatite rather than chlorapatite, reddish-brown biotite rather than greenish-brown biotite, and distinctive actinolite that is microscopically deep blue-green in the Z vibration direction. They also contain more ilmenite than magnetite, whereas the altered rocks associated with the Romeral orebodies contain more magnetite than ilmenite. Both the quartzites near the granodioritic batholith and the altered rocks near El Romeral, however, contain pyrite and sphene.

**Diabase dikes:** The granodioritic batholith is cut by dark gray diabasic dikes that contain 0 to 20 percent of calcic plagioclase phenocrysts (An\textsubscript{80-50}) and 0 to 5 percent of ferrosalite phenocrysts in a microcrystalline groundmass composed of plagioclase, biotite, ferrosalite, magnetite, and ilmenite. Secondary constituents include amphibole and chlorite after ferrosalite and epidote, calcite, and clay after plagioclase.

**Probable age:** The quartz-bearing diorite, the granodiorite, and the granite aplite are neither actinolitized nor cut by magnetite veins or pods, even where they lie near the North orebody, and therefore they are considered to be post-ore in age. Aguirre and Egert (1970) reported a lead alpha
age of $136 \pm 20$ m.y. on zircon from Punta de Piedra granodiorite. The median figure, 136 m.y., corresponds to the Jurassic-Cretaceous boundary (Casey, 1964), or to earliest Neocomian (Berriasian) time. However, the $\pm 20$ m.y. time range extends from middle Jurassic to middle Cretaceous (Aptian-Albian) time. Aguirre and Eger also reported that the Punta de Piedra granodiorite intrudes rocks at least as young as the uppermost limestone-bearing member of the Fm Arqueros ($K_{a_5}$), which contains the rudistid pelecypod Agria Blumenbachii (Struder), a good index fossil for latest Neocomian (Barremian) time. Casey (1964) suggested a time scale based on the limits 64-136 m.y. for the Cretaceous, and on the assumption that the 12 designated stages were of approximately equal duration (6 m.y. each). According to this scale, Barremian time lasted from 118 to 112 m.y. ago. Actual K/Ar dates for glauconites from Hauterivian, Barremian, and Aptian sediments range in age from 115 to 94 m.y. and average 106 m.y. (Casey, 1964). It would therefore appear that Barremian time is probably within the age range between 118 and 106 m.y. Thus Punta de Piedra granodiorites, which cut Barremian limestones, must be near the young end of the 116-156 m.y. age range suggested by lead-alpha age dating. A reasonable compromise that overlaps all three suggested age ranges would be $116 \pm 10$ m.y. for the age of the granodiorite batholith. The Romeral ore deposits, which are older than the Punta de Piedra batholith, and which cut andesite porphyries of probable early Neocomian age,
therefore are probably between 116 ± m.y. and 136 m.y., or middle Neocomian in age.
Figure 29. Generalized geologic map showing the locations of the Romeral magnetite deposits with respect to major rock units and faults.
Magnetite-bearing deposits of the Romeral area are shown in Figure 29. From south to north, they are:

1. South Extension orebody. South of the Main orebody, in the Romeral fault zone.

2. Main orebody. In a fractured roof pendant of actinolitized andesite porphyry in Romeral diorite.

3. Western ore pods. In actinolitized phyllite west of the Main orebody.


5. Subsurface ore pods between the Main orebody and the North orebody. Along a north-northeast-trending fault in phyllite.


9. Magnetite-bearing actinolite veinlets. Stockworks of magnetite-bearing veinlets are well developed in all major rock units older than the Punta de Piedra-La Totora batholith, especially in Romeral diorite and La Liga andesite porphyry.

10. Magnetite-apatite-bearing actinolite veins. Large veins are scattered within a 3 to 4 km radius of El Romeral, cutting all rocks older than the Punta de Piedra-La Totora batholith.

Detailed descriptions of these magnetite deposits follow.
Figure 30. The Main orebody. Oblique air photo, 1970. Generalized geology is marked in ink. North is toward the upper right-hand corner of the photo.
South Extension Orebody

The South Extension orebody lies along the Romeral fault trend, about 500 m south of the Main orebody. It is about 800 m long (N-S) and 100 m wide (E-W), and it dips steeply. The top of the magnetite zone is about 175 m below the surface.

The South Extension has been explored by magnetometer surveys and reconnaissance drilling, but it has not yet been mined.

Main Orebody

The Main orebody is about 850 m long, 250 m wide (maximum) and has a vertical extent of about 400 m (Figs. 31 and 32). Its western margin trends north-south, parallel to the Romeral fault, and dips steeply. The southeastern boundary trends N 10-30 E, and the northern limit trends about N 50 W, parallel to the steeply dipping contact between La Liga intrusive andesite porphyry and Paleozoic phyllite.

Ore grade ranges from about 20 percent of iron to more than 60 percent of iron. (Pure magnetite, Fe₂O₃, contains 72.4 percent of iron.) At the outer fringes of the Main orebody, disseminated magnetite grains, magnetite pods, and magnetite veins and veinlets are scattered through actinolitized andesite porphyry. In progressively higher grade zones, andesite remnants are progressively smaller and more actinolitized between coalescent magnetite grains, veinlets, and pods.
Figure 31.
Main and North Extension orebodies. This ore-zone map is based on exposures in the Main pit in the year 1970, and on assays of samples from 1450 10 m percussion drill holes on the 320-300 m levels.
Figure 32. Ore-zone section of the Main orebody.
As shown in Figures 31 and 32, the highest grade ores are found along steeply dipping sub-tabular zones that trend N-S, N 30-50 E, and N 20-45 W.

**Ore mineralogy and texture:** Megascopically, ore from the high-grade zones of the Main orebody appears to be almost pure, massive magnetite. Close examination, however, reveals small (0.1-2 mm) intergrowths of actinolite and discontinuous veinlets of green and white gangue minerals. Microscopically, the ore consists of spongy intergrowths of sievelike skeletal magnetite grains and actinolite microprisms with generally smaller amounts of clinozoisite, sphene, apatite, plagioclase (An$_{27}$ ± 5), scapolite, chlorite, and quartz (Figs. 33 and 34). Although the magnetite grains generally are very irregular in shape, the influence of octahedral symmetry is visible in overall shapes of some grains and in short, planar segments of grain boundaries that are mutually parallel or that intersect at 90° or 60°.

Magnetite veinlets are abundant and easily visible at the edges of the orebody, and microscopic magnetite veinlets also are common in even the most high-grade ore (Figs. 35 and 36). The veinlets have irregular, unmatching walls, and actinolitized wallrocks near the veinlets contain disseminated clots of skeletal magnetite. The abundance of veinlets suggest that small fractures and capillary microfractures were important distributaries for the iron-bearing fluids. The narrowness of some of the veinlets suggests that these fluids must have been very tenuous. The unmatching veinlet
walls and envelopes of disseminated magnetite further suggest progressive wallrock replacement adjacent to the veinlets.
Figure 33. Intergrowth of actinolite and poikiloblastic skeletal magnetite, Main orebody.

Figure 34. Spongy intergrowth of skeletal magnetite, actinolite, and chlorapatite, Main orebody.
Figure 35. Magnetite veinlets cutting actinolitized andesite porphyry, eastern margin of Main orebody.

Figure 36. Capillary magnetite veinlets and skeletal magnetite crystals in actinolitized rock, high-grade ore zone in western Main pit.
Composition of ore magnetite: Magnetite from the Main orebody has low content of titanium and vanadium. Polished ore magnetites are very pale pinkish to brownish gray in reflected light and are almost isotropic. These features suggest a low content of Ti (Uyttenbogaardt and Burke, 1971, p. 168). The unit-cell size of magnetite from one ore sample was measured by X-ray diffraction as a =8.398 ± 0.007Å as compared to 8.395 Å for pure magnetite and 8.408Å for magmatic magnetite from an intramineral dike. Spectrographic analysis of two samples of ore magnetite indicate 100 to 500 ppm Ti and 1000 ppm V.

Temperature and mode of formation: The low titanium content of Romeral ore magnetite suggests submagmatic temperatures of formation. This statement must be qualified because the magnetite does not contain intergrowths of ilmenite, as required for strict application of the titanium geothermometer of Buddington and Lindsley (1964). However, the magnetite commonly is intergrown with sphene, and therefore Ti probably was available in the system; if the magnetite had formed at high temperature, its thermally expanded lattice might have accommodated more Ti. The problem is not that simple, however, because substitution of Ti⁴⁺ also requires concomitant reduction of Fe³⁺ to Fe²⁺ to maintain the charge balance, and this process is dependent on fO₂. Furthermore, partition coefficients for Ti between sphene and magnetite are almost certainly not the same as those between ilmenite
and magnetite. Still, submagmatic temperatures of formation also are indicated by the close association of magnetite with actinolite, a typically metamorphic mineral that forms below 550°C (Liou, 1972). Thus, magnetite composition, and actinolite association, suggest that the ore formed at submagmatic temperatures, and ore textures suggest that the ore formed by veinlet filling and wallrock replacement.

**Gangue veinlets:** The Main orebody is interpenetrated by small veinlets of gangue materials (Figs. 37, 38, 39). Actinolite veinlets, the most common type, contain various percentages of magnetite, clinozoisite, apatite, and pyrite. Clinozoisite veinlets, also common, contain various amounts of actinolite, magnetite, sphene, scapolite, apatite, chlorite, pyrite, and chalcopyrite. They are more abundant on the east side of the Main orebody than on the west. Chlorite veinlets contain various amounts of plagioclase, apatite, calcite, quartz, magnetite, pyrite, and chalcopyrite. Green axinite-magnetite-apatite veinlets are rare. White calcite veinlets are common, and they are typically younger than all the green veinlets. They contain subsidiary apatite, sphene, clinozoisite, chlorite, plagioclase, and quartz.

The veinlets are discontinuous and are offset by other veinlets and fractures, indicating recurrent fracturing accompanied by changing and waning gangue-mineral deposition. In general the following age sequence can be observed:
Figure 37. Magnetite ore cut by actinolite veinlets, which are in turn cut by a calcite veinlet and a barren fracture, east Main pit.

Figure 38. Discontinuous actinolite and pyrite veinlets cutting magnetite ore, south Main pit.
1. actinolitic veinlets
2. clinozoisitic veinlets
3. pyritic veinlets
4. chloritic veinlets
5. calcitic veinlets
6. barren fractures

This sequence probably reflects temperature decay with time, along with progressive changes in composition of hydrothermal fluids.

In some areas, recurrent fracturing and faulting have resulted in formation of magnetite breccias. Early magnetite breccias are partially cemented by actinolite, whereas later ones are cemented by actinolite, apatite, and calcite (Fig. 67). Post-ore breccias contain actinolite fragments and have matrices of pulverized magnetite (Fig. 68).

Figure 39. Axinite-chlorapatite-magnetite veinlet cutting magnetite ore, northwest Main orebody.
Figure 40. Partly resorbed skeletal magnetite crystal containing scattered clots of maghemite; from partly resorbed ore near Intramineral (IM-4) dike contact, northwest Main pit.

Figure 41. Hematite lamellae in partly resorbed skeletal magnetite grain in ore near an intramineral (IM-4) dike contact; northwest Main pit.
Partially dissolved and oxidized magnetite grains near dikes:
Magnetite grains in ore adjacent to intraminal and late
dikes and in inclusions of ore within these dikes have
unusually delicate skeletal shapes and commonly contain
intergrowths of hematite and maghemite (Figs. 40, 41). The
delicate shapes of such grains suggest that the grains may
have been partly dissolved, and their hematite and maghemite
intergrowths suggest that they have been partially oxidized.
The fact that these grains are found in ore adjacent to
intramineral and late dikes and in inclusions within these
dikes suggests that their partial dissolution and oxidation
may have resulted from the emplacement of the dikes. These
dikes probably emanated hot water during their crystalliza-
tion (p. 205), and they probably also heated the local
groundwaters, causing limited solution and oxidation of
nearby magnetite grains.

Distribution of impurities in the ore: Phosphorous, sulfur,
titanium, and vanadium are considered to be impurities
because they affect the steel-making characteristics of iron
ore. Phosphorous zoning at El Romeral, based on assays
from 1450 percussion-drill holes on the 300-320 m levels,
indicates that the P content of the iron ore ranges from 0
to about 1 weight percent (Fig. 42). The zones of highest
phosphorous content are in the central and eastern parts of
the Main orebody parallel to and between northeast- and
northwest-trending high-grade iron-ore zones. (Compare
Figs. 31 and 42.) Apatite also is abundant in Romeral fault breccia northeast of the Main orebody (Fig. 70). Ore-zone sections based on deep drilling indicate that phosphorous content of the ore generally decreases with increasing depth (Franquesa, 1970).
Figure 4.2. Distribution of phosphorous in the Main orebody. Zoning is based on assays from 1450 track-drill holes on the 300-320 m levels.
Carbonate-bearing chlorapatite is the only phosphorus-bearing mineral known to be present in the Romeral ore (Appendix C, p. C-4, C-5). Microscopic chlorapatite is intergrown with actinolite and poikiloblastic magnetite, and coarser grained apatite is present with calcite in veinlets that cut the ore. Chlorapatite also is abundant in the Romeral fault zone west of the Main orebody, in the cataclasite zone northwest of the North orebody, and in actinolite-chlorapatite-magnetite veins outside the Romeral orebodies.

Sulfur zoning, based on samples from deep drilling, indicates that the S content of Romeral ore ranges from 0 to about 0.1 weight percent, with small zones containing as much as 0.5 weight percent (Franquesa, 1970). Sulfur content of the ore generally is low in the upper, northwestern part of the Main orebody and higher in the southeastern and deeper parts of the orebody. Pyrite is by far the most common sulfur-bearing mineral in the Romeral ore, but traces of chalcopyrite, chalcocite, and covellite also are present. Pyrite most commonly is formed in veinlets that cut the magnetite ore, but minor pyrite also is intergrown with magnetite.

Sphene is the common Ti-bearing mineral in and around the Main orebody. Ilmenite is very rare. Titanium zoning by Franquesa (1970) shows that Ti is somewhat more abundant in the upper and east-central parts of the Main orebody than in the western and lower parts. Thus its distribution
is similar to that of apatite. A sample of ore magnetite from the eastern part of the orebody (500 ppm Ti) was more titaniferous than a magnetite sample from the western part (100 ppm Ti). This suggests that not only is there more sphene in the eastern part of the orebody, but also the magnetite is more titaniferous there.

No vanadium minerals have been identified at El Romeral, but vanadium is present as a trace element in the ore magnetite (500-1000 ppm by emission spectrographic analysis). Vanadium zoning by Franquesa (1970) indicates that V content of the ore is highest in the southeastern and eastern parts of the Main orebody and lowest in the northwestern part.

Oxidized and leached ore: In the upper and northern parts of the Main pit, and in the North pit, and particularly in zones where clay minerals are superimposed on earlier actinolitic assemblages, magnetite has been partly oxidized to hematite (martite). Hematite lamellae are oriented along octahedral planes in magnetite as shown in Figure 43. The close association of martite with clay minerals suggests that martitization and argillization resulted from wallrock reaction with late, acidic, oxidizing hydrothermal fluids. Such fluids may also have dissolved and removed apatite from the rocks of the argillized zone.

Geothite is common in surface exposures of magnetite and along fractures in ores above the water table, which was intersected at about the 320 m level in the Main pit. The
geothite probably formed by weathering of ore in the aerated zone above the water table. Leaching of deleterious pyrite and chalcopyrite from ores in the aerated zone improved the ore quality above the water table. Redeposition of copper near the water table resulted in formation of several pods of supergene copper minerals.

Figure 43. Hematite (martite) after magnetite, cut by geothite-filled fracture, north Main pit.

**Sequence of ore-and-gangue-mineral deposition:** Ore-distribution patterns, together with age relationships between Romeral fault breccia, magnetite ore, intramineral dikes of four generations, and gangue-mineral assemblages in different parts of the Main orebody suggest that: (1) ore deposition began in the southwestern part of the orebody, near the Romeral fault, (2) as replacement fronts progressed northeastward, and possibly upward, apatite-scapolite-clinozoisite-sphene deposition increased, and (3) as replacement fronts retreated to the central and lower parts of the orebody, apatite-scapolite-clinozoisite-sphene
deposition decreased, and pyrite deposition increased. Ore deposition preceded emplacement of intramineral (IM-1) dikes in the southern part of the Main orebody but followed it in the northeastern part. This suggests that ore deposition progressed from south to northeast. Similarly, ore deposition preceded emplacement of intramineral (IM-2) dikes in the western part of the orebody but followed it in the eastern part, and this suggests continued eastward movement of replacement fronts. Inasmuch as apatite, scapolite, clinozoisite, and sphene are more abundant in the upper and eastern parts of the orebody than in its western part, this suggests that rates of deposition of these minerals increased as replacement fronts progressed northeastward. Furthermore, concentration of apatite in the matrix of Romeral fault breccia containing actinolite-magnetite fragments, suggests that apatite deposition followed actinolite-magnetite deposition in the northwestern part of the orebody. Intramineral (IM-3) dikes are post-ore on both sides of the orebody but are partially replaced by magnetite in its central part. This suggests that replacement fronts retreated toward the central part of the orebody during the waning stages of ore deposition. Pyrite veinlets, which cut the ore and the intramineral (IM-3) dikes, are concentrated in the eastern, central, and lower parts of the orebody. This suggests that as replacement fronts retreated toward the central and lower parts of the orebody, pyrite deposition increased. The chlorite-oligoclase-magnetite-chalcopyrite veinlets that cut intramineral (IM-4) dikes record a weak, dying stage of magnetite
deposition that followed cessation of large-scale magnetite-actinolite deposition in the Main orebody. Calcite veinlets formed after magnetite deposition had ceased.

Intimate intergrowths of actinolite with poorly formed, poikiloblastic, skeletal magnetite grains suggest rapid, nearly simultaneous deposition of these minerals. Crystal growth experiments at the Stanford Earth Sciences Laboratory and elsewhere have shown that:

1. Very rapid crystallization results in dendritic crystallization.

2. Rapid crystallization of two or more phases results in intimate intergrowths in which poorly formed, poikilitic, skeletal guest crystals contain many inclusions of host material.

3. Slow growth results in "cleaning up" of the guest phases by exclusion from the host material (R. H. Jahns, personal communication, 1973).

Spongy, skeletal magnetite grains have much larger surface areas than clean, octahedral grains of comparable volume, and they therefore have larger numbers of unsatisfied surface charges. The existence and persistence of these metastable, high-energy grain shapes therefore suggests rapid deposition and no later textural equilibration by annealing.

This does not necessarily imply low pressure or extremely steep temperature gradients that cause dumping in xenothermal deposits. In fact, stratigraphic evidence indicates about 2 kb of total pressure, and well developed, non-telescoped mineral zonation indicates steep, but not extreme,
temperature gradients. Rapid deposition evidently occurred simply because the hydrothermal fluids overshot iron saturation, became supersaturated in some critical degree, and then rapidly "dumped" magnetite until they once again approached equilibrium with magnetite. Continuing temperature decrease made re-solution of magnetite unlikely.

Western Ore Pods

Many pods of iron ore are present west of the Main orebody (Figs. 29, 30, 31). Many are conformable pods of magnetite and martite in amphibolitized phyllites between the Romeral fault zone and the diorite contact. Others are fragmental ore pods caught up in either the early Romeral fault or the late North-northeast fault. Magnetite, actinolite, and apatite breccia fragments in the early Romeral fault are locally surrounded by apatite-rich matrices (Fig. 70). Magnetite-martite fragments in the North-northeast fault zone are imbedded in fault gouge that is locally rich in pulverized magnetite and martite.

North Extension Orebody

Many conformable pods of iron ore occur in amphibolitized, chloritized, and argillized phyllites in the northern part of the Main pit, north of the Main orebody. The largest mapped pod, called the North Extension orebody, is cut off by the North-northeast fault, which also cuts off the south end of the North orebody. Before being displaced
by about 200 m in a right-lateral sense along this post-ore fault, the North Extension and North orebodies probably were semi-continuous. Magnetic anomalies along the fault trend between the orebodies may well represent displaced slivers of magnetite strewn along the fault zone. Such magnetite slivers probably will be encountered as the north end of the Main pit is stripped back to provide stable slopes in the soft, argillized phyllites.

North Orebody

Before it was mined, the North orebody was a semi-continuous group of concordant magnetite-martite pods in steeply dipping actinolitized biotite schists (Fig. 44). Its strike length was about 300 m (N 20 W), its dip length was about 100 m (70°SW), and it was 20 to 50 m thick. The North orebody is now essentially mined out, and all that remain are discontinuous magnetite-martite pods at the bottom and along the west and north sides of the North pit.

Massive ore pods in the North pit have gradational contacts with surrounding actinolitized biotite schists. Relict schistose texture, which is well preserved in the lower grade rocks, also is discernable in some of the high-grade ore pods. This suggests that replacement was the dominant mode of ore deposition in the North orebody.

Although phyllites west and north of the North orebody contain scattered pods of magnetite, most of the ore is concentrated in the coarser grained schists. Within these
Figure 44. Northern magnetite deposits of El Romeral.
Figure 44b. Generalized composite section, showing the North orebody.
schists, magnetite is concentrated along actinolitized biotitic layers (Fig. 14). Evidently the schists were more permeable than the finer grained phyllites to ore-bearing hydrothermal fluids and the biotitic layers of the schists presumably were chemically more susceptible to replacement by actinolite and magnetite than were the quartz-oligoclase layers.

The magnetite of the North orebody, like that of the Main orebody, is pale pinkish-brownish gray in reflected light and is almost isotropic. These characteristics suggest a low Ti content, which, together with the close association of magnetite with actinolite, suggests formation at sub-magmatic temperatures. Texturally, the magnetite grains in the North orebody are typically anhedral, skeletal, and poikiloblastic, which suggests rapid deposition.

Along the western side of the North pit, many skeletal magnetite grains are rod shaped, and their long axes plunge parallel to crenulations in the schist. These rodlike shapes may have resulted in part from preferential magnetite deposition along the axes of crenulation folds. However, textural indications of rolling and shearing during grain growth can be noted, as follows, and as illustrated in diagrams a, b, and c of Fig. 45:

a. Some magnetite grains are broken, and the pieces are slightly offset relative to one another. Other pieces are torn free of the magnetite rods and are strewn behind them.
Figure 45. Perspective sketch of rolled, broken, skeletal magnetite rods in schist between the left-lateral Romeral fault and the North orebody. Arrows show orientation of implied left-lateral shear couple.
b. Biotite booklets are packed against the northern sides of the magnetite rods, are deflected around them, and form trains behind them. This configuration suggests differential movement between the magnetite rods and the biotite booklets.

c. Skeletal magnetite rods, cut perpendicular to their long axes, have curved radial and concentric structural elements that might have resulted from rotational movement during their growth.

These textural characteristics suggest that the rodlike magnetite grain shapes may have resulted in part from rolling and shearing of magnetite grains during their growth. Such rolling and shearing of growing magnetite rods could have been caused by left-lateral movement on the Romeral fault during ore deposition. The fault zone is about 100 m west of the North orebody.

Northern Magnetite-Apatite-Actinolite Cataclasites

Northwest of the North orebody is a wedge-shaped body magnetite-apatite-actinolite-rich cataclasite that is bounded by two branches of the Romeral fault. The western fault branch trends N 20-40 W and separates cataclasite from Romeral diorite. The eastern fault branch trends N 10 W and separates cataclasite from altered phyllite that contains concordant pods of actinolite and magnetite. Both faults dip steeply to the southwest, and attitudes of foliations within the cataclasite wedge are in the range N 15-40 W, 60-80 SW (Fig. 44a).

The rocks of the cataclasite wedge contain abundant
magnetite, actinolite, and apatite, and they have textures that suggest extreme deformation. Schistose cataclasites are predominant in the southwestern part of the wedge, whereas deformed breccias are predominant in its northwestern part. Apatite-rich cataclasite dikes cut these breccias in the central part of the cataclasite wedge.

The schistose cataclasites contain biotite, apatite, magnetite, and axinite. Schistosity is defined by subparallel arrangement of biotite plates, and foliation is defined by interlayering of biotite-rich, apatite-rich, and magnetite-rich layers, lenses, pods, and boudins. Long axes of magnetite boudins plunge steeply to the southwest, as do long axes of apatite grains (Figs. 46, 48).

The deformed breccias contain angular fragments of magnetite-actinolite rock in an apatite-rich matrix (Figs. 47, 49). Subtle foliations are defined by lenticular clusters of fragments and by alternating layers of fine- and medium-grained apatite in the matrix. Pods and veins of magnetite-actinolite rock as large as 10 m thick and 50-100 m long are interspersed with the breccia and are oriented N 20-30 W, 60-80 SW, parallel to the attitude of foliation in the breccia. The pods appear to be large fragments caught up in the breccia, whereas the veins appear to cut the breccia.

The apatite-rich cataclasite dikes range in thickness from about 4 cm to 1 m. Foliation within them is oriented parallel to their walls (N 20-30 W, 60-80 SW) and is defined by alternating fine- and medium-grained layers of apatite
Figure 46. Schistose apatite-magnetite-biotite-axinite cataclasite, as viewed downward from the northeast. Foliation is oriented N 34 W, 70 SW, and lineation plunges steeply to the southwest.

Figure 47. Deformed breccia with angular fragments of magnetite-actinolite in an apatite-rich matrix.
Figure 47. Thin section of schistose cataclasite, cut N 56 E, 90°, perpendicular to the foliation of the cataclasite. Long axes of magnetite grains and grain clusters, and long axes of apatite grains plunge steeply to the southwest (nichols at 45°).

Figure 49. Thin section of deformed breccia with angular fragments of magnetite in an apatite-rich matrix (nichols at 45°).
Figure 50. Mottled breccia (left) intersected by black, magnetite-rich replacement veins (center), and cut by an apatite-rich cataclasite dike (right).
Figure 51. Microscopic view of dike-like apatite-rich cataclasite, looking north. Low birefringence of apatite grains indicates that apatite c axes are nearly perpendicular to the section and therefore plunge gently. (Nichols are a 45°).
grains and by trains of almost microscopic actinolite needles that are deflected around streamlined clusters of magnetite fragments. Long axes of magnetite grain clusters, of actinolite needles, and of apatite grains plunge gently (Figs. 50, 51).

The various textures of the cataclasite suggest extreme deformation concurrent with replacement of phyllite by magnetite, actinolite, and apatite. While mineralization was in progress, left-lateral displacements on the Romeral fault drove the cataclasite wedge between the east branch of the fault and the unyielding block of Romeral diorite (Fig. 72, b, c). Steeply plunging magnetite and apatite rods in schistose cataclasite probably resulted from rolling and stretching of grains caused by the left-lateral displacement. Gently plunging grains of apatite and magnetite in the cataclasite dikes may have resulted from rolling and stretching of grains during upward flowage of apatite-rich materials in the severely squeezed cataclasite wedge. Inasmuch as the cataclasite dikes cut earlier cataclasites, the dikes probably formed during late stages of Romeral ore deposition.

Northern Magnetite Pods in Schist

Many pods of massive magnetite are scattered through biotite schists as much as 3.6 km north-northwest of the North orebody (Fig. 29). These pods are elongate parallel to the host-rock schistosity. Magnetite in and around the pods is commonly accompanied by chlorapatite and actinolite.
Magnetite-bearing Veinlets

Stockworks of magnetite-bearing actinolite veinlets cut all major primary rock units older than the Punta de Piedra batholith, and are particularly well developed in the La Liga andesite porphyry and Romeral diorite. In general the abundance of veinlets and the ratio of magnetite to actinolite in them increase toward the Romeral orebodies. However, such veinlets are essentially absent from the quartzo-feldspathic zone west of the orebodies, where they have been destroyed by post-veinlet alteration.

Magnetite- and Apatite-bearing Actinolite Veins

Actinolite veins containing magnetite and apatite are scattered within a 3 to 4 km radius of El Romeral (Fig. 29). Most of them dip steeply and strike at high angles (45 to 90°) to the trend of the Romeral orebodies. They cut all major igneous rock units older than the Punta de Piedra batholith, and they also cut silicified and feldspathized diorite west of the orebodies. Inasmuch as silicification and feldspathization lagged behind deposition of the magnetite ores, and the actinolite-magnetite-apatite veins cut the quartzo-feldspathic rocks, those veins probably formed later than the orebodies.

The veins range in thickness from about a centimeter to
several meters. They are surrounded by envelopes of fine-grained, actinolitized wallrock that range in thickness from centimeters to about ten meters (Fig. 52). Widths of actinolitized envelopes around the veins generally vary in direct proportion to vein widths and in inverse proportion to their distance from El Romeral.

![Composite sketch](image)

Figure 52. Composite sketch, showing characteristics typical of actinolite-magnetite-apatite veins in the El Romeral-La Escoba area.

Actinolite is the most abundant mineral in the veins and in the metasomatized wallrocks peripheral to them. Near the Romeral orebodies magnetite is more abundant than apatite in the veins, but with increasing distance from El Romeral, the apatite content increases, and the magnetite content decreases. Many veins have been mined on a small
scale for apatite, and at La Escoba, about 3.5 km northeast of the North orebody, several large veins have been extensively worked. The actinolite prisms, which generally are perpendicular to vein walls and hence exhibit crude comb structure, range in size from about 0.5 mm x 5 mm to about 5 cm x 50 cm, depending on vein width. Wallrocks adjacent to the veins are altered to fine-grained actinolite that typically is intergrown with magnetite, apatite, epidote, sphene, quartz, calcite, and chlorite. Large, randomly oriented actinolite prisms with abundant inclusions of magnetite are found adjacent to some veins. Vein actinolite in the La Escoba area is partly altered to talc, but most of it in the Romeral area is unaltered.

Vein magnetite is present in interstices between the actinolite prisms, as small anhedral grains disseminated through the actinolite, in central parts of some veins where it is intermixed and interlayered with apatite, actinolite, epidote, sphene, and quartz, and in veinlets that cut obliquely across the large veins. The magnetite is flat gray and isotropic in reflected light, which suggests that its impurity content is low. Its unit cell edge, as measured by X-ray diffraction, is 8.396 ± 0.004 Å, as compared to 8.395 Å for pure magnetite. Spectrographic analysis indicates about 1000 ppm Ti and 1000 ppm V.

The vein apatite is carbonate-bearing chlorapatite. Large crystals of apatite (1-3 cm), bundles of apatite needles, and anhedral clots of apatite are found between
larger actinolite prisms. Apatite is also found in crustified central portions of veins and in veinlets that cut obliquely across the larger veins.

The needlelike apatite crystals are commonly tubelike in that they contain elongate, microscopic fluid inclusions (0.0001 x 0.006 mm). The fluid inclusions are oriented parallel to the long axes of their host crystals and are almost completely filled with colorless, probably aqueous liquid of low refractive index (approximately 1.345 ± 0.015). No bubbles are visible in most of the tubes, but in larger ones minute bubbles are barely visible at about 1000x magnification. The consistent orientation of these fluid inclusions parallel to the long axes of their host crystals, and their lack of association with the basal cleavage direction of apatite, suggest that these fluid inclusions formed during crystal growth, and are therefore primary.

Fluid inclusions in the vein quartz range in size from 0.003 to 0.009 mm. They are randomly distributed and have subhedral negative quartz-crystal shapes. They consistently contain a small bubble of gas, a 0.001 mm greenish, equant, moderately birefringent crystal, and an orange-brown flake that skitters in response to molecular movement in the fluid. The lack of planar or conchoidal arrangement of these fluid inclusions suggests that they did not form along fractures, but formed before fracturing, during crystal growth. Their consistent four-phase compositions also suggest that these fluid inclusions are primary, and that the solid phases
they contain are daughter minerals that crystallized out of their trapped fluids after sealing (Roedder, 1967).

The magnetite- and apatite-bearing actinolite veins described above have previously been regarded as mafic dikes (Melo, 1965), and as pegmatites that were emplaced as apatite-magnetite-amphibole magmas very rich in highly mobile fluids (Park, 1972). The fact that the vein amphibole is actinolite rather than hornblende suggests that vein formation took place at temperatures below 550°C. Experimental work by Liou (1972) indicates that the upper stability limit for actinolite at 2 kb is about 550°C, and above this temperature hornblende is the stable amphibole. Since no hornblende remnants have been observed in the veins, the actinolite is assumed to be a primary vein constituent, indicative of conditions of vein formation. Considering that iron-oxide magmas have not been observed experimentally at temperatures below 815°C, whereas the veins probably formed at temperatures below 550°C, it seems unlikely that the veins formed from ore magmas. Furthermore, several lines of evidence suggest that hydrothermal fluids were important in vein formation. The presence of primary fluid inclusions in apatite and quartz suggests that aqueous fluids were present during formation of these minerals. The actinolite alteration of wallrocks adjacent to veins further suggests that hydrothermal fluids permeated into the wallrocks and reacted with them. In addition, crustification in central parts of some veins suggests hydrothermal open-
space filling, and the thin veinlets that cut the larger veins imply tenuous hydrothermal fluids rather than viscous iron-oxide-rich magmas. Thus, the actinolite-magnetiteapatite veins of El Romeral are here called veins, rather than dikes, in order to emphasize the direct importance of hydrothermal fluids rather than magmas in their formation.
Figure 53. Alteration zones surrounding the Romeral magnetite deposits.
ALTERED ROCKS

Products of Early Thermal Metamorphism

Oligoclase-diopside rock: Fine-grained, light greenish-gray oligoclase-diopside rock crops out about 400 m south-southeast of the Main orebody (Fig. 53). Relict porphyritic texture is commonly preserved, indicating that the rock formed by alteration of La Liga andesite porphyry. Nearer the orebody, oligoclase-diopside rock is cut by actinolite veinlets and is progressively more altered to actinolite rock. Farther east, clinozoisite and montmorillonite are overprinted on oligoclase-diopside rock. These relationships indicate that alteration of La Liga andesite to oligoclase-diopside rock occurred prior to extensive actinolitization and formation of clinozoisite and montmorillonite.

Approximate mineralogical and chemical compositions of a typical sample of oligoclase-diopside rock are given in Table 6. In Figure 54 the chemical composition of oligoclase-diopside rock is compared with that of the pre-existing La Liga andesite porphyry. Minor addition of SiO₂ and CaO, and minor removal of Al₂O₃ were the only significant chemical changes involved in transformation of the porphyry.

Oligoclase-diopside rock formed early in the Romeral alteration sequence, and diopside is stable at higher temperatures than actinolite (Liou, 1972). Furthermore, diopside requires no OH⁻ for its formation, and conversion of andesite porphyry to oligoclase-diopside rock was nearly
Figure 54. Chemical changes involved in alteration of La Liga andesite porphyry to oligoclase-diopside rock. Circled points on the left side of the graph represent mean compositions calculated from 7 samples, and the error bars represent a range of one standard deviation from the mean.

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<th>Chemical Composition</th>
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isochemical. Therefore, it would appear that the oligoclase-diopside rock resulted from an early, prograde, relatively dry, nearly isochemical thermal metamorphism that preceded retrograde hydrothermal actinolitization.

**Cordieritic rocks:** Phyllites and schists north of the Main orebody and east of the Romeral diorite contain as much as 23 volume percent of poikiloblastic cordierite (Fig. 17). In some rocks near the diorite contact, cordierite porphyroblasts are as large as 2 to 5 mm, and in hand sample they resemble plagioclase phenocrysts. The cordierite evidently was formed in response to prograde thermal metamorphism of the aluminous phyllites and schists, while oligoclase-diopside rock formed from the more calcic andesite porphyry.

**Products of Early Metasomatism**

**Dioritized rocks:** Andesites and quartzites near contacts with Romeral diorite contain poikiloblastic crystals of plagioclase \( \text{An}_{27-32} \), actinolitic hornblende, and diopside. The poikiloblastic nature of these crystals and their distribution along diorite contacts suggests that they formed by contact metasomatism related to diorite intrusion and crystallization. Inasmuch as the metasomatic minerals are similar to those of Romeral diorite, the metasomatized rocks are here termed dioritized rocks (Fig. 55).

Dioritized andesite porphyry zones are as much as 100 m wide. Since the dioritized andesites contain plagioclase
crystals approximately 1 cm across, they have previously been mapped as ocoites (Melo, 1965). However, the poikiloblastic nature of these plagioclase crystals suggests that they are porphyroblasts rather than phenocrysts and that the rock is dioritized La Liga andesite porphyry rather than ocoite.

Quartzites as much as 100 m outside the east mass of Romeral diorite contain poikiloblastic plagioclase, amphibole, and pyroxene crystals (Fig. 10), and the abundance of such crystals increases toward the diorite contact. In apposition, diorite as much as 100 m within the east mass contains dioritized inclusions of quartzite, and the abundance of such inclusions increases toward the quartzite contact.

Figure 55. Dioritized quartzite from 10 m north of the contact between quartzite and east-mass Romeral diorite. The dioritized rock contains poikiloblastic plagioclase (An29±2), diopside, and actinolitic hornblende. Relict quartz is abundant, but hematite is absent.
The compositions of plagioclase porphyroblasts in dioritized rocks are similar to those of rims of zoned plagioclase crystals in Romeral diorite. This suggests that dioritization took place late in the crystallization history of the diorite, when plagioclase rims were forming. The abundance of amphibole in the dioritized rock also suggests that dioritization may have occurred when the diorite was largely crystalline, and when late, interstitial residual diorite magma became water saturated. Hot aqueous fluids may have exsolved from such residual magma and moved outward from the pluton, carrying dissolved dioritic constituents to be deposited in quartzites and andesites near the diorite contacts.

Hematite-enriched and hematite-depleted quartzites: Iron contents of quartzite samples taken on a traverse perpendicular to the diorite contact suggest that hematite has been leached from dioritized quartzites adjacent to the diorite contact and redeposited in hematite-rich quartzites about 80 to 145 m from that contact (Fig. 56).
Quartzites less than about 80 m from the diorite contact are grayish to pinkish white, and they contain only about 2.5 to 4 weight percent of iron. By contrast, quartzites 80 to 145 m from the diorite contact are pink to dark red-brown and contain as much as 16.5 weight percent of iron, whereas hematite-banded quartzites more than about 145 m from the diorite contact contain about 5 to 6 weight percent of iron. The hot, aqueous fluids that caused dioritization of the quartzites probably also dissolved hematite from quartzites near the diorite contact and redeposited it at lower temperatures in quartzites 80 to 145 m from the contact.
Later, hematite was leached from quartzites along the East fault zone, during post-ore argillization and sericitization, which occurred during collapse of the Romeral hydrothermal system. Still later, hematite was leached from quartzites near the post-ore Punta de Piedra granodiorite east of El Romeral. This occurred during post-granodiorite hydrothermal alteration, which resulted in deposition of epidote, actinolite, brown biotite, fluorapatite, ilmenite, pyrite, and sphene in quartzites near the granodiorite contact.

Deuterically Actinolitized Diorite

The Romeral diorite contains about 26 ± 5 percent of actinolite, which is very evenly distributed through large volumes of rock. Some of the actinolite is interstitial to plagioclase, and some is in poikilitic intergrowths with plagioclase and augite. If hornblende ever was present, it all has been converted to actinolite. In addition, disseminated ilmenite grains generally are rimmed and interpenetrated by sphene. The ubiquitous distributions of both actinolite and sphene in the diorite, irrespective of distance from the Romeral orebodies, and independent of later actinolite veinlets, suggests that these minerals are the products of pervasive deuteritic alteration. The fact that post-diorite, pre-ore meladiorite dikes are not actinolitized suggests that this deuteritic alteration took place soon after consolidation of the diorite mass, and before emplacement of
the meladiorite dikes.

Much of the deuteric actinolite in Romeral diorite contains dust-sized particles of magnetite. Furthermore, inasmuch as actinolite contains 13.8 percent of Fe₂O₃ + FeO, whereas actinolitic hornblende contains 12.6 percent, augite contains 9.5 percent, and plagioclase contains none, it would appear that iron was added to the Romeral diorite during deuteric actinolitization.

Products of Ore-Related Hydrothermal Alteration

Rocks in and around the Romeral magnetite deposits underwent extreme mineralogical, textural, and chemical changes during ore deposition and related hydrothermal alteration (Fig. 53). Andesite porphyries and schists in the ore zones have been almost completely replaced by magnetite with subsidiary actinolite, and wallrocks surrounding the orebodies have been almost completely replaced by actinolite with subsidiary magnetite, plagioclase, clinzoisite, sphene, chlorapatite, chlorite, and scapolite. Such actinolitic assemblages are superimposed on oligoclase-diopside rock and La Liga andesite porphyry southeast of the Main orebody, and on hornfelses, phyllites, and schists north of the Main orebody and around the North orebody. West of both orebodies quartzo-feldspathic alteration has been extensively superimposed on actinolitized phyllites and actinolitized Romeral diorite. Northeast of the Main orebody, advanced argillic and sericitic alteration assemblages are
superimposed on actinolitized, chloritized and feldspathized hornfelses and phyllites. North of El Romeral, at La Escoba, actinolite in large veins is partly altered to talc. La Escoba andesites near the veins contain actinolite, scapolite, muscovite, and chlorite, and more generally the La Escoba andesites contain abundant microscopic actinolite and secondary green biotite.

**Amphibolitized metasedimentary rocks:** Dark green felty rocks near the Romeral fault, west of the Main orebody, are amphibolitized phyllites (Fig. 53). These rocks are in gradational contact with less altered phyllites, and their foliation is parallel to that of the phyllites. Microprobe analysis suggests that the amphibole in these rocks contains about 0.5 atoms of Al per formula weight (Appendix G). According to Deer, Howie, and Zussman (1962), the division between actinolite and hornblende-series amphiboles is arbitrarily taken at \( \text{Si} \leftrightarrow \text{Al} \) replacement of 0.5 Al atoms per formula weight; thus the amphibole of the felty green rock is transitional between actinolite and hornblende. According to experimental work by Liou (1972), transitional actinolite-hornblende suggests transitional greenschist-amphibolite-facies metamorphism at temperatures near 550°C.

Phyllites northeast of the Main orebody, schists surrounding the North orebody, and phyllites surrounding those schists also have been amphibolitized, as shown in Figure 53.
**Actinolite-plagioclase rocks:** Dark green rocks, containing abundant actinolite and secondary plagioclase along with marialitic scapolite, magnetite, and clinozoisite, occupy a 250-600 m wide zone east and southeast of the Main orebody (Table 7, and Tables A 53 and A 54, Appendix A). Many of these rocks are greenish-black and aphanitic, but secondary plagioclase locally is responsible for a porphyroblastic texture that resembles the primary texture of La Liga andesite porphyry. Relicts of oligoclase-diopside rock and of La Liga andesite porphyry are common, and both these relicts and the actinolitized rocks are cut by stockworks of veinlets containing actinolite, magnetite, clinozoisite, sphene, calcite, chlorapatite, chlorite, and tourmaline. Several steeply dipping pods of coarsely crystalline actinolite-magnetite rock also cut the actinolite-plagioclase rock, as shown in Figure 53.

Although compositions of actinolite-plagioclase rocks are erratic, comparison of mean chemical compositions of La Liga andesite porphyry and actinolite-plagioclase rock indicates that in general, conversion of one to the other involved major additions of ferrous and ferric iron, moderate addition of magnesium, and minor additions of water, phosphate, carbonate, chloride, sulfide and borate (Fig. 57). Major amounts of silica and aluminum were removed, along with moderate amounts of sodium and minor titanium and potassium. Calcium contents of individual samples vary widely, but the mean CaO content of actinolite-plagioclase rocks is about
Figure 57. Chemical changes involved in alteration of La Liga andesite porphyry to actinolite-plagioclase rock. Circled points on the left side of the graph represent mean compositions calculated from seven samples of La Liga andesite porphyry, and the error bars represent a range of one standard deviation from the mean. Circled points on the right side of the graph represent mean compositions of actinolite-plagioclase rock as calculated from five samples. The arrowed bars represent total ranges of compositions observed.
equal to that of La Liga andesite porphyry.

**TABLE 7**

Mineralogical and Chemical Compositions* of Actinolite-Plagioclase Rock

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Mean Vol.%</th>
<th>An</th>
<th>Chemical Constituent</th>
<th>Mean Wt.%</th>
</tr>
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<td>40.8</td>
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<td>SiO₂</td>
<td>44.6</td>
</tr>
<tr>
<td>plagioclase</td>
<td></td>
<td></td>
<td>TiO₂</td>
<td>0.5</td>
</tr>
<tr>
<td>primary</td>
<td>1.9</td>
<td>50</td>
<td>Al₂O₃</td>
<td>10.4</td>
</tr>
<tr>
<td>secondary</td>
<td>23.0</td>
<td>32</td>
<td>Fe₂O₃</td>
<td>8.0</td>
</tr>
<tr>
<td>scapolite</td>
<td>7.2</td>
<td></td>
<td>Fe</td>
<td>8.0</td>
</tr>
<tr>
<td>magnetite</td>
<td>6.1</td>
<td></td>
<td>Fe(sulfide)</td>
<td>1.9</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>6.0</td>
<td></td>
<td>MnO</td>
<td>tr</td>
</tr>
<tr>
<td>diopside</td>
<td>3.2</td>
<td></td>
<td>MgO</td>
<td>7.6</td>
</tr>
<tr>
<td>penninite</td>
<td>2.8</td>
<td></td>
<td>CaO</td>
<td>11.0</td>
</tr>
<tr>
<td>pyrite</td>
<td>2.5</td>
<td></td>
<td>Na₂O</td>
<td>2.5</td>
</tr>
<tr>
<td>tourmaline</td>
<td>1.4</td>
<td></td>
<td>K₂O</td>
<td>0.3</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>1.4</td>
<td></td>
<td>P₂O₅</td>
<td>0.5</td>
</tr>
<tr>
<td>calcite</td>
<td>1.1</td>
<td></td>
<td>Cl</td>
<td>0.3</td>
</tr>
<tr>
<td>sphene</td>
<td>1.1</td>
<td></td>
<td>F</td>
<td>tr</td>
</tr>
<tr>
<td>montmorillonite</td>
<td>0.5</td>
<td></td>
<td>CO₂</td>
<td>0.7</td>
</tr>
<tr>
<td>quartz</td>
<td>0.4</td>
<td></td>
<td>S</td>
<td>2.2</td>
</tr>
<tr>
<td>hematite</td>
<td>0.1</td>
<td></td>
<td>B₂O₃</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td>total</td>
<td>99.5</td>
<td></td>
<td>total 100.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>H₂O⁺</td>
<td>1.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>H₂O⁻</td>
<td>0.1</td>
</tr>
</tbody>
</table>

*Mean compositions are based on six samples, see Tables A 53 and A 54, Appendix A.

**Actinolite-sphene zone:** Along the southeast margins of the Main orebody is a 20-60 m wide zone of yellowish-green, crystalloblastic rock consisting of actinolite, olive-green sphene, diopside, calcite, chlorapatite, and magnetite (Fig. 53 and Table 8). This rock contains remnants of dark green, aphanitic actinolite-plagioclase rock.
Crystalloblastic actinolite and sphene apparently were superimposed on earlier actinolite-plagioclase rocks during middle to late stages of alteration related to deposition of the Main orebody.

### TABLE 8
Mineralogical and Chemical Compositions* of Actinolite-Sphene Rock

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Mean Vol.%</th>
<th>Chemical Constituent</th>
<th>Mean Wt.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>actinolite</td>
<td>49.6</td>
<td>SiO₂</td>
<td>39.2</td>
</tr>
<tr>
<td>sphene</td>
<td>15.1</td>
<td>TiO₂</td>
<td>8.7</td>
</tr>
<tr>
<td>diopside</td>
<td>14.3</td>
<td>Al₂O₃</td>
<td>1.5</td>
</tr>
<tr>
<td>calcite</td>
<td>12.2</td>
<td>Fe₂O₃</td>
<td>2.9</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>5.6</td>
<td>FeO</td>
<td>6.7</td>
</tr>
<tr>
<td>magnetite</td>
<td>2.0</td>
<td>MnO</td>
<td>tr</td>
</tr>
<tr>
<td>quartz</td>
<td>0.6</td>
<td>MgO</td>
<td>9.2</td>
</tr>
<tr>
<td>plagioclase</td>
<td>0.2 (An 31)</td>
<td>CaO</td>
<td>23.5</td>
</tr>
<tr>
<td>hematite</td>
<td>0.1</td>
<td>Na₂O</td>
<td>0.3</td>
</tr>
<tr>
<td></td>
<td>total</td>
<td>K₂O</td>
<td>tr</td>
</tr>
<tr>
<td></td>
<td></td>
<td>P₂O₅</td>
<td>2.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cl⁻</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>CO₂</td>
<td>4.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td>H₂O⁺</td>
<td>1.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>H₂O⁻</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>total</td>
<td>100.4</td>
</tr>
</tbody>
</table>

*Mean compositions are based on three samples (see Tables A 55 and A 56, Appendix A).

Clinozoisitic rocks: Clinozoisite is intergrown with magnetite and actinolite in the Main orebody, and clinozoisite veinlets cut the Main orebody. Intramineral (IM-2) dioritite dikes are partly replaced by clinozoisite on the eastern side of the Main orebody, and clinozoisite is scattered through the rocks of the actinolitic zones east of
that orebody. Farther to the east, clinozoisite, along with secondary plagioclase ($\text{An}_{17} \pm 5$), is interspersed through andesite peripheral to the actinolitic zones, and clinozoisite has partly replaced many plagioclase crystals in andesitic and dioritic rocks as far as 500 to 700 m east of the Main orebody. Clinozoisite also is present in rocks to the west of the Main orebody, in actinolitic veinlets that cut the Romeral diorite and the La Liga andesite, and in later magnetite-pyrite-biotite veinlets, which have aplitic alteration envelopes. Clinozoisite veinlets also cut actinolitized and feldspathized phyllites north of the North orebody. All of this suggests that while clinozoisite was forming along with magnetite and actinolite, it also was forming in somewhat cooler rocks peripheral to the zone of actinolitization. Then, as the Romeral thermal gradient began to decay, the clinozoisite front retreated, and late clinozoisite was superimposed on earlier clinozoisite-bearing, actinolitic and magnetitic rocks.

Products of Late Hydrothermal Alteration

**Biotitic rocks**: Very fine-grained, greenish-brown secondary biotite is pervasively disseminated through the volcanioclastic andesites of La Escoba. In those rocks, the biotite has partially replaced fine-grained plagioclase and actinolite grains. In the altered rocks near El Romeral, such greenish-brown secondary biotite is rare. Similar biotite is found, however, in the late, biotitic dioritite dikes
that cut the orebodies, and in late, magnetite-pyrite-biotite veinlets that have aplitic, quartzo-feldspathic alteration envelopes.

**Quartzo-feldspathic (aplitized) rocks:** West of the Romeral orebodies is a large, irregularly shaped zone in which dioritic rocks have been altered to aplitic rocks (Fig. 58). These rocks contain various proportions of oligoclase (An\textsubscript{10-20}), quartz, albite (An\textsubscript{8-10}), and microcline, along with minor tourmaline, and various amounts of remnant diorite, actinolite, zoned andesine, ilmenite, and magnetite. Diorite remnants, ranging in size from a few millimeters across to about 450 m across, are scattered throughout the aplitic zone. These diorite remnants have been deuterically actinolitized and cut by stockworks of actinolite veinlets. They also are cut by later magnetite-bearing veinlets which have aplitic, quartzo-feldspathic envelopes, or selvages. These aplitic selvages are in gradational contact with their dioritic wallrocks, and the selvage rocks are similar in composition and texture to the aplitic rocks of the large quartzo-feldspathic zone (Figs. 59, 61). Where the selvaged veinlets are closely spaced, only small remnants of diorite remain between them. Thus, the selvaged veinlets are regarded as forerunners of pervasive aplitization. Where the rocks have been completely aplitized, no remnants of diorite mineralogy or texture remain, and the rocks are white to light tan, fine- to
Figure 58. Distribution of aplitized and feldspathized rocks in the Romeral area.
medium-grained, equigranular, and in places, myrmekitic (Figs. 61, 63).

The selvaged veinlets of El Romeral generally are less than 1 mm thick, and they are sparsely and unevenly filled with low-Ti magnetite, pyrite, biotite, tourmaline, quartz, and clinozoisite. Their bleached, aplitic selvages have half-widths ranging from about 1 to 5 cm; and gradational boundaries between the selvage rocks and the green-mottled diorite wallrocks are 1 to 3 cm wide. Modal analyses of 5-mm slices of bleached rock and diorite, taken at successive distances from a veinlet, provide a notion of mineralogical changes across an aplitic selvage adjacent to a magnetite-pyrite-biotite veinlet (Table A 64, Appendix A). Chemical variations across the bleached envelope, as calculated from the modal analyses, are plotted in Figure 63. Actinolite and zoned andesine crystals have been progressively replaced along grain boundaries by quartz and oligoclase in the transitional zone between the aplitic selvage and its diorite wallrock. Although actinolite has been destroyed in the selvage rock, the overall iron content of this rock is about equal to that of the pre-existing actinolitized diorite. The iron is somewhat more erratically distributed, however, and it is present as magnetite rather than actinolite. Total iron is enriched at the outer edge of the bleached zone, where magnetite has been deposited in actinolite in the diorite. This pattern of chemical distribution suggests that iron moved outward from
Figure 59. Actinolitized diorite cut by veinlets with quartzo-feldspathic alteration envelopes (or aplitic selvages), near the western margin of the large zone of aplitic, quartzo-feldspathic rocks, west of the Romeral orebodies.

Figure 60. Aplitic, quartzo-feldspathic rock formed by pervasive hydrothermal alteration of Romeral diorite, west of the Main orebody.
Figure 61. Gradational boundary between aplastic quartzofeldspathic rock (left) and actinolitized Romeral diorite (right). This sample is from the margin of a quartzofeldspathic alteration envelope adjacent to a thin veinlet containing magnetite, biotite, pyrite, tourmaline, quartz, and clinozoisite. The veinlet is in actinolitized Romeral diorite along the western margin of the large, aplastic quartzofeldspathic zone west of the Romeral orebodies.

Figure 62. Microtexture of aplastic quartzofeldspathic rock from the central subzone of most intense quartzofeldspathic alteration. Myrmekitic intergrowths suggest rapid and simultaneous crystallization of quartz and oligoclase in the presence of an aqueous fluid. This rock also contains traces of black tourmaline.
Figure 63. Chemical variation across an aplitic selvage adjacent to a magnetite-biotite-pyrite veinlet in Romeral diorite.
the veinlet and was preferentially precipitated in actinolite at the margin of the bleached selvage. Furthermore, the pattern suggests that while silica, sodium, minor titanium, minor iron, and trace potassium were added to rocks adjacent to the veinlet, calcium, magnesium, and aluminum were removed from those rocks.

With the aid of feldspar staining techniques and modal analyses, the following subzones can be recognized within the large quartzo-feldspathic zone west of the Romeral orebodies (Fig. 58):

1. Oligoclase-albite-quartz (eastern subzone).
2. Quartz-oligoclase-albite (central subzone).
3. Oligoclase-quartz-microcline (western subzone).
4. Oligoclase-microcline (small northern subzone).
5. Oligoclase-quartz (small southern subzone).

Mean mineralogical compositions of rocks from three of the subzones are given in Table 9. More complete mineralogical and chemical data are given in Appendix A, Tables A 57 through A 62. Secondary plagioclase (An20-8) is predominant near the Main orebody, quartz is predominant in the central subzone, and microcline is present in the outermost subzones, along with oligoclase and/or quartz. Remnants of strongly actinolitized diorite containing low-Ti magnetite are common in aplitic rocks near the Main orebody. By contrast, remnants of deuterically actinolitized diorite, and relict disseminated ilmenite are common in the rocks of the outermost subzone.
Table 9
MINERALOGICAL COMPOSITIONS* OF QUARTZO-FELDSPATHIC ROCKS

<table>
<thead>
<tr>
<th>Minerals</th>
<th>Oligoclase-Albite Rock (Eastern Subzone)</th>
<th>Quartz-Albite Rock (Central Subzone)</th>
<th>Oligoclase-Albite Rock (Western Subzone)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean Vol.%</td>
<td>An</td>
<td>Mean Vol.%</td>
</tr>
<tr>
<td>primary plagioclase:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>core</td>
<td>7.2</td>
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<td>4.7</td>
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<td>inner rim</td>
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</tr>
<tr>
<td>microcline</td>
<td>0.2</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>quartz</td>
<td>7.8</td>
<td>40.7</td>
<td></td>
</tr>
<tr>
<td>penninite</td>
<td>0.2</td>
<td>1.1</td>
<td></td>
</tr>
<tr>
<td>biotite</td>
<td>0.3</td>
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<td></td>
</tr>
<tr>
<td>actinolite</td>
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</tr>
<tr>
<td>clinozoisite</td>
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<td>0.2</td>
<td></td>
</tr>
<tr>
<td>tourmaline</td>
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<td>0.4</td>
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<tr>
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</table>

* Modal analyses from which the mean compositions were calculated are given in Appendix A, Tables A 57, A 59, and A 61.
Figure 64. Chemical variation across the quartzo-feldspathic alteration zone.
Myrmekytic textures, such as those shown in Figure 62, are present in some of the rocks of the central subzone where replacement of diorite was essentially complete. Such textures probably resulted from rapid, simultaneous crystallization of quartz and feldspar at subsolidus temperatures, in the presence of an aqueous fluid. Similar textures have been produced experimentally in this manner (Jahns, personal communication, 1973).

Chemical variations across the quartzo-feldspathic zone are shown in Figure 64. Sodium is enriched in the inner subzone, near the orebodies, silica is enriched and aluminum is depleted in the central subzone, and potassium is enriched in the outermost subzone. Calcium, magnesium, titanium, and ferrous iron are depleted in the rocks of all subzones, but ferric iron is slightly and progressively enriched in successive outer, central, and inner subzones, so that total iron is only moderately depleted in the quartzo-feldspathic zone as a whole.

**Feldspathized rocks**: White, aphanitic oligoclase (An_{12-22}) surrounds subparallel remnants of dark green, actinolitized phyllite in feldspathized rocks north of the North orebody (Figs. 58 and 65). Such rocks also are present northwest of the North orebody, where they contain microcline as well as oligoclase, and west of the Main orebody, between the Romeral fault and the diorite contact. In addition, remnants of such rocks are present in the argillic materials north-northeast
of the Main orebody, and slivers of such rock are common along the eastern, western, and northern margins of the cataclasite zone northwest of the North orebody (Fig. 53).

Figure 65. White, oligoclase-rich rock with remnants of dark green, actinolitized phyllite, from north of the North orebody. The oligoclase has been cut by a small actinolite-magnetite-apatite vein.

Several lines of evidence suggest that feldspathic and quartzo-feldspathic alteration lagged behind ore deposition and pervasive actinolitization at El Romeral:

1. Feldspathized rocks contain actinolitized phyllite remnants, and aplitized rocks contain relict actinolitized diorite. Near the orebodies, these actinolitized remnants commonly contain low-Ti magnetite, similar to the Romeral ore magnetite. This suggests that feldspathic and quartzo-feldspathic alteration followed pervasive actinolitization and ore deposition at El Romeral.
2. The minerals found in veinlets with quartzo-feldspathic alteration envelopes are typically late minerals in the Romeral paragenetic sequence. Pyrite and clinozoisite are found in late veinlets that cut the Romeral orebodies and in late alteration assemblages superimposed on intramineral dikes. Tourmaline is found with late chlorite, superimposed on actinolitized rocks southeast of the Main orebody, and in quartz-albite pegmatite dikes that cut the orebodies. Greenish-brown biotite, similar to the veinlet biotite, is superimposed on actinolitized andesites near La Escoba and is found in late dikes that cut the orebodies. Quartz is found in late actinolite-magnetite-apatite veins and in late quartz-albite pegmatite dikes. These observations suggest that the magnetite-pyrite-biotite-tourmaline-quartz-clinozoisite veinlets and their quartzo-feldspathic selvages formed late in the Romeral sequence.

3. The secondary plagioclase found in the feldspathic and quartzo-feldspathic zones is more sodic (An20-8) than that of the ore zones (An27±5). This suggests lower temperatures of formation, and inasmuch as temperature probably decayed with time, it also suggests that feldspathic and quartzo-feldspathic alteration followed ore deposition.

4. Aplitic, tourmaline-bearing quartz-oligoclase-albite rocks of the central subzone contain the same minerals as pegmatite dikes that cut the orebodies. This suggests that the aplitic rocks may have formed at about the same time as the post-ore pegmatite dikes.

However, the fact that quartzo-feldspathic rocks are cut by actinolite-magnetite-apatite veins and by late diorite dikes, suggests that quartzo-feldspathic alteration occurred before dioritic igneous activity and hydrothermal actinolite-magnetite-apatite deposition had completely ceased in the Romeral area.
Post-ore Hydrothermal Alteration

Chloritic rocks: Actinolitic rocks peripheral to the Romeral orebodies are partially chloritized. Within the Main orebody, actinolite in intramineral dikes is partially chloritized, and chlorite-bearing magnetite-oligoclase-chalcopyrite vein-lets cut both the magnetite ore and the intramineral (IM-4) dikes. North of the Main orebody and east of the North orebody, spheroidal chlorite-nontronite knots are common in actinolitized phyllites. Chlorite is typical of relatively low-temperature, greenschist-facies metamorphic assemblages, whereas actinolite is typical of somewhat higher temperature, transitional greenschist-amphibolite facies metamorphic assemblages. Chlorite apparently was superimposed on magnesium-rich actinolitic rocks as the Romeral thermal gradient began to decay, soon after Romeral ore deposition and actinolitization.

Montmorillonitic rocks: In the east mass of Romeral diorite, 700 to 900 m east of the Main orebody, the cores and inner rims of most zoned andesine crystals are partially altered to montmorillonite. Minor montmorillonite also is present in argillized phyllites northeast of the Main orebody. The montmorillonite in both of these areas may have formed during incipient argillization that followed Romeral ore deposition and chloritization but preceded extreme kaolinitization and sericitization.
Advanced argillic and sericitic rocks: A zone of soft, light-colored, clay-rich materials lies northeast of the Main orebody and along the southern and eastern sides of the North orebody (Fig. 53). The clay-rich materials northeast of the Main orebody contain ghostlike remnants of actinolitized, feldspathized, and chloritized phyllites. Dickite is the predominant clay mineral near the North Extension orebody and at the southern end of the North orebody. Northeastward the dickite gives way to kaolinite and pyrophyllite, mixed with relict actinolite, oligoclase, chlorite, and montmorillonite. Most magnetite within the dickite and kaolinite-pyrophyllite zones has been oxidized to hematite. Within the East fault zone, the fault gouge has been altered to a snow white mixture of kaolinite and sericite. Near the East fault zone, phyllites contain kaolinite and sericite, and just east of that fault zone, small outcrops of biotite schist and hematitic quartzite also contain sericite.

This northeastward progression from dickite, to kaolinite and pyrophyllite, to sericite might reflect a northeastward increase in the potassium contents of the pre-argillic rocks. In the dickite zone, biotitic phyllites were altered to actinolite-chlorite-magnetite rocks before they were argillized. Thus, potassium was removed before argillization took place. In the sericite zone, however, relict potassium-bearing biotite probably still was present in the phyllites and schists when hydrogen-ion metasomatism began. Therefore, it appears possible that the potassium content of the sericite
zone may have been inherited from pre-existing biotite phyllites and schists. Minor mobilization of such potassium might account for the sericite in the East fault gouge zone and in the hematitic quartzites. Alternatively, the eastward progression from dickite, to kaolinite and pyrophyllite, to sericite, could be attributed to eastward temperature increase, possibly resulting from intrusion of the post-ore Punta de Piedra granodiorite batholith east of El Romeral.
ROMERAL FAULT SYSTEM

The Romeral fault system is an interbranching network of steeply dipping faults with north-northwest, north, and northeast strikes (Fig. 66). The predominant sense of movement on the early faults was left-lateral, whereas it was right-lateral on the post-ore faults. Some of the early faults have been displaced by the late faults. Therefore, the early faults can be understood best if the later movements are first reconstructed. For this reason, the faults of El Romeral are herein discussed in reverse chronological order.

Late North-northeast fault: The youngest major fault in the Romeral system is the post-ore North-northeast fault, which lies west of the Main orebody, cuts the north end of the North-extension orebody, and lies southeast of the North orebody. Its zone of breccia and gouge ranges from 10 to 50 m in thickness and contains fragments of magnetite and of white argillized rock in a dark gray matrix of pulverized magnetite and hematite (Fig. 66).

Slickensides measured southeast of the Main orebody plunge 20°S, and several offset features indicate about 200 ± 50 m of post-ore right-lateral movement. The early Romeral fault is offset about 225 ± 25 m where it is cut by the North-northeast fault west of the Main orebody. The North orebody is separated about 200 ± 25 m with respect to the North-extension orebody, and magnetic anomalies suggest
Figure 66. The Romeral fault system, showing the late northeast fault, the late faults in the Main ore-body, the East fault, and the early Romeral fault.
that several slivers of magnetite are scattered along the fault zone between the two orebodies. Schists southeast of the North orebody are displaced about 150 m along the North-northeast fault. The lesser displacement is attributable to diversion of movement onto the pre-existing East fault (Fig. 66).

**Late faults in the Main orebody:** Several northeast- and northwest-trending right-lateral faults that cut the Main orebody (Fig. 66) are expressed by breccia zones ranging from a few cm to 5 m in thickness. The breccias contain fragments of magnetite, and in some of the fault zones these fragments are surrounded by rinds of radially oriented prisms of actinolite (Fig. 57). The spaces between adjacent actinolite rinds are partially filled with apatite, and open spaces are lined with calcite. In other late fault zones the magnetite fragments are surrounded by pulverized magnetite (Fig. 58). Both apatite and fine-grained magnetite are undesirable for steel making, and both are concentrated in late fault zones. Continued mapping and projection of these fault zones would therefore help in prediction of ore-quality distribution.

As shown in Figure 66, several intramineral dikes are offset 10-50 m in a right-lateral sense along northeast- and northwest-trending faults. This suggests that the faults are related to the late, right-lateral North-northeast fault.
Figure 67. Fault breccia with fragments of magnetite cemented by actinolite, chlorapatite, and calcite. This specimen is from a north-northeast striking, steeply dipping right-lateral fault that cuts the ore and the intramineral dikes on the southeast side of the Main orebody.

Figure 68. Fault breccia with fragments of magnetite and actinolite in a matrix of pulverized magnetite. This specimen is from a branch of the late NNE fault in the southwestern part of the Main pit.
East fault: The East fault is a post-ore break that lies east of both orebodies, strikes north-northwest, and dips 65 to 70° SW. Its gouge zone is 50 to 60 m thick (Fig. 44), and consists mostly of fine-grained quartz, muscovite, and kaolinite. Early movement on the East fault was dip-slip and reverse, whereas later movement on the northern part of this fault was right-lateral. Slickensides along the south part of the fault plunge 55° NW, indicating nearly dip-slip movement, and a structural section across the North pit suggests that schists west of the fault have been moved upward about 150 m with respect to phyllites east of the fault. North of its intersection with the North-northeast fault, however, it has gently plunging slickensides and mullions indicating that its latest movement was strike-slip. Part of the right-lateral movement of the North-northeast fault evidently was shunted onto the north part of the East fault.

The east wall of the North pit is parallel to the East fault zone, which consists of soft, white, kaolinized and sericitized gouge (Fig. 69).

Early Romeral fault: The early Romeral fault strikes north-northwest and lies west of the Romeral orebodies (Fig. 66). Its southern part dips 75°E to vertical, and its northern part dips 70°W to vertical. Its breccia zone is 10 to 40 m thick, and west of the Main orebody it contains fragments of magnetite, actinolite, and apatite in an apatite-rich matrix containing magnetite and actinolite (Fig. 70).
Figure 69. East fault along the east wall of the North pit. Landslides from the soft, white, kaolinized and sericitized gouge zone have prevented deepening of the North pit to recover remaining pods of magnetite.
Many of the magnetite-actinolite fragments are deformed into streamlined shapes because of renewed (post-breccia) movement along the fault (Fig. 71).
Figure 70. Fault breccia from the Romeral fault zone southwest of the Main orebody.

Matrix of breccia consists of:
- apatite (>10%)
- actinolite (10%)
- magnetite (10%)

Long axes of mineral grains in matrix are aligned as shown.

Figure 71. Cataclasite from the Romeral fault zone northwest of the Main orebody.
Offset features suggest about 700 m of left-lateral movement on the early Romeral fault. The andesite-phyllite contact near the north end of the Main orebody is separated about 500 m from its probable counterpart west of that fault. Restoration of 200 m of right-lateral movement on the late North-northeast fault suggests 700 m of left-lateral movement on the early Romeral fault opposite the Main orebody. Similarly, a sliver of Romeral diorite west of the Main orebody probably was displaced about 700 m northward along the early Romeral fault before it was displaced about 200 m southward along the late North-northeast fault (Fig. 72).

The early Romeral fault branches northwest of the North orebody, one branch extending along the diorite contact and the other branch extending through phyllites east of the diorite contact. A sliver of Romeral diorite, northwest of the North orebody, is displaced about 350 m north along the northeastern branch of the early Romeral fault (Figs. 66 and 72c). Presumably the remainder of the 700 m displacement was taken up on the western branch of the fault and in the cataclasite zone between the two branches of the fault. This cataclasite zone is a wedge of relatively soft phyllite, schist, and apatite-magnetite-actinolite rock caught between the two branches of the Romeral fault. Reconstruction of fault movements suggests that this wedge, which is at a bend in the fault, was severely sheared, and was also severely compressed against a massive block of relatively unyielding diorite.
a. Early left-lateral movement along the Romeral fault and subsidiary northeast-striking, steeply dipping fractures, moving the andesite contact northward relative to the bulge in the diorite contact. This faulting could have resulted in a fluid pressure drop and provided passageways for hydrothermal fluids.

b. Ore deposition along north-south and northeast-striking, steeply dipping fracture zones as left-lateral movement continued.

Figure 72. Reconstructed movements along the Romeral fault system.
c. Reverse movement on the East fault, and continued left-lateral movement on the Romeral fault, resulting in:

1. movement of diorite slivers 350-700 m northward,

2. formation of en-echelon gash fractures which were filled by IM dikes and/or mineralized,

3. formation of cataclasites between the east and west branches of the fault northwest of the North orebody.

d. Post-ore right-lateral movements along the North-northeast fault, resulting in the present structural configuration.

1. At the northeast end of the North-northeast fault, right-lateral movement was shunted onto the pre-existing East fault.

2. Right-lateral movements on minor late faults in the Main orebody offset intramineral dikes.

Figure 72. (continued)
In the Main orebody, tensional gash fractures related to left-lateral movement on the Romeral fault are filled by intramineral dikes. Many of these dikes are lenticular, have lenticular apophyses, and contain unrotated, lenticular inclusions, all of which suggests that the dikes filled tensional fractures. Furthermore, the dikes are oriented in a northwest-trending, en echelon pattern, which is to be expected for gash fractures related to north-south, left-lateral movement. Similarly, several northwest-trending zones of high-grade ore in the eastern part of the Main orebody may have resulted from hydrothermal replacement along tensional fractures related to left-lateral movement on the Romeral fault.

Regional fault pattern: The Romeral fault system is one of a discontinuous series of breaks delineating the southern 300 km of the Atacama strike-slip fault system, which parallels the Chilean coast for about 1000 km, north of El Romeral. Several other major magnetite deposits, such as El Tofo, El Algorrobo, and Cerro Imán are located near north-trending faults of the Atacama system. Early fault movements that accompanied ore deposition and intrusion of intramineral dikes at El Romeral were probably left-lateral. At Cerro Imán, another magnetite deposit of Neocomian age, 300 km north of El Romeral, northwest-trending dikes similar to the intramineral and late dikes of El Romeral cut the orebody. If these dikes, like those of El Romeral, filled
tensional fractures related to strike-slip faulting, their orientation would suggest left-lateral movement. Thus, the predominant sense of movement on the Atacama fault system may have been left-lateral during Neocomian time. Post-ore movements however have been predominantly right-lateral. There has been about 200 ± 50 m of post-ore right-lateral movement on the late North-northeast fault at El Romeral, and in northern Chile the Atacama fault is active, with right-lateral stream offsets of as much as 1.5 km (St. Amand and Allen, 1965).

A possible explanation for this reversal in the sense of movement on the Atacama fault system comes from plate tectonic observations by Larson and Pitman (1972). Their worldwide correlation of sea-floor magnetic anomalies of Mesozoic age suggests that during the time span of 130 to 60 m.y. ago, the oceanic Phoenix plate moved southeastward as the South American plate moved westward. With this configuration, oblique subduction at the margin of the South American plate would have had a strong left-lateral component, which might have caused left-lateral movement along the Atacama fault system. Later, when the Chile rise became active in about Eocene time (Herron, 1972, p. 1676), the oceanic Southeast Pacific plate began to move northeastward toward the westward-moving South American plate. Oblique subduction resulting from this configuration would have a right-lateral component that might result in right-lateral movement along the Atacama fault system. One problem with
this explanation, however, is that magnetite fragments in breccias of right-lateral faults that cut the Main orebody are cemented by actinolite and apatite. This suggests that right-lateral fault movement closely followed ore deposition and therefore began well before Eocene time.
The Romeral magnetite deposits were formed at a depth of approximately 7 ± 3.5 km. The deposits probably are late Neocomian in age, and their tops lay at about the level of the Paleozoic-Jurassic unconformity. They probably were formed under about 5000 ± 3500 m of volcaniclastic andesites of Jurassic age (p. 72) and about 1250 m of Arqueros andesite sills and flows of Neocomian age (p. 74). The deposits therefore formed at a depth of about 6750 ± 3500 m, or rounding upward, 7 ± 3.5 km.

Pressure

Lithostatic pressure, $P_s$, at 7 ± 3.5 km depth would be 2 ± 1 kb, and hydrostatic pressure, $P_f$, would be 0.7 ± 0.35 kb. In tight microfractures fluid pressures probably approached lithostatic pressure, whereas in fracture zones, fluid pressures were probably intermediate between lithostatic and hydrostatic, and in actinolite-magnetite-apatite veins with textures suggesting open-space filling, fluid pressures may have approached hydrostatic pressure. Thus, rock pressure was approximately 2 kb, fluid pressures ranged from about 0.7 kb to 2 kb, and median fluid pressure was about 1.4 ± 0.7 kb during formation of the Romeral magnetite deposits.
Magmatic Differentiation Trends

The successively younger La Liga andesite and Romeral diorite plutons are progressively depleted in iron with respect to Arqueros andesites. Some post-diorite, pre-ore dikes are iron-enriched, whereas others are iron-depleted. Intraminal diorite dikes are enriched in iron, and late dikes are clearly divisible into iron-rich and iron-poor sets. In order to illustrate these differences, to show their development with time, and to indicate the other chemical changes that offset changes in iron content, the graphs of Figure 73 were constructed. In these graphs, the mean chemical composition of each igneous rock type is plotted against its approximate order of emplacement.

Although most of the igneous rocks of El Romeral are more or less deuterically and/or hydrothermally altered, the least altered samples available were used in this study of magmatic differentiation trends, and veinlets were avoided in the point counts from which the compositions were calculated. In each sample used, textural and mineralogical features suggested that the igneous components of the rock composition were predominant over superimposed deuteric or hydrothermal components.

As shown in Figure 73, sympathetic changes in Fe, MgO, and CaO generally are balanced by complementary changes in SiO₂, Al₂O₃, and Na₂O. High values in the patterns of Fe and MgO may reflect new arrivals of relatively mafic magmas,
whereas gradually decreasing trends may represent gradual depletion of mafic constituents. Gradual decrease in CaO and complementary increase in Na$_2$O probably reflect progressive crystallization of calcic plagioclase. Extreme compositional differences between iron-rich and iron-poor late dikes suggest simultaneous and extreme differentiation along two divergent compositional trends.

Harker diagrams provide second view of compositional variations within igneous rock suites (Fig. 74), that of variation in each constituent versus variation in silica. Such variation diagrams have been used to characterize rock suites in terms of their alkalinity (the Peacock index), to suggest whether rock suites are consanguineous (Turner and Verhoogen, 1960), and to indicate magmatic differentiation trends (Krauskopf, 1967).

Intersection of the Romeral curves for CaO and Na$_2$O + K$_2$O at about 58 percent of silica indicates that the Romeral igneous suite is calc-alkaline by the Peacock classification (Barth, 1962). The degree of compositional scatter shown on Harker diagrams suggests either extreme differentiation, or non-consanguineous origin for the rocks, or strong post-magmatic alteration of the rocks. Mineralogical similarities and close temporal and spatial relationships between the bodies of igneous rock that comprise the suite suggest that the rocks are consanguineous. Although some of the compositional scatter may have resulted from differential post-magmatic alteration, textural and mineralogical features of
each sample used in this study suggested that the igneous components of its composition were predominant over superimposed deuteric or hydrothermal components. Most of the compositional scatter shown on the Harker diagrams therefore is thought to have resulted from igneous differentiation.

To interpret Harker variation diagrams in terms of magmatic differentiation trends, "One must keep in mind that increase in silica from left to right necessarily means decrease in other oxides, regardless of differentiation" (Krauskopf, 1967, p. 398). The slope of the "line of no differentiation" shown on each of the diagrams in Figure 74 indicates the amount of change in weight percent of each constituents that would be necessary to offset a progressive change in silica content. Inasmuch as Arqueros andesites represent relatively undifferentiated andesites, the mean composition of Arqueros andesite is taken as an initial composition, and the "line of no differentiation" is plotted to pass through the point representing that composition. Differentiation of a constituent is significant only if it causes a markedly larger or smaller change than that indicated by the "line of no differentiation" for that constituent.

The following differentiation trends are apparent on the Harker variation diagrams for the Romeral igneous suite:

1. Ferric, ferrous, and total iron decrease significantly with increasing silica (to the right of the initial point). Conversely, iron increases significantly with decreasing silica (to the left of the initial point).
Figure 74. Harker variation diagrams illustrating differentiation trends in the Romeral igneous rock suite. The points indicate mean rock compositions, and the error bars indicate ranges of plus-or-minus one standard deviation. The numbers indicate approximate order of emplacement. The short-dashed lines are "lines of no differentiation," and the heavy lines show generalized differentiation trends.
2. Lime and magnesia generally decrease with increasing silica.

3. Soda and alumina concentrations are somewhat scattered, but they generally increase with increasing silica, or decrease less steeply than the "line of no differentiation."

4. Titania and potash concentrations generally are scattered near their respective "lines of no differentiation."

The compositional contrasts between iron-enriched and iron-depleted rocks are increasingly extreme in progressively younger rocks. (Compare rocks 3 and 4 to 7, 8 and 9, and to 19 and 20 on the iron vs silica diagrams.) Thus magmas of increasingly contrasting compositions irrupted in the Romeral area during the time interval between crystallization of the pre-ore Romeral diorite pluton and emplacement of the post-ore Punta de Piedra granodiorite batholith. This suggests that after crystallization of the upper, presently exposed part of the Romeral diorite pluton, the Romeral area was underlain by a body of residual magma, which became progressively differentiated with respect to iron and silica. The iron-poor dikes presumably filled fractures that tapped progressively iron-depleted upper (?) portions of such a differentiating body of magma, whereas the iron-rich dikes presumably tapped its progressively iron-enriched lower (?) portions (Fig. 81).

An Alk-Fe-Mag diagram (Figure 75) provides a third view of differentiation trends and compositional groupings within the Romeral igneous rock suite. Compositions of the oldest rocks, those of the Arqueros-La Liga andesite group, cluster
near the center of the magnesium-iron-alkali field. These rocks probably are relatively undifferentiated andesites. Romeral diorite is slightly depleted in iron and magnesium and enriched in alkalies with respect to these earlier andesites. Most post-diorite, pre-ore dikes also are depleted in iron and magnesium and enriched in alkalies. However, the post-diorite, pre-ore porphyritic meladiorite dikes and the intramineral dioritite dikes are enriched in iron and depleted in alkalies. The post-ore dikes exhibit two divergent compositional trends, one toward iron and alkali enrichment, and the other toward iron depletion and alkali enrichment.

![Figure 75. Alk-Fe-Mg diagram for the igneous rocks of El Romeral. Alk = wt percent of Na2O + K2O, Fe = wt percent of FeO (Fe2+ + Fe3+ calculated as FeO), Mg = wt percent of MgO, and Alk + Fe + Mg = 100. The numbers indicate the approximate sequence of emplacement.](image-url)
Table 10, which shows the variation in compositions of plagioclases in successively younger Romeral igneous rocks, provides a fourth view of Romeral magmatic differentiation trends.

### Table 10

**IGNEOUS PLAGIOCLASE COMPOSITIONS**

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In general, successively younger Romeral igneous rocks contain increasingly sodic plagioclases. This would seem to suggest progressive Ca depletion and residual Na enrichment in a series of residual magmas. There are exceptions to the general trend, however. Porphyritic meladiorite, for example, contains somewhat more calcium than Romeral diorite,
and its plagioclases are considerably more anorthitic than those of Romeral diorite. By contrast, the intramineral dioritites contain almost as much Ca as Romeral diorite, but their plagioclases are considerably more sodic than those of the diorite.

Crystallization Temperatures and Water Contents of Romeral Magmas

Romeral diorite: Textural interpretation (Table A-21) suggests the following crystallization sequence for Romeral diorite:

1. early magmatic plagioclase cores (An 47 ± 4);
2. magmatic plagioclase inner rims (An 38 ± 4), unzoned plagioclase crystals (An36 ± 6), augite crystals, and ilmenite-magnetite grains;
3. latest magmatic plagioclase rims (An 24 ± 2), and actinolitic hornblende crystals (only preserved in meladiorite);
4. deuteric actinolite and sphene;
5. hydrothermal actinolite, magnetite, sphene, clinozoisite, chlorapatite, chlorite, biotite, tourmaline, pyrite, and chalcopyrite.

Ilmenite-magnetite grains in Romeral diorite are partly altered to sphene and cannot be used for T-fo2 determination. However, unaltered ilmenite-magnetite grains in Romeral meladiorite provide the basis for estimates of temperature and oxygen fugacity during equilibration of ilmenite with magnetite, in meladiorite. Inasmuch as the ilmenite-magnetite intergrowths of Romeral diorite are similar in appearance to
those of meladiorite, the risky assumption is made that they formed under similar conditions. The compositions of coexisting magnetite and ilmenite from meladiorite, estimated from their unit-cell parameters as measured by x-ray diffraction, are magnetite $\text{Mg}_63\text{Fe}_{37}$, ulvospinel $\text{Mg}_{37}$, and ilmenite $\text{Fe}_{80}$-hematite $\text{Fe}_{20}$ (Appendix F). In addition, the ilmenite crystals also contain about 5 percent of hematite microlamellae along (0001). According to the curves of Buddington and Lindsley (1965), which give compositions of coexisting magnetite and ilmenite as functions of temperature and oxygen fugacity, these compositions suggest magnetite-ilmenite equilibration at temperatures near $1075 \pm 50^\circ\text{C}$ and $f_{\text{O}_2} = 10^{-8} \times 1\text{atm}$ (about one log unit above the $f_{\text{O}_2}$ of the NiNiO$_2$ buffer at $1075^\circ\text{C}$).

Inasmuch as solidus temperatures depend greatly on water contents of residual magmas, it is necessary to estimate water contents of late residual magmas in order to estimate solidus temperatures. The presence of dioritized quartzites and andesites near contacts of Romeral diorite, together with the ubiquity of deuteric actinolite and sphene in the diorite, suggest that when the Romeral diorite had almost completely crystallized, its interstitial residual magma reached water saturation. According to Burnham (1967), an andesitic magma at about 7 km depth would approach water saturation at about 6 weight percent of water. Thus it is inferred that the latest residual diorite magma contained about 6 weight percent of water.

Given the inferred crystallization history of Romeral
diorite, given an approximate temperature for the formation of ilmenite-magnetite intergrowths in the diorite, and given the inference that the latest residual diorite magma contained about 6 weight percent of water, then experimental andesite melting curves by Eggler and Burnham (1973) can be used to derive an approximate crystallization-temperature curve for Romeral diorite, as shown in Figure 76. Temperatures for the main events depicted on this curve are given in Table 11.

Table 11

ROMERAL DIORITE CRYSTALLIZATION TEMPERATURES

<table>
<thead>
<tr>
<th>Events</th>
<th>Approximate Temperatures</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Initial magma intrudes to 7 ± 3.5 km depth.</td>
<td>&gt;1175°C</td>
</tr>
<tr>
<td>2. Initial plagioclase crystallizes.</td>
<td>1175°C</td>
</tr>
<tr>
<td>3. Initial augite and ilmenite crystallize.</td>
<td>1075 ± 50°C</td>
</tr>
<tr>
<td>4. Initial magnetite crystallizes.</td>
<td>1035°C</td>
</tr>
<tr>
<td>5. Initial amphibole crystallizes.</td>
<td>900°C</td>
</tr>
<tr>
<td>6. Rock is solidified.</td>
<td>780°C</td>
</tr>
</tbody>
</table>
Figure 76. T-X section at 5kb for Mt. Hood andesite containing various amounts of water and CO₂ (after Eggler and Burnham, 1973, Fig. 4, p. 2524). Other diagrams by Eggler and Burnham (1973) suggest that "out temperatures" are relatively insensitive to changes in total pressure between 2kb and 5kb. Thus, even though total pressure at El Romeral probably was 2 ± 1 kb, and even though the andesitic and dioritic rocks of El Romeral are not exact compositional equivalents of Mt. Hood andesite, the curves shown above probably are approximately applicable to the Romeral magmas and rocks.
Extrapolation of the Romeral diorite crystallization curve to temperatures above the initial ilmenite-augite crystallization temperature (1075 ± 50°C), suggests that plagioclase began to form at about 1175° ± 50°C. This further suggests that the diorite magma initially contained about 1.5 ± 1.5 weight percent of water. As anhydrous minerals, such as plagioclase, augite, and ilmenite formed, the residual magmas probably became progressively enriched in water.

Intramineral dioritites: Relatively unaltered intramineral diorite dikes southeast of the Main orebody contain small plagioclase and actinolitic hornblende phenocrysts in a saccharoidal groundmass of plagioclase, actinolitic hornblende, and titanomagnetite. These minerals are thought to be primary (p. 87, 88), and thus their textural interrelationships suggest slow crystallization of the small phenocrysts, followed by rapid, almost simultaneous crystallization of the saccharoidal groundmass minerals at lower temperatures. The abundance of primary amphibole (more than 15

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1 A bothersome inconsistency arises here. The high crystallization temperature (determined for ilmenite from the porphyritic meladiorite by the method of Buddington and Lindsley (1964), suggests a relatively dry magma according to the crystallization-temperature curves given by Eggler and Burnham (1973) for hydrous andesites. The porphyritic meladiorite, however, contains about 27 volume percent of actinolitic hornblende, which, if it is primary, suggests a high magmatic water content according to Holland (1965).
weight percent) suggests that intramineral dioritites crystallized from residual diorite liquids that contained more than 7 weight percent of water (Holland, 1972). According to Burnham (1967), however, an andesitic magma containing 7 weight percent of water would become water saturated at about 9 km depth, about 2 km below where the intramineral dikes intruded the growing Romeral magnetite deposits. Rapid upward movement of such water-saturated magma (already containing plagioclase and amphibole phenocrysts) into tensional fractures (p. 85), would have necessitated rapid exsolution of about 1 percent of water, which would have increased mineral-nucleation rates. This effect, combined with upward temperature decrease, would have caused rapid, simultaneous nucleation and growth of plagioclase, amphibole, and titano-magnetite crystals to form the observed saccharoidal groundmass textures. An approximate crystallization-temperature curve, based on the crystallization sequence and the magmatic water contents postulated above, and on experimental melting curves for hydrous andesites (Eggler and Burnham, 1973) is shown in Figure 76. Crystallization temperatures from this curve are given in Table 12.
Table 12
INTRAMINERAL DIORITITE CRYSTALLIZATION TEMPERATURES

<table>
<thead>
<tr>
<th>Events</th>
<th>Approximate Temperatures</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Plagioclase phenocrysts begin to form in water-rich residual diorite magma at 9 km depth.</td>
<td>about 1000°C</td>
</tr>
<tr>
<td>2. Actinolitic hornblende phenocrysts begin to form at 9 km depth.</td>
<td>about 930°C</td>
</tr>
<tr>
<td>3. Magma moves upward rapidly into tensional fractures, losing about 1 percent of water as it rises to 7 km depth.</td>
<td>rapid temperature decrease</td>
</tr>
<tr>
<td>4. Plagioclase, actinolitic hornblende, and titanomagnetite nucleate simultaneously and grow rapidly to form the saccharoidal groundmass.</td>
<td>885-760°C</td>
</tr>
<tr>
<td>5. Rock is completely solidified.</td>
<td>780°C</td>
</tr>
<tr>
<td>6. Titanomagnetite is partially oxidized.</td>
<td>600°C</td>
</tr>
<tr>
<td>7. Hydrothermal actinolite, magnetite, scapolite, and chlorapatite replace actinolitic hornblende and plagioclase.</td>
<td>550-475°C (p. 212)</td>
</tr>
<tr>
<td>8. Hydrothermal clinozoisite, chlorite, pyrite, chalcopyrite, and sphene replace earlier magmatic and hydrothermal minerals.</td>
<td>&lt;500°C (p. 213)</td>
</tr>
</tbody>
</table>

Inasmuch as intramineral magmas probably contained about 6 weight percent of water when they arrived at 7 km depth, whereas intramineral rocks generally contain only about 2.5 percent of water, about 3.5 percent of water probably was lost during crystallization. This hot water, perhaps mixed with groundwaters heated by dike emplacement,
may have partly dissolved and oxidized wallrock ore-magnetite grains, thereby causing them to acquire unusually delicate skeletal shapes and intergrowths of hematite and maghemite.

Magmatic Densities

If gravitational sorting of mineral grains is to be considered as a possible mechanism for magmatic differentiation, then the densities of the magmas and minerals involved should be estimated. Mineral densities are given in mineralogical textbooks, but magmatic densities are more difficult to determine in that they vary with composition, temperature, and pressure. According to Bottinga and Weill (1970), "Densities of silicate melts can be calculated from their compositions with an accuracy comparable to that of actual measurements in complex geological systems" (p. 181). "The density, $\rho$, of a multicomponent phase may be expressed as

$$\rho = \frac{\Sigma_{i} X_{i} M_{i}}{\Sigma_{i} X_{i} \bar{V}_{i}},$$

where $X_{i}$ is the mole fraction of component $i$, $V_{i}$ its partial molar volume, and $M_{i}$ its gram formula weight" (p. 169). Using partial molar volumes and temperature-pressure correction factors given by Bottinga and Weill (1970), the following magmatic densities were calculated for Romeral magmas at the inferred conditions specified in Table 13.
Table 13
MAGMATIC DENSITIES

<table>
<thead>
<tr>
<th>Magma</th>
<th>Inferred water content (weight percent)</th>
<th>$T$ (°C)</th>
<th>$P$ (kb)</th>
<th>Density</th>
</tr>
</thead>
<tbody>
<tr>
<td>La Liga andesite</td>
<td>1.5</td>
<td>1200</td>
<td>2</td>
<td>2.44</td>
</tr>
<tr>
<td>Romeral diorite</td>
<td>1.5</td>
<td>1200</td>
<td>2</td>
<td>2.43</td>
</tr>
<tr>
<td>dacite porphyry</td>
<td>5 (?)</td>
<td>1200</td>
<td>2</td>
<td>2.21</td>
</tr>
<tr>
<td>porphyritic meladiorite</td>
<td>5 (?)</td>
<td>1200</td>
<td>2</td>
<td>2.37</td>
</tr>
<tr>
<td>intramineral dioritite</td>
<td>7</td>
<td>1000</td>
<td>3</td>
<td>2.24</td>
</tr>
<tr>
<td>late biotite dioritite</td>
<td>7</td>
<td>1000</td>
<td>3</td>
<td>2.25</td>
</tr>
<tr>
<td>late quartz-albite pegmatite</td>
<td>7</td>
<td>900</td>
<td>3</td>
<td>2.12</td>
</tr>
</tbody>
</table>

It is particularly interesting that intramineral dioritite and late biotite dioritite magmas probably were less dense than Romeral diorite magma. This is surprising in that the iron-enriched dioritite dike rocks are more dense than Romeral diorite. The reason for this unexpected result is that on the basis of the high amphibole and/or biotite contents of the dioritites, their magmas are inferred to have contained more water than the diorite. Evidently the tendency toward increased magmatic density caused by their increased iron contents, was overbalanced by the tendency toward decreased density caused by their increased water
Viscosities as well as densities of magmas must be estimated if gravitational sorting of mineral grains is to be evaluated as a possible mechanism for magmatic differentiation. Shaw (1965) gave curves for experimentally determined viscosities of several magmas at super-liquidus temperatures. He also empirically related magmatic viscosities to the ratio \( R = \text{Si} + \text{Al} / \text{O} \), where Si, Al, and O are the total numbers of silicon, aluminum, and oxygen atoms in 100 grams of molten rock. For Romeral diorite with 2 percent of water, \( R = 0.426 \). This suggests a viscosity of about \( 10^3 \) poises at the liquidus \((1175^\circ C)\). Similarly, for intramineral dioritite magma with 7 percent of water, \( R = 0.370 \), which suggests a viscosity of about \( 10^3.5 \) poises at the dioritite liquidus \((1000^\circ C)\).

Shaw extrapolated his viscosity curves to sub-liquidus temperatures at about the slopes determined experimentally at super-solidus temperatures (minimum viscosity curve, Figure 79). However, Murase and McBirney (1973) measured viscosities of dry andesitic magmas, glasses, and crystalline rocks and found that viscosity increases abruptly as temperature falls below the liquidus (maximum viscosity curve, Figure 79). They attributed this phenomenon to the combined effects of increasing molecular association and crystal suspension. In water-bearing magmas such as the Romeral
diorite magma and its progressively "wet" differentiates, however, the effects of increasing molecular association probably would have been partially offset by the effects of \( \text{OH}^- \) interference with Si-O and Al-O bonding. Thus the viscosities of the Romeral magmas probably would not have increased as abruptly as those of the dry magmas studied by Murase and Mc Birney.

Bartlett (1969) suggested a different approach to the problem of magmatic viscosity. He derived an equation relating grain density and grain-size distributions to magmatic viscosity (Table 14). Given mineral densities and approximate magmatic density, and assuming a given temperature difference between the floor and ceiling of the magma chamber, trial magmatic viscosities can be assigned, and grain-size distributions can be calculated. As shown in Table 14, the size of ilmenite, augite, and plagioclase grains observed in Romeral diorite are consistent with grain-size distributions that theoretically would result from crystal settling and convective back-diffusion of mineral grains in a magma with a viscosity of about \( 10^{4.5} \) poises.
The Bartlett equation for grain-size distributions resulting from crystal settling and convective back-diffusion in magmas of various viscosities is:

$$\ln\left(\frac{N_{P2}}{N_{P1}}\right) = -\left\{\frac{(\rho_p - \rho_m)g}{18\kappa K} \left(\frac{1700vK}{\alpha_{P}(T_1 - T_2)}\right)^{1/2}\right\} dp^2$$

where, $N_{P2}$ = mineral particle population density in magma near the ceiling

$N_{P1}$ = mineral particle population density in magma near the floor

$\rho_p$ = specific gravity of settling mineral (g cm$^{-3}$)

$\rho_m$ = specific gravity of magma (g cm$^{-3}$)

$g$ = gravity acceleration (980 cm sec$^{-2}$)

$\alpha_P$ = magma viscosity (poise)

$T_1 - T_2$ = floor temperature minus ceiling temperature (°K)

and $dp$ = particle diameter.

The derivation of this equation, together with a complete explanation of its symbols, and a discussion regarding its possible applications, is given by Bartlett (1969).

Assuming magmatic viscosity $= 10^{4.5}$ poises, and $T_1 - T_2 = 125^\circ$C, calculated mineral-grain distributions fit the grain distributions of Romeral diorite as follows:

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Observed grain diameter (mm)</th>
<th>Trial grain diameter (mm)</th>
<th>Calculated $N_{P2}/N_{P1}$ for the trial diameter</th>
</tr>
</thead>
<tbody>
<tr>
<td>ilmenite</td>
<td>0.01-0.25</td>
<td>0.1</td>
<td>.82</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.25</td>
<td>.28</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.5</td>
<td>.006</td>
</tr>
<tr>
<td>Plagioclase</td>
<td></td>
<td>1.0</td>
<td>.89</td>
</tr>
<tr>
<td>core (An47)</td>
<td>0.06-4.0</td>
<td>2.0</td>
<td>.61</td>
</tr>
<tr>
<td>inner rim (An38)</td>
<td></td>
<td>4.0</td>
<td>.15</td>
</tr>
<tr>
<td>augite</td>
<td>0.2-1.0</td>
<td>0.5</td>
<td>.85</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.0</td>
<td>.53</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.0</td>
<td>.08</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>(in melaniorite)</td>
<td>0.08-0.2</td>
<td>.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>.5</td>
<td>.98</td>
</tr>
</tbody>
</table>

Table 14
MAGMATIC VISCOSITY AND GRAIN-SIZE DISTRIBUTION
Figure 77. Viscosities of Romeral diorite and intramineral dioritite magmas as a function of temperature. The minimum viscosity curve for Romeral diorite was calculated by the method of Shaw (1965). The maximum curve is from viscosity measurements on dry Mount Hood andesite by Murase and Mc Birney (1973). The approximate viscosity curve combines the superliquidus viscosity calculated by the Shaw method with an approximate viscosity calculated from observed grain-size distributions, using Bartlett's equation (Table 14). The approximate viscosity curve for intramineral dioritite is an interpolation between the superliquidus viscosity determined by the Shaw method and the high viscosity of solid rock at the solidus temperature.
Thus, assuming that crystal settling and convective mixing did cause the size distributions of plagioclase, ilmenite, and augite grains in Romeral diorite, then the viscosity of the diorite probably was about $10^{4.5}$ poises when settling and convection stopped. This presumably would have occurred at some temperature below 1075 °C, the initial ilmenite-augite crystallization temperature, but above 900 °C, the initial amphibole crystallization temperature (approximate viscosity curve for Romeral diorite, Figure 79). Plagioclase rims and interstitial amphibole probably formed after settling and convection had stopped. The approximate viscosity curve for intramineral diorite (Figure 79) is an interpolation between the superliquidus viscosity as determined by the Shaw method and the high viscosity of solid rock at the solidus temperature.

**Temperatures of Thermal Metamorphism and Hydrothermal Metasomatism**

**Amphibolite to greenschist-facies minerals:** Experimental studies of equilibrium phase relations between amphibolite and greenschist-facies mineral assemblages by J. G. Liou (1972, in press) indicate that the assemblage actinolite + oligoclase + epidote, which is typical of altered zones around El Romeral, is stable between 550 °C and 475 °C at 2 kb. Within this range anorthite contents of plagioclases and Al contents of actinolites vary directly with temperature. Above 550 °C actinolite converts to hornblende, and oligoclase
to andesine. The resulting assemblage, typical of the amphibolite facies, may include clinopyroxene, magnetite, sphene, and sulfides. Below 475° oligoclase is converted to albite, and the resulting greenschist-facies assemblage typically includes actinolite, chlorite, epidote, quartz, calcite, magnetite, and sphene.

Most of the alteration-mineral assemblages in and around the Romeral magnetite deposits are transitional between the amphibolite and the greenschist facies. Assuming equilibrium was approached, they probably formed at temperatures between 550 and 475°C. The early dioritized rocks and the oligoclase-diopside rocks have amphibolite-facies affinities, the ore-related actinolitic assemblages are transitional, and the late clinozoisite-oligoclase, chloritic, and quartzofeldspathic assemblages have greenschist-facies affinities. Their probable temperatures of formation were as follows:

1. Dioritized rocks within about 100 m of Romeral diorite contacts contain secondary plagioclase (An27-32), actinolitic hornblende and diopside. This assemblage suggests "wet" metasomatism at temperatures near 550°C.

2. Oligoclase-diopside rocks probably formed during relatively dry prograde thermal metamorphism at temperatures near 550°C.

3. The actinolite-plagioclase assemblage associated with the Main orebody suggests temperatures of formation near 550-500°C. Secondary plagioclases
range in composition from An$_{32}$, which suggests temperatures near 550°, to An$_{22}$, which suggests temperatures near 500°C. Amphibole from felty green amphibolitized phyllites west of the Main orebody is transitional between actinolite and hornblende (Appendix G), which suggests temperature of formation near 550°C. Amphibole from actinolite veins that cut the Main orebody has a lower Al content (Appendix C-1), which suggests temperature of formation near 500°C.

4. The clinozoisite-oligoclase (An$_{12-22}$) assemblage that is superimposed on the outer edges of the actinolite-plagioclase zone probably formed at about 475 ± 50°C on the basis of the An content of its plagioclase.

5. The quartzo-feldspathic rocks contain plagioclases ranging in composition from An$_8$ to An$_{20}$ along with quartz and minor microcline. This assemblage suggests temperatures of formation near 475 ± 50°C at pH above about 4 (Liou, 1973; Meyer and Hemley, 1967, p. 212-213).

Ore magnetite: Although most Romeral ore deposition accompanied actinolite-plagioclase deposition at estimated temperatures between 550 and 500°C, some ore deposition continued during formation of the clinozoisite-oligoclase-bearing rocks, at estimated temperatures between 500 and
475°C. Deposition of ore magnetite was nearly complete, however, when chlorite-albite-magnetite veinlets formed at estimated temperatures below 475°C.

Vein magnetite: Vein magnetite, like the ore magnetite of El Romeral, has low impurity content (Appendix F), and it coexists with actinolite, chlorapatite, clinozoisite, quartz, and calcite, an assemblage that suggests sub-magmatic temperatures of formation near 500 ± 50°C. The fact that the actinolite-magnetite-apatite veins cut quartzo-feldspathic rocks, which are in part post-ore in age, and which formed at estimated temperatures near 475°C., suggests that the veins may also have formed at temperatures near 475°C. Alternatively, the veins may have tapped hotter hydrothermal fluids than those responsible for quartzo-feldspathic alteration, and the veins, like the ore, may have formed at temperatures between 550 and 475°C.

Post-ore argillic alteration: Dickite, kaolinite, pyrophyllite, sericite, and martite that are superimposed on actinolitized and chloritized biotite phyllites and schists northeast of the Main orebody probably represent retrograde hydrothermal alteration following magnetite ore deposition. The eastward progression from dickite, to kaolinite, to pyrophyllite, to sericite could represent an eastward increase in the ratio of ak⁺/ah⁺ at about 350 ± 50°C (Meyer and Hemley, 1967, p. 212). Such an eastward increase in ak⁺/ah⁺ could have resulted from an eastward increase in
the relict biotite contents of the pre-argillic phyllites and schists (p. 173).

Thermal Gradients

Vertical thermal gradients: Steep geothermal gradients probably existed during ore deposition at El Romeral. While ore deposition took place at about 7 km depth (p. 189) and at temperatures near 500 ± 50°C (p. 213), a body of water-saturated residual dioritic magma is inferred to have existed about 2 km below the deposit, at temperatures near 1000°C (p. 204, p. 205). Thus the thermal gradient between the zone of ore deposition and the subjacent body of magma was approximately 500° in 2000 m. Above the zone of ore deposition the thermal gradient was approximately 475° in 7 km (Fig. 77).

Figure 78. Vertical thermal gradient inferred to have existed during Romeral ore deposition, compared to normal geothermal gradients according to Rittmann (Kern and Weisbrod, 1967, Fig. 13, p. 90).
The clinozoisite-oligoclase assemblage that formed east of the Romeral magnetite deposits, and the quartzo-feldspathic assemblage that formed to the west, probably both formed at temperatures in the range 475 ± 50°C (Fig. 78).

Figure 79. Simplified east-west section through El Romeral, showing approximate thermal gradients inferred to have existed during ore deposition. Vertical gradients were estimated on the basis of considerations explained in the text (p. 216). Lateral gradients were estimated on the basis of mineral zonation and by analogy with heat-flow models postulated by Jaeger (Verhoogen and others, 1970, p. 558). Effects of residual heat from the La Liga andesite porphyry and Romeral diorite plutons have not been taken into account.
Decay of thermal gradients with time: Superimposition of late argillic alteration on earlier actinolitic and chloritic alteration-mineral assemblages implies post-ore decay of the abnormal vertical thermal gradients that accompanied ore deposition. The Romeral clay-mineral assemblage suggests temperatures of formation near 350 °C (p. 215). Thus by the time the clay minerals formed, the thermal gradient between the Romeral ore deposits and the surface had decreased to approximately 325° in 7 km.

Compositions of Hydrothermal Fluids

Water commonly is the most abundant fluid associated with volcanic and geothermal areas, and by analogy, water probably was the most abundant fluid involved in formation of the Romeral magnetite deposits. Most water, however, passed through the deposits leaving almost no trace, except as combined in hydroxyl-bearing minerals and as trapped in microscopic fluid inclusions.

The extent to which non-magmatic water may have been involved in hydrothermal iron transport is unknown. If the La Liga andesite and Romeral diorite plutons represent bodies of magma that fed submarine andesite flows and sills of the Arqueros formation, then the Romeral deposits probably formed about 7 km ± 3.5 km below the sea floor, and it is possible that the wallrocks were saturated with seawater, and that connate water was involved in hydrothermal transport. Oxygen isotope studies might provide an answer but are beyond the
scope of this study.

Fluid inclusions: Fluid inclusions were observed in apatite and quartz from actinolite-magnetite-apatite veins, in apatite from an apatite-rich cataclasite dike, and along actinolite grain boundaries in samples from the Main orebody. Complete analysis of fluid inclusions would involve measurement of filling and freezing temperatures, microchemical analysis, and stable isotope analysis. Although such measurements are beyond the scope of this study, simple optical, mineralogical, and chemical observations were made, and these observations, along with interpretations based upon them, are given below.

Four-phase fluid inclusions in quartz grains from actinolite-magnetite-apatite veins have the following observed characteristics:

1. inclusion size: 0.003 to 0.009 mm
2. shape: subhedral negative quartz crystal shape
3. distribution: random
4. content:
   a. colorless liquid, r.i. = 1.345 ± .015 (probably aqueous)
   b. colorless gas bubble
   c. green anisotropic prismatic crystal with inclined extinction and r.i. = 1.64 (probably actinolite)
   d. orange-brown flake that skitters because of Brownian movement (probably hematite)
5. liquid/liquid + gas ratio = 0.97 ± .02 (by volume).

The random distributions and consistent four-phase contents of these inclusions suggest that they are primary and that their enclosed actinolite and hematite grains are daughter minerals that crystallized after fluid entrapment.
This suggests that when the fluid was entrapped it contained all of the chemical constituents of actinolite ($\text{Ca}_{1.9}\text{Na}_{0.1}$)

\[
\left(\text{Mg}_{3.2}\text{Fe}^{3+}_{0.1}\text{Fe}^{2+}_{1.4}\text{Al}_{0.2}\right)\left(\text{Si}_{7.7}\text{Al}_{0.3}\right)\text{O}_{22}^0\text{OH}_2
\]

and hematite ($\text{Fe}_2\text{O}_3$) in solution. Furthermore, these constituents were present in sufficient concentrations to cause crystallization of actinolite and hematite as temperature decreased after entrapment. The high ratio of liquid/liquid + gas suggests that the fluid was trapped as a supercritical fluid at temperatures and pressures consistent with those postulated for vein formation at El Romeral ($500 \pm 50^\circ\text{C}$, $1.4 \pm 0.7 \text{ kb}$), (Roedder, 1967, p. 531, Fig. 12.2).

Two-phase fluid inclusions in tubelike acicular apatite crystals from actinolite-magnetite-apatite veins have the following observed characteristics:

1. inclusion size: 0.0001 by 0.006 mm
2. shape: elongate
3. distribution: in tubelike acicular apatite crystals
4. orientation: parallel to the long axes of host apatite crystals
5. content:
   a. colorless liquid, r.i. = 1.345 ± 0.015, (probably aqueous)
6. liquid/liquid + gas ratio = 0.97 ± .02

The shapes and orientations of these fluid inclusions suggest that their host apatite crystals tended to grow as hollow, fluid-filled tubes, but that as crystal growth progressed, these fluid-filled tubes were sealed. Such tubelike growth suggests temporary starvation of the centers of basal-plane faces relative to faster-growing crystal edges that would have had easier access to moving nutrient
solutions. This in turn implies that materials were supplied to growing vein apatite crystals by mass flow of fluid, rather than by diffusion (Roedder, 1967, p. 519), which presumably would have provided a more evenly distributed supply of materials. By contrast, apatite grains in the metasomatic Romeral orebodies and in the actinolitized rocks that surround them, are conspicuously devoid of fluid inclusions. This suggests that diffusion was more important than mass flow of fluid, in transmitting chemical constituents to and from replacement sites in the orebodies, and in the actinolitized rocks that surround them.

Several observations and two simple chemical tests were made on fluid inclusions in apatite from an apatite-rich cataclasite dike, as summarized below:

1. inclusion size: 0.005 to 0.05 mm
2. shape: irregular but generally elongate parallel to the foliation of the cataclasite
3. distribution: random
4. content:
   a. colorless liquid, r.i. = 1.345 ± .015 (probably aqueous)
   b. no gas bubbles visible at 500x magnification
   c. A colorless cubic crystal (probably halite) was found in one fluid inclusion.
5. Cl⁻ test:
   a. About 5 grams of apatite were crushed in about 1 cc of distilled water to liberate fluids from fluid inclusions.
   b. Two drops of AgNO₃ were added.
   c. A white precipitate (AgCl) formed, indicating the presence of Cl⁻ ions in the wash water.
   d. Little of no apatite probably was dissolved in the cool, distilled wash water during crushing. Therefore, the Cl⁻ ions probably came from fluid inclusions.
6. pH test:
   a. Step 5a was repeated.
   b. The pH of the wash water was measured with pH paper.
   c. The wash water tested neutral (pH-7).

These fluid inclusions probably are secondary. Their orientations parallel to cataclasite foliation suggests that they formed during tectonic emplacement of the cataclasite dike, which probably occurred during late stages of Romeral ore deposition (p. 138). Chemical testing and the rare occurrence of halite crystals in the fluid inclusions suggest that the entrapped fluids are concentrated chloride brines. This suggests that late hydrothermal fluids were more saline than earlier ones. The crude pH test suggests that the entrapped fluids are somewhat alkaline. Distilled water in equilibrium with atmospheric CO₂ would be slightly acidic, but the pH test indicated that the wash water was neutral after the apatite was crushed in it. Therefore it is postulated that the wash water was neutralized by alkaline fluids released from the fluid inclusions by crushing them. If the entrapped fluids were indeed alkaline, this would suggest that after reaction with wallrocks and deposition of magnetite and apatite, the "spent" hydrothermal fluids were slightly alkaline.

No fluid inclusions were found within actinolite grains, but fluid inclusions are common along actinolite grain boundaries in samples from the Main orebody. These fluid inclusions are visible only at magnifications of about 500x or greater. They are vermicular in shape, and they contain
colorless, probably aqueous fluids. Although these fluid inclusions are probably secondary and thus cannot be regarded as samples of the ore-forming hydrothermal fluids, they do represent aqueous fluids that passed through the Romeral deposits at some time.

**Acidic hydrothermal constituents:** The presence of apatite at El Romeral suggests that weak acids of the phosphate radical, such as $\text{H}_3\text{PO}_4(\text{aq})$ and $\text{H}_2\text{PO}_4^-$ probably were constituents of the ore-forming hydrothermal fluids of El Romeral. The presence of chloride ions in the apatite and scapolite of El Romeral suggests that chlorine-bearing species such as $\text{Cl}_2$, $\text{HCl}$, and $\text{Cl}^-$ were also important constituents of those fluids. By contrast, the absence of fluorapatite (by far the most common type of apatite in most igneous rocks and hydrothermal veins) suggests that fluorine was absent from the Romeral hydrothermal fluids.

The presence of carbonate ions in Romeral calcite and apatite suggests that acids of the carbonate radical, such as $\text{H}_2\text{CO}_3(\text{aq})$ and $\text{HCO}_3^-$ were present in the Romeral hydrothermal fluids, and the presence of borate ions in the tourmaline and the axinite of El Romeral, suggests that acids of the borate radical, such as $\text{H}_3\text{BO}_3(\text{aq})$ and $\text{H}_2\text{BO}_3^-$ also were present in those fluids. In addition, the presence of late sulfide minerals, such as pyrite and chalcopryite, suggests that $\text{S}_2$, $\text{H}_2\text{S}$, $\text{HS}^-$, and $\text{S}^2-$ were constituents of late hydrothermal fluids of El Romeral.
Chemical Exchanges between Hydrothermal Fluids and Wallrocks

Assuming that altered La Liga andesite porphyry and Romeral diorite originally had about the same compositions as relatively unaltered La Liga andesite porphyry and Romeral diorite, and assuming that changes in the compositions of the altered rocks were brought about through chemical exchanges with hydrothermal fluids, then such chemical exchanges can be evaluated by comparing the compositions of altered and unaltered rocks. Chemical constituents that are more abundant in the altered rocks than in the unaltered rocks presumably were added to the altered rocks and removed from the fluids. By contrast, constituents that are less abundant in the altered rocks than in the unaltered rocks presumably were removed from the altered rocks and taken into solution in the fluids.

An approximate accounting for each chemical constituent added to or removed from the rocks of each of the orebodies and alteration zones in La Liga andesite and Romeral diorite was made as follows:

1. Mean chemical compositions of representative groups of samples of fresh and altered rocks, expressed in terms of weight percentages of oxide constituents, were converted to compositions expressed in terms of grams of oxides per cm$^3$ of rock:

\[
g(\text{oxide}) \times \frac{1}{100g(\text{rock})} \times \frac{1}{100\cdot \text{cm}^3(\text{rock})} = \frac{g(\text{oxide})}{\text{cm}^3(\text{rock})}.\]

The rock densities used in this calculation, like the weight percentages of oxides, were
calculated from the mineralogical compositions of the rocks, as determined from thin-section point counts.

2. Mean chemical compositions of rocks, expressed in terms of grams of oxides per cm$^3$ of rock, were converted to compositions expressed in terms of moles of elements per cubic meter of rock:

   a. $\frac{g(\text{oxide})}{\text{cm}^3(\text{rock})} \div \text{molecular weight (oxide)} \cdot \frac{\text{number of non-oxide atoms per molecule}}{\text{cm}^3} = \frac{\text{moles (element)}}{\text{cm}^3}$

   b. $\frac{\text{moles (element)}}{\text{cm}^3(\text{rock})} \times 10^6 \frac{\text{cm}^3}{\text{m}^3} = \frac{\text{moles (element)}}{\text{m}^3(\text{rock})}$.

3. Mean molar compositions of unaltered rocks were subtracted from those of altered rocks to obtain mean molar changes involved in wallrock alteration:

   \[
   \frac{\text{moles (element)}}{\text{m}^3(\text{altered rock})} - \frac{\text{moles (element)}}{\text{m}^3(\text{unaltered rock})} = \frac{\text{moles (element)}}{\text{m}^3(\text{altered rock})}.
   \]

Mean molar changes involved in conversion of La Liga andesite porphyry to magnetite ore and actinolitic rock, and in conversion of Romeral diorite to quartzo-feldspathic rock are summarized in Table 15. Qualitative changes involved in argillization of actinolitized phyllite also are summarized in Table 15.

In the Main orebody, abundant magnetite and actinolite, along with less abundant oligoclase, clinozoisite, scapolite, and minor apatite and pyrite, replaced La Liga andesite
porphyry. This required exchange of large amounts of Fe, smaller amounts of Mg, and minor amounts of P, S, Cl, and C from the hydrothermal fluids, for large amounts of Si and Al, smaller amounts of Na and Ca, and minor amounts of Ti and K from the replaced rocks. In the actinolite-plagioclase zone, hydrothermal Fe, Mg, S, Ca, C, P, and B were exchanged for wallrock Al, Si, Na, Ti, and K. By contrast, in the quartzo-feldspathic zone, hydrothermal Si, Na, K, and B were exchanged for wallrock Ca, Mg, Al, Fe, Ti, P, C, and Cl. In the post-ore argillic zone, hydrothermal H+, H2O, K, O2, and B were exchanged for Mg, Ca, Si, and Fe from the magnetite-bearing, actinolitized, and chloritized phyllites that were altered to argillites containing dickite, kaolinite, pyrophyllite, sericite, and martite.

Thus the compositions of hydrothermal fluids passing through the Romeral system at any one time probably changed as the fluids moved upward and progressively exchanged chemical constituents with wallrocks of different compositions at progressively shallower depths and lower temperatures. Furthermore, inasmuch as quartzo-feldspathic alteration generally followed actinolitization and ore deposition, and argillic alteration followed quartzo-feldspathic alteration, it would appear that the compositions of Romeral hydrothermal fluids generally changed with time, from Fe-Mg-rich, to Na-Si-rich, to acid-rich.
Table 15

CHEMICAL CHANGES IN THE ALTERED ROCKS OF EL ROMERAL

<table>
<thead>
<tr>
<th>Host Rock</th>
<th>Altered Rock</th>
<th>Moles x 10² added per cubic meter</th>
<th>Moles x 10² removed per cubic meter</th>
</tr>
</thead>
<tbody>
<tr>
<td>La Liga</td>
<td>Ore, Main, and So.</td>
<td>406.2 Fe</td>
<td>-155 Si</td>
</tr>
<tr>
<td>andesite</td>
<td>Extension</td>
<td>4.0 Mg</td>
<td>96.2 Al</td>
</tr>
<tr>
<td>porphyry</td>
<td>Orebodies</td>
<td>1.7 P</td>
<td>35.7 Na</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.6 Cl</td>
<td>23.2 Ca</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.2 CO</td>
<td>2.8 Ti</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.1 S</td>
<td>1.9 K</td>
</tr>
<tr>
<td>La Liga</td>
<td>actinolite-plagioclase rock</td>
<td>56.8 Fe</td>
<td>-33.3 Al</td>
</tr>
<tr>
<td>andesite</td>
<td></td>
<td>20.1 Mg</td>
<td>27.0 Si</td>
</tr>
<tr>
<td>porphyry</td>
<td></td>
<td>20.6 S</td>
<td>12.6 Na</td>
</tr>
<tr>
<td></td>
<td></td>
<td>7.7 Ca</td>
<td>3.6 Ti</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5.3 C</td>
<td>0.4 K</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.3 P</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.6 Cl</td>
<td></td>
</tr>
<tr>
<td>La Liga</td>
<td>actinolite-sphene rock</td>
<td>87.2 Ca</td>
<td>-87.0 Al</td>
</tr>
<tr>
<td>andesite</td>
<td></td>
<td>38.3 Mg</td>
<td>-37.9 Si</td>
</tr>
<tr>
<td>porphyry</td>
<td></td>
<td>34.3 C</td>
<td>-34.6 Na</td>
</tr>
<tr>
<td></td>
<td></td>
<td>32.3 Ti</td>
<td>1.1 K</td>
</tr>
<tr>
<td></td>
<td></td>
<td>22.9 Fe</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>9.6 P</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>3.5 Cl</td>
<td></td>
</tr>
<tr>
<td>Romeral</td>
<td>oligoclase-albite-quartz rock</td>
<td>33.7 Na</td>
<td>-41.3 Ca</td>
</tr>
<tr>
<td>diorite</td>
<td></td>
<td>29.9 Si</td>
<td>-22.6 Mg</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.8 K</td>
<td>-6.3 Fe</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4.6 Ti</td>
<td>1.1 Al</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.1 Cl</td>
<td>0.3 P</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.3 B</td>
<td></td>
</tr>
<tr>
<td>Romeral</td>
<td>quartz-oligoclase-albite rock</td>
<td>75.3 Si</td>
<td>-43.5 Ca</td>
</tr>
<tr>
<td>diorite</td>
<td></td>
<td>3.5 Na</td>
<td>-32.2 Al</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.3 B</td>
<td>-30.2 Mg</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.2 K</td>
<td>-9.2 Fe</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3.6 Ti</td>
<td>0.3 P</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.1 Cl</td>
<td></td>
</tr>
</tbody>
</table>
Table 15 (continued)

CHEMICAL CHANGES IN THE ALTERED ROCKS OF EL ROMERAL

<table>
<thead>
<tr>
<th>Host Rock</th>
<th>Altered Rock</th>
<th>Moles x $10^2$ added per cubic meter</th>
<th>Moles x $10^2$ removed per cubic meter</th>
</tr>
</thead>
<tbody>
<tr>
<td>Romeral diorite</td>
<td>oligoclase-quartz-</td>
<td>$+ 44.7$ Si</td>
<td>- 36.8 Ca</td>
</tr>
<tr>
<td></td>
<td>microcline rock</td>
<td>$+ 7.7$ Na</td>
<td>- 32.5 Mg</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$+ 6.1$ K</td>
<td>- 10.3 Al</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$+ \text{tr B}$</td>
<td>- 9.8 Fe</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>- 3.2 Ti</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>- 0.3 P</td>
</tr>
<tr>
<td>Romeral diorite</td>
<td>oligoclase-quartz</td>
<td>$+ 32.6$ Si</td>
<td>- 19.3 Ca</td>
</tr>
<tr>
<td></td>
<td>rock</td>
<td>$+ 5.5$ Na</td>
<td>- 13.2 Mg</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$+ 0.2$ K</td>
<td>- 12.4 Al</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>- 4.4 Fe</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>- 1.1 Ti</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>- 0.3 P</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>- 0.1 Cl</td>
</tr>
<tr>
<td>actinolitized,</td>
<td>advanced argillic</td>
<td>$H^+$</td>
<td>Al(?)</td>
</tr>
<tr>
<td>chloritized</td>
<td>rock</td>
<td>$H_2O$</td>
<td>Mg</td>
</tr>
<tr>
<td>phyllite</td>
<td></td>
<td>$K^+$</td>
<td>Ca</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$O_2$</td>
<td>Si</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$B$</td>
<td>Fe</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Na</td>
</tr>
</tbody>
</table>
Chemical reactions: Balanced chemical equations represent mole-for-mole chemical exchanges within a closed system, whereas replacement involves volume-for-volume exchange in an open system. Thus balanced chemical equations do not necessarily represent the replacement process accurately. Nevertheless, it is useful to write hypothetical chemical reactions for the chemical exchanges that take place at sites of solution, deposition, and/or replacement. For example, alteration of andesine to actinolite + oligoclase might be represented as follows:

\[
\text{3NaCaAl}_3\text{Si}_2\text{O}_6 + 4\text{SiO}_2 + 3\text{Mg}^{2+} + 2\text{Fe}^{2+} + 2\text{H}^+ \rightleftharpoons \text{Ca}_2\text{Mg}_3\text{Fe}_2\text{Si}_8\text{O}_{22}(\text{OH})_2 + \text{Na}_3\text{CaAl}_5\text{Si}_{11}\text{O}_{32} + 4\text{Al}^{3+} \]

Thus actinolitization of plagioclase would consume SiO\text{2}, Mg\text{2+}, Fe\text{2+}, and H\text{+}; and it would liberate Al\text{3+}.

On the basis of his experiments on equilibrium solubility of magnetite in hydrothermal fluids, Helz (1971) postulated the following reactions for hydrothermal solution and deposition of magnetite:

1. \( \text{Fe}_2(\text{OH})_5^{+} + \frac{1}{3}\text{H}_2(\text{g}) \rightleftharpoons \frac{2}{3}\text{Fe}_3\text{O}_4(\text{s}) + 7/3\text{H}_2\text{O}(\text{g}) + \text{H}^+(\text{aq}) \)

2. \( \text{Fe}_2(\text{OH})_6(\text{aq}) + \frac{1}{3}\text{H}_2(\text{g}) \rightleftharpoons \frac{2}{3}\text{Fe}_3\text{O}_4(\text{s}) + 10/3\text{H}_2\text{O}(\text{g}) \)

3. \( \text{Fe}^{2+}(\text{aq}) + 4/3\text{H}_2\text{O} \rightleftharpoons \frac{1}{3}\text{Fe}_3\text{O}_4(\text{s}) + 1/3\text{H}_2(\text{g}) + 2\text{H}^+ \)
Helz (1971) further concluded that Fe(OH)$_3^+$ would be the dominant species in neutral or acidic solutions at temperatures above 500°C. Its solubility would be directly proportional to both temperature and acidity. Fe(OH)$_3$ should be the second-most abundant species present at temperatures above 500°C. Its solubility would be directly proportional to temperature and independent of acidity. Fe$^{2+}$ and Fe(OH)$^+$ would be very insoluble at 500°C and at pH above 4 (Helz, 1971, p. 53-55).

At least two reactions might be important with regard to apatite solubility:

1. $3\text{H}_3\text{PO}_4(aq) \rightleftharpoons 3\text{H}^+ + 3\text{H}_2\text{PO}_4^- \quad (K = 10^{-2.1})$

2. $5\text{Ca}^+ + \text{Cl}^- + 3\text{H}_2\text{PO}_4^- \rightleftharpoons \text{Ca}_5(\text{PO}_4)_3\text{Cl}(s) + 6\text{H}^+$

Inasmuch as $\text{H}^+$ appears on the right-hand side of the apatite deposition reaction, solubility of apatite would be enhanced by increased acidity.

Alteration of andesine to clinozoisite + albite would consume SiO$_2$ and Na$^+$, and it would liberate H$^+$ ions, as follows:

$4. \text{Fe(OH)}^+(aq) + \frac{4}{3}\text{H}_2\text{O} \rightleftharpoons \frac{1}{3}\text{Fe}_3\text{O}_4(s) + \frac{1}{3}\text{H}_2(g) + \text{H}^+(aq)$

magnetite
Chloritization of biotite in Romeral phyllites would consume $H^+$ and liberate $K^+$ and $SiO_2$, as follows:

$$2K(Mg,Fe)_{3}AlSi_{3}O_{10}(OH)_{2} + 4H^+ \rightleftharpoons$$

$$Al(Mg,Fe)_{5}AlSi_{3}O_{10}(OH)_{8} + (Mg,Fe)^{2+} + 2K^+ + 3SiO_2$$

Argillization of andesine in phyllites north of the Main orebody would consume $H^+$ and liberate silicic acid, as follows:

$$2NaCaAl_{3}Si_{5}O_{16} + 6H^+ + H_2O \rightleftharpoons$$

$$2Na^+ + 2Ca^+ + 3Al_{2}Si_{2}O_{5}(OH)_{4} + 4H_4SiO_4(aq)$$

Hydrothermal Mineral Solubilities

**Effects of temperature, acidity, and salinity on hydrothermal solubility of magnetite:** Experimental work by Helz (1971) indicates that dilute HCl strongly enhances the solubility of magnetite in hydrothermal solutions at temperatures above $300^\circ C$. Helz determined the equilibrium solubility of magnetite in artificial hydrothermal fluids containing NaHCO$_3$, NH$_3$, NaOH, pure water, NaCl, and HCl. In all but the
HCl-bearing fluids, he found negligible magnetite solubility at temperatures up to $300^\circ$C, and a maximum of only 20 ppm Fe at $500^\circ$C. Thus temperature increase alone does not appear to dramatically increase magnetite solubility at temperatures up to $500^\circ$C, and salinity has little effect. By contrast, Helz found that dilute HCl solutions ($10^{-2}$ to $10^{-3}$ molar) produced solubilities ranging from about 30 ppm below $275^\circ$C to over 1000 ppm at $500^\circ$C. Thus increased acidity, at temperatures above $300^\circ$C greatly enhances the hydrothermal solubility of magnetite. On the other hand, if iron were dissolved in a slightly acidic hydrothermal fluid at temperatures above $500^\circ$C, decreasing temperature and/or decreasing acidity would promote magnetite deposition.

Although HCl was the only acid used in the experiments made by Helz, it seems likely that other acids might also enhance the solubility of magnetite at high temperatures. It would seem, for instance, that phosphoric acid, which is the active ingredient in commercial rust removers, might be particularly effective in this regard, even though it is a relatively weak acid.

**Effects of temperature, acidity, and salinity on the hydrothermal solubility of chlorapatite:** Apatite is very insoluble in pure water at room temperature, but its solubility probably is enhanced by acidity, by temperature increase above about $400^\circ$C, and by the presence of dissolved neutral salts. Calcium phosphate, like most carbonates and
phosphates, is insoluble in neutral and alkaline solutions, but dissolves in acid solutions (Krauskopf, 1967, p. 88). Although chlorapatite \((\text{Ca}_5\text{(PO}_4\text{)}_3\text{Cl})\) is even less soluble in pure water than simple calcium phosphate \((\text{Ca}_3(\text{PO}_4)_2)\), its solubility, like that of calcium phosphate, probably is enhanced by increased acidity. As an extreme example, apatite is quickly dissolved at room temperature in concentrated nitric acid. According to Smith (1963, p. 244), hydrothermal solubility of apatite probably also is enhanced by temperature increase above about 400°C, and by increase in the concentration of dissolved neutral salts.

Chemical Environments of Hydrothermal Ore Deposition and Alteration

\(\text{pH}\): Slight acidity, perhaps resulting from dissociation of acids such as \(\text{HCl}, \text{H}_3\text{PO}_4,\) and \(\text{H}_2\text{CO}_3\), would have enhanced the capacity of the Romeral hydrothermal fluids to carry ore-forming constituents. However, the fact that secondary oligoclase was stable along with magnetite suggests that the hydrothermal fluids responsible for ore deposition were not very acidic at sites of magnetite deposition. Furthermore, to the extent that actinolitization of andesine consumed \(\text{H}^+\) ions, such actinolitization would have tended to increase \(\text{pH}\) and thereby promote deposition of magnetite and apatite.

The abundant secondary plagioclase \((\text{An}_8-20)\) in the altered rocks of the quartzo-feldspathic zone suggests that the hydrothermal fluids responsible for quartzo-feldspathic alteration were nearly neutral, and that the ratio of \(\log_{10}\text{NaCl/}^{\text{m}}\text{HCl}\) was
greater than about three in those fluids (Meyer and Hemley, 1967, p. 212-215).

By contrast, the advanced argillic alteration of actinolitized and chloritized phyllites north of the Main orebody probably involved strongly acidic hydrothermal fluids. Such strong hydrogen-ion metasomatism probably required dissociation of a strong acid, such as H₂SO₄ (Meyer and Hemley, 1967, p. 220-223). This suggests convective mixing of late, H₂S-bearing hydrothermal fluids of magmatic derivation, with oxygenated groundwaters, to produce H₂SO₄.

Oxidation of magnetite to martite in the argillic zones probably also was caused by reaction of magnetite with such acidic, oxygenated hydrothermal fluids.

**Oxygen fugacity:** Experimental work on the stability fields of magnetite, actinolite, clinozoisite, and other minerals, suggests that oxygen fugacities were low during ore deposition and associated actinolitic and clinozoisitic alteration. According to Ernst (1967, p. 57), ferrotremolite (and presumably actinolite) form at oxygen fugacities below about fO₂ = 10⁻²⁵ bars at 500°C and 3 kb total pressure. At progressively higher oxygen fugacities, hedenbergite + magnetite (fO₂ = 10⁻²⁵ to 10⁻²² bars), andradite + magnetite (fO₂ = 10⁻²² to 10⁻¹⁸ bars), and hematite (fO₂ = 10⁻¹⁸ bars) (Lindsley, 1962-63) are favored. Thus the abundance of actinolite and magnetite at El Romeral, and the absence of hedenbergite, the scarcity of garnet, and the absence of
primary hematite, suggest that oxygen fugacity was low
during formation of the Romeral ore deposits, probably at
about the $f_{O_2}$ of the quartz + magnetite + fayalite + $O_2$
buffer reaction ($f_{O_2} = 10^{-24.5}$ at $500^\circ C$, and 2 kb) (Buddington
and Lindsley, 1964, p. 315). Experiments by Holdaway (1972,
in press) on the thermal stability of Al-Fe epidote as a
function of $f_{O_2}$ and Fe content, suggest that in the presence
of abundant iron, and at $550-475^\circ C$, clinozoisite is favored
over epidote only at low oxygen fugacity. Thus the abun-
dance of clinozoisite, rather than epidote, in association
with the Romeral magnetite deposits is consistent with the
low oxygen fugacity postulated on the basis of the magnetite-
actinolite association.

Sulfur fugacity: Pyrite and chalcopyrite generally are
late to post-ore in age at El Romeral. This suggests that
sulfur fugacity was very low during deposition of most of
the Romeral magnetite, but that it increased somewhat during
late ore deposition. Sulfur fugacity diagrams by Holland
(1965) suggest that if temperatures were near $400^\circ$ when
pyrite-magnetite veinlets formed, and if $f_{O_2}$ was near that
of the quartz-fayalite-magnetite-$O_2$ buffer reaction, then
$f_{S_2}$ was probably near $10^{-7}$ atm when the pyrite formed.
Actually, $f_{S_2}$ was probably somewhat higher than $10^{-7}$ atm,
because the magnetite-pyrite veins probably formed at tem-
peratures nearer $475^\circ$ than $400^\circ C$. 
EVALUATION OF PREVIOUS MODELS FOR PROCESSES OF ORE FORMATION

It has previously been suggested that the iron deposits of Chile were formed by magmatic segregation followed by ore-magma injection (Geijer, 1931; Brüggen, 1934), by pneumatolytic and hydrothermal replacement (Ruiz, 1943, 1965), by deuteric release (Ruiz and others, 1967), by remobilization of iron from buried Precambrian iron formation and other iron-rich rocks (Park, 1972), and by release of iron from basaltic and oceanic crust and pelagic sediments during progressive partial melting along subduction zones (Sillitoe, 1972). It is now possible to evaluate these processes in light of the evidence described herein.

Magmatic Segregation and Ore-Magma Injection

Geijer (1931) supposed that the andesitic host rocks of the iron deposits of El Algarrobo are younger than the surrounding granodiorite into which they were intruded as the lower part of a volcanic feeder. He proposed that a residual magma of iron oxides was then injected into the andesite to form the ore.

Geijer's hypothesis cannot apply to El Romeral for two reasons. The andesitic and schistose host rocks are older than the surrounding diorite, and the magnetite-actinolite deposits formed by hydrothermal replacement rather than by magmatic injection (chapter 7). Still, there is some evidence for magmatic segregation at El Romeral, and that
element of Geijer's hypothesis may be applicable.

Brüggen (1934) was the first to describe an aplitic quartzo-feldspathic zone at El Algorrobo. He proposed "that the 'aplite' was derived from a granodioritic magma, that by process of segregation, lost its ferromagnesian minerals; these minerals were concentrated at depth and subsequently injected into the meta-andesites as a mixture of magnetite and amphibole" (Ruiz and others, 1968, p. 332).

Brüggen's observation of the iron-deficient "aplite" zone was important. Similar quartzo-feldspathic zones are present near most of the important magnetite deposits of northern Chile, including El Algorrobo, Ojos de Agua, Alcaparra, and Huantemé (Ruiz and others, 1968); El Tofo, Cerro Imán, and El Romeral (personal observation, 1970). However, Brüggen's hypothesis is inapplicable to El Romeral on two major accounts. The quartzo-feldspathic zones were formed by post-magmatic alteration rather than by magmatic segregation (p.160-169), and the magnetite-actinolite deposits were formed by hydrothermal replacement rather than by magmatic injection (p.106-126).

Pneumatic and Hydrothermal Replacement

Ruiz (1943, 1965) recognized that the magnetite and associated minerals at Algorrobo were emplaced predominantly by fracture filling and metasomatic replacement. He attributed the replacement and vein filling to a pneumatolytic or in part hydrothermal phase composed of H\textsubscript{2}O, CaCl\textsubscript{2}, Cl\textsubscript{2},
NaCl, KCl, FeCl₃, SiCl₄, PCl₄, B₂O₃, H₂S, and CO₂, and derived from granodioritic magma. He further suggested that this fluid phase probably separated when the magma was completely liquid or had only a small proportion of crystals.

Ruiz' vein-filling and replacement hypothesis probably is essentially correct for the hydrothermal stage of iron concentration at El Romeral. However, there are minor problems with the details of the process. For instance, the Romeral magnetite deposits are not early magmatic but are post-magmatic with respect to the neighboring diorite. The ore-bearing fluid phase did not separate from the diorite itself but probably from an iron-enriched, water-enriched residual magma represented by the intramineral dikes and derived from the diorite by processes of magmatic differentiation. Another minor problem is that experimental work by Helz (1971) has shown that chloride complexes are not important contributors to the hydrothermal solubility of iron. Furthermore, thermodynamical calculations by Krauskopf (1964) have shown that although volatilities of iron chlorides may become important where HCl is exceptionally abundant, volatility is an unlikely explanation for iron transport in general. Thus it is questionable how much Romeral iron was transported as FeCl₃, as suggested by Ruiz.

Deuteric Release

Mackin and Ingerson (1960) explained the origin of the
magnetite deposits of Iron Springs, Utah, in terms of deuteric release of iron from bleached rocks adjacent to magnetite veinlets in a nearby intrusion of quartz monzonite. Mackin (1968) suggested that the magnetite veinlets served as "roots" through which deuteric ore-forming fluid was drawn from the crystal mush, thus forming bleached, iron-depleted selvage rocks along the veinlet walls, and supplying iron, which diffused through the fluids in the veinlets to sites of deposition in limestones along the margins of the intrusion.

Ruiz and others (1968) suggested that bleached, aplitic, quartzo-feldspathic rocks common near many Chilean magnetite deposits were formed by deuteric alteration of dioritic or granodioritic host rocks and that replacement of the ferromagnesian minerals in those rocks would have liberated large quantities of iron, which might have been redeposited to form the orebodies.

Inasmuch as bleached, aplitic, quartzo-feldspathic rocks are closely associated with the magnetite deposits of El Romeral, the deuteric release hypothesis is attractive, in that it would explain both the origin of the bleached rocks and the origin of the orebodies. However, several features of the geology of El Romeral do not seem to fit the deuteric release model.

First, the selvaged veinlets of El Romeral cut deuterically actinolitized Romeral diorite that is inter-penetrated by actinolite veinlets. Therefore, the diorite
probably was solidified, deuterically actinolitized, fractured, and cut by stockworks of actinolite veinlets before being cut by the selvaged magnetite veinlets. Thus the selvaged veinlets probably did not form by deuteric release of fluid from the diorite crystal mush.

Second, although ferromagnesian silicate minerals were destroyed during aplitization of Romeral veinlet-selvage rocks, the aplitic selvage rocks (as represented by several samples from one selvage) contain relict ilmenite and secondary magnetite, and they are not appreciably depleted in iron. Furthermore, dioritic wallrocks along the outer margins of aplitic selvages contain somewhat more iron than average Romeral diorite. These observations suggest that iron moved outward from the magnetite veinlets and was deposited as disseminated magnetite in the selvage rocks and in dioritic rocks adjacent to them. Thus the selvaged magnetite veinlets of El Romeral apparently served as distributaries for post-deuteric, Fe-Na-Si-bearing hydrothermal fluids, rather than as collectors for deuterically released ore-forming fluids.

Third, the aplitized and feldspathized rocks near the Romeral orebodies contain remnants of pervasively actinolitized, magnetite-bearing diorite and phyllite. This suggests that aplitization and feldspathization followed pervasive actinolitization and ore deposition at El Romeral. Further, it suggests that the iron in the Romeral ore and actinolitic zones could not have been supplied by destruction
of the ferromagnesian minerals of Romeral diorite during aplitization, since such aplitization apparently did not happen until after the ore deposits had, for the most part, formed.

Fourth, although the aplitic rocks of the large quartzo-feldspathic zone west of the Romeral orebodies are depleted in iron with respect to Romeral diorite, it is unlikely that enough iron was removed from those rocks to account for the Romeral magnetite deposits and their associated actinolitized rocks. An approximate accounting was made for the amount of each chemical constituent added to or removed from the rocks of each of the orebodies and alteration zones affecting either Romeral diorite or La Liga andesite porphyry, as follows:

1. Chemical changes involved in alteration of andesite to ore and actinolitic rock, and of diorite to aplitic rock, were computed in terms of mean molar gains and losses per cubic meter of rock (p. 225), and Table 15.

2. Mean molar changes per cubic meter of altered rock were multiplied by the areas of corresponding alteration zones. This gave the approximate number of moles of each constituent gained or lost by the rocks in a horizontal, one-meter-thick slice through each alteration zone.

The results of this calculation, given in Table 16, suggest that about 11 times as much iron was added to the rocks of a one-meter slice through the orebodies and actinolitic zones as was removed from the rocks in such a slice.
through the quartzo-feldspathic zone. Only the Main and South Extension orebodies were used in this calculation, because only those orebodies replaced La Liga andesite porphyry, which is relatively uniform in composition. Neither the North Extension orebody, nor the North orebody, nor the northern magnetite pods, nor any of the rocks in their associated actinolitic zones were counted, because they replaced metasedimentary rocks of various and undetermined compositions. Thus the amount of iron added to the Romeral ore and actinolitic zones has been underestimated in this respect, and it would appear that at least 11 times as much iron was added to the rocks of a one-meter slice through the orebodies and actinolitic zones as was removed from the rocks in such a slice through the quartzo-feldspathic zone. Even this statement needs qualification, however, because it is based on the assumption that the mean compositions and areas used in the calculations are sufficiently precise and accurate that the results of the calculation are meaningful. In an effort to determine approximate limits on the precision and accuracy of the mass-balance calculation for iron, estimated magnitudes of discrepancies and errors were compiled, as shown in Table 17.

Analytical discrepancies between total iron contents determined by wet chemical analysis and those determined by calculation from modal analyses ranged from 2 percent to 11 percent of the total iron content as determined by wet chemical methods. Fortunately, however, these discrepancies
nearly cancel each other in the Fe calculations, and therefore they have been ignored.

Since none of the rock types considered is uniform in composition, it is questionable whether the mean compositions of the small groups of samples analyzed adequately represent the mean compositions of the rocks in the zones from which they were collected. Inasmuch as the reliability of the mean increases as the number of samples increases, the reliabilities of the compositions used in these calculations can be ranked in terms of the numbers of samples of each rock type analyzed. In this respect, the ore compositions would appear to be more reliable than the other rock compositions. However, the approximate median iron content of the ore was used in the calculation, and the ore composition was reconstructed from assays and general mineralogical observations, and not calculated from modal analyses. Thus the ore grade used in the calculation can only be regarded as approximate. Nevertheless, by visual estimation from ore-zone sections and plans, the median iron content of the ore probably is close to the mean (for a cutoff grade of 30 percent of iron). Thus an estimated uncertainty range of ±5 percent of iron has been assigned for the ore grade used in the calculations. In the cases of La Liga andesite porphyry, Romeral diorite, and quartzo-feldspathic rock, limits of precision were placed at ± one standard deviation in the mean. If a second value of the mean were calculated from a second set of measurements, the difference between
Table 16
MASS BALANCE, QUARTZO-FELDSPATHIC ROCKS VERSUS ORE AND ACTINOLITIC ROCKS

<table>
<thead>
<tr>
<th></th>
<th>Total, ore total, quartzo and actinolite zones</th>
<th>Total, quartzo feldspathic zone</th>
<th>Ore + actin. + Δqtzo.- feld.</th>
<th>Ratios of additions, removals.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Δ moles x10^5 zone-meter</td>
<td>Δ moles x10^5 Romeral-meter</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Si</td>
<td>-440.9</td>
<td>594.5</td>
<td>+153.6</td>
<td>+1.3x</td>
</tr>
<tr>
<td>Ti</td>
<td>-38.2</td>
<td>-46.1</td>
<td>-84.3</td>
<td></td>
</tr>
<tr>
<td>Al</td>
<td>-397.7</td>
<td>-161.8</td>
<td>-559.5</td>
<td></td>
</tr>
<tr>
<td>Fe(mean)</td>
<td>+1037.9</td>
<td>-93.8</td>
<td>+944.1</td>
<td>+11.1x</td>
</tr>
<tr>
<td>Fe(max.)</td>
<td>+1617.9</td>
<td>-31.4</td>
<td>+1586.5</td>
<td>+55.1x</td>
</tr>
<tr>
<td>Fe(min.)</td>
<td>+608.6</td>
<td>-258.3</td>
<td>+360.3</td>
<td>+ 2.4x</td>
</tr>
<tr>
<td>Mg</td>
<td>+155.9</td>
<td>-331.7</td>
<td>-175.8</td>
<td>- 2.1x</td>
</tr>
<tr>
<td>Ca</td>
<td>+40.6</td>
<td>-483.4</td>
<td>-442.8</td>
<td>-11.9x</td>
</tr>
<tr>
<td>Na</td>
<td>-152.8</td>
<td>+198.7</td>
<td>+45.9</td>
<td>+ 1.3x</td>
</tr>
<tr>
<td>K</td>
<td>- 6.0</td>
<td>+30.3</td>
<td>+24.3</td>
<td>+ 5.0x</td>
</tr>
<tr>
<td>Cl</td>
<td>+19.9</td>
<td>- 1.2</td>
<td>+18.7</td>
<td>+18.7x</td>
</tr>
<tr>
<td>C</td>
<td>+46.3</td>
<td>-tr</td>
<td>+46.3</td>
<td></td>
</tr>
<tr>
<td>P</td>
<td>+21.1</td>
<td>- 3.7</td>
<td>+17.4</td>
<td>+ 5.7x</td>
</tr>
<tr>
<td>S</td>
<td>+83.4</td>
<td>-tr</td>
<td>+83.4</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>+8.3</td>
<td>+ 2.4</td>
<td>+10.7</td>
<td></td>
</tr>
</tbody>
</table>
two means would be expected to be less than the standard deviation in either set (Parratt, 1961, p. 92). In the case of quartzo-feldspathic rock, a range of \( \pm \) one standard deviation puts the lower limit just below zero, which is impossible. Thus, the limits 0 and \( \pm \) one standard deviation were used. In the cases of actinolite-plagioclase rock and actinolite-sphene rock the maximum and minimum values encountered were used as limits.

Uncertainties in the areas of the alteration zones result from partial cover and from uncertain placement of gradational boundaries. Inasmuch as the area of the quartzogeldspathic zone probably is about 1.4 times that of the ore and actinolitic zones, error in the mean composition of quartzo-feldspathic rock is weighted 1.4 times more heavily than error in the mean compositions of ore and actinolitic rocks.

Using the maximum and minimum values given in Table 17, it was calculated that the amount of iron added to the rocks in a 1 m slice through the Main and South Extension ore and actinolitic zones was not less than 2.4 times and not more than 55.1 times the amount removed from the rocks in such a slice through the quartzo-feldspathic zone.

If the quartzo-feldspathic zones had 2.4, or 11.1, or 55.1 times the vertical extent of the ore and actinolitic zones, it is possible that the rocks of the quartzo-feldspathic zone lost as much iron as was deposited in the rocks of the ore and actinolitic zones. However, reconnaissance
Table 17

UNCERTAINTIES IN THE MASS-BALANCE CALCULATIONS

<table>
<thead>
<tr>
<th></th>
<th>La Liga Andesite Porphyry</th>
<th>Roseral Diorite</th>
<th>Main Orebody</th>
<th>South Extension Orebody</th>
<th>Actinolite-Plagioclase Zone</th>
<th>Actinolite-Sphene Zone</th>
<th>Quartzofeldspathic Zone</th>
<th>Ore and actinolitic zones in metasedimentary rocks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Discrepancies between calculated and wet chemical analyses</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>absolute discrepancies</td>
<td>+0.9% Fe</td>
<td>+35% Fe</td>
<td></td>
<td></td>
<td>+0.85% Fe</td>
<td></td>
<td>+0.04% Fe</td>
<td></td>
</tr>
<tr>
<td>relative discrepancies</td>
<td>+2%</td>
<td>+11%</td>
<td></td>
<td></td>
<td>+9%</td>
<td></td>
<td>+5%</td>
<td></td>
</tr>
<tr>
<td>total relative</td>
<td>+2%</td>
<td>+11%</td>
<td></td>
<td>-(+9%)</td>
<td>-(+3%)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>discrepancies</td>
<td>numbers of samples</td>
<td>7</td>
<td>5</td>
<td>1450</td>
<td>70</td>
<td>5</td>
<td>3</td>
<td>15</td>
</tr>
<tr>
<td>location indices</td>
<td>median Fe</td>
<td>4.1% Fe</td>
<td>3.4% Fe</td>
<td>~4.5% Fe</td>
<td>~4.5% Fe</td>
<td>13.8% Fe</td>
<td>1.8% Fe</td>
<td></td>
</tr>
<tr>
<td>estimated accuracy</td>
<td>dispersion indices</td>
<td>±1.8% Fe</td>
<td>±0.7% Fe</td>
<td></td>
<td>9.9% Fe</td>
<td>3.0% Fe</td>
<td>±2%</td>
<td></td>
</tr>
<tr>
<td>standard deviations</td>
<td>ranges:</td>
<td></td>
<td></td>
<td></td>
<td>maximum</td>
<td></td>
<td>maximum</td>
<td></td>
</tr>
<tr>
<td></td>
<td>minimum</td>
<td></td>
<td></td>
<td></td>
<td>17.8% Fe</td>
<td>12.1% Fe</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>relative compositional</td>
<td>dispersing areas (x10^2m^2)</td>
<td>±44%</td>
<td>±20%</td>
<td>±11%</td>
<td>±11%</td>
<td>±28%</td>
<td>±60%</td>
<td>±100%</td>
</tr>
<tr>
<td>estimated uncertainties</td>
<td>in zone-area measurements</td>
<td>±10%</td>
<td>±20%</td>
<td>±30%</td>
<td>±10%</td>
<td>124.8 x10^2 m^2</td>
<td>&gt;105.2 x10^2 m^2</td>
<td></td>
</tr>
<tr>
<td>error resulting from</td>
<td>ignoring altered</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>metasedimentary rocks</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>
observations made at El Tofo and Cerro Imán suggest that quartzo-feldspathic zones may typically have about the same vertical dimensions as their associated orebodies. The magnetite-actinolite orebody at El Tofo is similar to the upper parts of El Romeral in that it has stockworks veining and relatively high apatite and low pyrite contents. By contrast, the magnetite-actinolite orebody at Cerro Imán is similar to the lower parts of El Romeral in that it has relatively low apatite and high pyrite content. It also is surrounded by actinolitized rocks but is associated with only a small zone of quartzo-feldspathic rocks. Thus it would appear that quartzo-feldspathic zones typically have about the same vertical dimensions as their associated orebodies, and they may have less vertical extent than their associated amphibolitic zones. If this is true for El Romeral, then it is unlikely that the rocks of the Romeral quartzo-feldspathic zone lost enough iron to account for the ore and actinolitic zones.

Furthermore, as shown in Table 15, calculations based on mean compositions suggest that most other constituents also are out of balance. More P, C, Cl, and S were added to the rocks in the ore and actinolitic zones than were removed from those of the aplitic zone; more Si, Na, and K were added to the rocks of the aplitic zone than were removed from those of the ore and actinolitic zones; more Ca and Mg were removed from the rocks of the aplitic zone than were deposited in those of the ore and actinolitic zones; and B was added, and
Al, and Ti were partially removed from the rocks of both the aplitic and the ore and actinolitized zones.

This apparent lack of mass balance for most chemical constituents indicates that the process of ore formation cannot be treated as a simple exchange between two parts of a closed system. Instead, both of these parts of the system probably were open to exchange of ions with through-passing hydrothermal fluids. Such fluids probably also were open to chemical exchanges with other masses of rock, magma, water, and possibly air, and they also probably served as media through which ions moved by diffusion between sites of ionic exchange.
Remobilization of Iron
from Buried Precambrian Iron Formation

Park (1972) noted the presence of ferruginous quartzites near El Romeral and El Tofo, and he listed and briefly described other showings of iron-rich metasedimentary rocks around the Pacific Basin. He acknowledged that Precambrian exposures are few, but he asserted that those few commonly contain showings of iron formation, and he suggested that buried Precambrian iron formation may be more common around the Pacific Basin than is generally supposed. With this in mind, he envisioned a process of partial mobilization of iron in deeply buried Precambrian iron formations, or other iron-rich rocks, to explain the origin of metasomatic and volcanic magnetite deposits of the Pacific Basin.

This thought-provoking hypothesis is difficult to substantiate in terms of the rocks exposed at El Romeral. Ferruginous quartzites are present at El Romeral, and iron has, in places, been hydrothermally leached from them. However, it seems unlikely that this iron was redeposited to form the Romeral orebodies. Iron leached from quartzites along the northern contact of the east mass of Romeral diorite apparently was redeposited as hematite, in quartzites about 80 to 145 m from the diorite-quartzite contact, and about 1 km from the magnetite orebodies. Furthermore, although hematite has been leached from quartzites in the post-ore granodiorite batholith east of El Romeral, such
hematite probably was not mobilized until after the Romeral orebodies had formed. However, the ferruginous quartzites dip steeply, and they could extend to considerable depths. It is therefore possible that iron removed from quartzites at greater depths could have been redeposited to form the orebodies.

In the process suggested by Park (1972, p. 347), "intrusive rocks and their companion fluids" work their way up through iron-rich materials. Iron presumably is mobilized by the "companion fluids," which thereby become iron-bearing hydrothermal fluids, or even volatile-rich "ore magmas." Such "companion fluids" might be wallrock fluids, heated by the magma, or magmatic fluids, or both. Low-density wallrock fluids would tend to rise along deep fractures, but since wallrock porosity, permeability, and water content probably decrease with increasing depth, particularly in crystalline rocks, it is questionable how rapidly such fluids would migrate into deep fractures. Another possibility, however, is that dissolved iron might move by diffusion through relatively static wallrock fluids. If such fluids were heated by magmatic intrusions, the thermal gradients established might drive ionic diffusion through them. But again, it is questionable how effective this process would be in low-porosity wallrocks with low water contents. If, by contrast, the "companion fluids" were magmatic, there would be other limitations to be considered. Inasmuch as the solubility of water in silicate magmas
increases with increasing depth, it is unlikely that significant amounts of aqueous fluid can coexist with silicate magma at depths much below 10 km, except in association with small bodies of water-saturated magma undergoing pegmatitic crystallization (Burnham, 1967, p. 66). Thus, if the Romeral magnetite deposits formed at 7 ± 3.5 km depth (p. 189), and if it is doubtful that magmas and magmatic "companion fluids" can coexist below 10 km depth, then the depth interval through which iron could have been mobilized by such fluids would appear to be somewhat limited. Still another possibility is that iron-bearing wallrock fluids might migrate into rising, water-undersaturated magmas and become dissolved in those magmas. However, rates of diffusion of dissolved water in silicate magmas are several orders of magnitude too small to be geologically significant (Burnham, 1967, p. 47), and therefore, even though water would tend to diffuse into water-undersaturated magmas, this process could be effective only along the outer shell of a body of magma. Furthermore, if iron had been added to the Romeral magmas in this way, the rocks that crystallized from those magmas presumably would be iron-rich, and they are not. In fact, Romeral diorite and La Liga andesite porphyry are iron-poor with respect to average andesites, and although intramineral dioritites are somewhat iron-rich, they probably formed by differentiation of dioritic magma (p. 93). Moreover, the intramineral magmas probably rose within altered dioritic and andesitic rocks, and they probably had little or no contact with
iron-rich metasedimentary rocks or wallrock fluids.

Still another problem in applying the iron-remobilization hypothesis to El Romeral is that the Romeral ore and gangue minerals suggest a dioritic source, rather than a quartzitic source, for the Romeral iron. The secondary plagioclase, actinolite, clinozoisite, chlorapatite, and sphene, which are associated with the magnetite in the Romeral orebodies, have close mineralogical and chemical affinities with the dioritic rocks of El Romeral. By contrast, the ores are notably quartz-poor, and thus are not compositionally similar to the quartzites. This suggests that the iron in the orebodies, along with the other ore- and gangue-forming constituents, came from dioritic magmas, rather than from hematitic quartzites. More generally, most of the metasomatic magnetite deposits of the Pacific Basin are closely associated with hypabyssal intrusions of intermediate composition, but very few are associated with known iron-rich metasedimentary rocks.

Furthermore, it is questionable whether buried Precambrian iron formations are extensive around the Pacific Basin. The ferruginous quartzites of Chile probably are neither Precambrian, nor extensive, nor thick (p. 66), and whereas most extensive deposits of Precambrian iron formation are 2000 to 3000 m.y. old, Stewart (1972) gives convincing evidence that rocks of the North American cordilleran geosyncline were deposited directly on oceanic crust after a late Precambrian continental separation less than 850 m.y. ago.
Nevertheless, magnetite-actinolite deposits similar to El Romeral are numerous in the California-Nevada iron belt, where no metasedimentary iron deposits of Precambrian age are known.

It has not been, and perhaps cannot be proven that the iron in the Romeral ore deposits did not come from deeply buried Precambrian iron formations of other iron-rich rocks. However, the classical hydrothermal hypothesis - that metals are derived from magmas by separation of hydrothermal fluids during late stages of magmatic crystallization - appears to explain more of the features observed at El Romeral than does the iron-remobilization hypothesis.

Iron Release by Partial Melting of Subducted Basalts and Pelagic Sediments

Sillitoe (1972) observed that metal zones in the cordillera of South America are aligned parallel to the coast and are generally arranged in the following pattern from west to east: Fe; Cu; Pb, Zn, Ag; Sn, Mo. He suggested that these metals were progressively and sequentially mobilized by progressive partial melting of subducted basaltic crust and pelagic sediments at progressively greater depths. This hypothesis is attractive as an explanation of metal zonation in the South American cordillera. El Romeral and other magnetite deposits of the Chilean iron province are closer to the Peru-Chile trench than are the major copper deposits of Chile. However, as a general explanation
for formation of metasomatic magnetite deposits, this hypothesis is less than satisfying. In western North America, for instance, many magnetite deposits lie far east of the subduction zone that probably existed during their formation. Furthermore, these magnetite deposits are intermingled with Cu, W, Mo, Pb-Zn-Ag, Au, Hg, and Sb deposits. This problem may be more apparent than real, however. If the effects of continental accretion, strike-slip faulting, and basin-and-range faulting and volcanism could be substracted, then two belts of mineral deposits would emerge, and their locations would correspond to those of imbricate subduction zones postulated by Lipman and others (1972) on the basis of K$_2$O contents of volcanic rock suites. Within each of these belts, the pattern observed by Sillitoe would appear to be crudely developed. This lends credence to Sillitoe's contention that metals are sequentially mobilized by progressive partial melting of subducted materials at progressively greater depths along inclined seismic zones. Nevertheless, mobilization of metals as constituents of calc-alkaline magmas by progressive partial melting probably would provide only the first small step in the concentration of metals that leads to the formation of high-grade ore deposits such as El Romeral.
MULTI-STAGE MODEL FOR IRON CONCENTRATION AT EL ROMERAL

In view of rock compositions, structural configurations, and age relationships observed at El Romeral; within the boundary conditions postulated for processes of ore formation at El Romera; within the framework of current plate-tectonic theory; and in light of recent experiments on the behavior of iron in magmas and in hydrothermal fluids; it is now possible to evaluate the possibilities for iron enrichment at several stages of magmatic and hydrothermal development.

Stage 1 - Partial Melting

According to paleomagnetic evidence, South America began to move away from Africa about 150 m.y. ago (Phillips and Forsyth, 1972). At about the same time, andesitic island-arc volcanism began in northern Chile (Ruiz and others, 1965). On the basis of current plate-tectonic models, and in light of rock-melting experiments by Green and Ringwood (1969) and Eggler and Burnham (1973), it is postulated that this volcanism was caused by partial melting of basaltic and/or peridotitic materials that were subducted beneath the western edge of South America as it moved away from Africa and overrode first the east-moving Farallon plate (150 to 130 m.y.), then the southeast-moving Phoenix plate (130 to 60 m.y.) (Larson and Pitman, 1972), and then the east-northeast-moving Southeast Pacific plate (60 m.y. to present) (Herron,
The andesitic and dioritic igneous activity that preceded and accompanied ore deposition during Neocomian time at El Romeral is viewed as a part of this subduction-related magmatism.

According to the following reasoning, the K$_2$O content of the Romeral igneous rock suite (K$_2$O = 0.6 percent at 55 percent of silica) suggests that the Romeral magmas were generated at depths of about 80 to 100 km. Benioff defined inclined seismic zones dipping beneath volcanic belts from nearby trenches and suggested that these seismic zones delineate tectonic dislocations that serve as preferred loci for production of magmas with heat generated by strain (Dickinson and Hatherton, 1967, p. 801). Dickinson and Hatherton (1967) noted that andesitic volcanoes never occur within the region of shallow seismicity (less than 80 km), and that K$_{55}$ contents of andesitic rock suites increase with increasing depths to subjacent seismic zones. They suggested that the linearity of volcanic chains erupting andesites reflects lines of intersection of Benioff inclined seismic zones and the subhorizontal Gutenberg low-velocity layer. They further suggested that "the partition of potash between incipient melt and residual crystalline phases should be such that the potash content of the melt rises as confining pressure rises" (Dickinson and Hatherton, 1967, p. 803).

The andesitic and dioritic rocks of El Romeral are within a linear belt of andesitic rocks that defines an island-arc system of late Jurassic to early Cretaceous age
(Ruiz and others, 1965). In accordance with the concept of uniformitarianism, the relationships described by Dickinson and Hatherton (1967) should be applicable to andesites of the Mesozoic island arc of Chile. Thus, according to plots of $K_{55}$ versus depth by Dickinson (1968), the value $K_{55} = 0.6$, characteristic of the Romeral andesite-diorite suite, suggests that the Romeral magmas were generated at depths of about 80 to 100 km.

Nielson and Stoiber (1973) criticized the use of $K_{55}$-versus-depth curves for determining depths of paleosubduction zones. Although they confirmed that $K_2O$ contents of andesitic suites generally increase with increasing depths to subjacent seismic zones, they found considerable differences, both in $K_2O$ levels, and in rates of $K_2O$ increase with depth, in rock suites from different island arcs. Regardless of these differences, however, the Dickinson-Hatherton observation that andesitic volcanoes never occur within less than 80 km of their subjacent Benioff zones probably gives a reliable minimum estimate for the depth of Romeral magma generation. Furthermore, if it is assumed that the Mesozoic Chilean arc was more similar to the modern Central American arc than to other circum-Pacific and oceanic arcs, then the Nielson-Stoiber plot of $K_{55}$ versus depth for the Central American arc can be applied. For an andesitic rock suite characterized by $K_{55} = 0.6$, this plot would suggest a subjacent seismic zone at 80 to 90 km depth. Thus it seems reasonable to postulate that the Romeral magmas were
generated by partial melting along an inclined seismic zone, well within the mantle, at depths between about 80 and 100 km.

Inasmuch as continental crust rarely is more than about 50 km thick, and inasmuch as relatively buoyant sialic crust probably is not easily subducted, it is doubtful that Pre-
cambrian iron formation was involved in the partial melting that generated the Romeral magmas. Thus iron probably was mobilized as a normal constituent of dioritic magmas formed by partial melting of subducted basalt and/or peridotite, at 80 to 100 km depth.

Stage 2 - Magmatic Ascent

As the dioritic Romeral magmas moved upward through subcrustal and continental crustal materials, they may have assimilated iron, perhaps from iron-rich metasedimentary rocks such as the iron-bearing quartzites that crop out near El Romeral, El Tofo, and Lago LieuLieu (in central Chile). However, if significant iron assimilation had taken place, one would expect La Liga andesites porphyry and Romeral diorite to be abnormally iron rich, and they are not. Average andesites of the Arqueros Fm contain 6.6 ± 2.4 percent of iron, and circumpacific andesites typically contain 3 to 8 percent of iron. In this context, neither La Liga andesite porphyry, with 4.1 ± 1.8 percent of Fe, nor Romeral diorite, with 3.4 ± 0.3 percent of iron, appears to be enriched in iron. Thus it is doubtful whether either of these large plutons, which are closely associated with the Romeral
magnetite deposits, assimilated much iron en route to their hypabyssal environment of crystallization.

**Stage 3 - Magmatic Differentiation**

The successive La Liga andesite and Romeral diorite plutons are progressively iron depleted with respect to Arqueros andesites. This suggests either that iron minerals formed and progressively settled out, or that they crystallized in progressively smaller amounts from successive plutonic magmas.

The post-diorite, pre-ore dikes exhibit two divergent compositional trends, one toward iron and magnesium enrichment along with silica depletion, and the other toward silica and alkali enrichment along with iron depletion. This suggests either that a post-diorite, pre-ore body of residual magma became increasingly differentiated, or that increasingly diverse magmas were generated in the mantle. The mineralogical similarities between the dikes and their close temporal and spacial relationships suggest differentiation from a common body of magma.

Intramineral dioritite magmas were enriched in iron, magnesium, and water. Enrichment in iron and magnesium might have been caused by resorption of iron-bearing oxide minerals and augite, but water probably was residually concentrated.

The extreme differences between iron-rich and iron-poor late dikes suggest hydrothermal removal of iron from the
upper parts of a subjacent residual magma body, along with simultaneous iron enrichment in the lower parts of the magma body.

Crystal settling and partial resorption: If early iron-bearing minerals had settled into deeper, hotter parts of the magma body and had been partly resorbed there, this would have caused progressive iron depletion in the upper parts of the magma body and progressive iron enrichment in its lower parts. This hypothesis is attractive in that it would account for the increasingly divergent compositional trends observed in the dike rocks of El Romeral. It also would explain the origin of an iron-enriched magma, from which iron-bearing hydrothermal fluids might have evolved, as well as the origin of a sodic-silicic magma, from which the hydrothermal fluids involved in quartzo-feldspathic alteration might have evolved.

However, there are several flaws in this hypothesis as applied to El Romeral. Plagioclase, not magnetite, probably was the earliest mineral to form from the Romeral magmas, and the next minerals to form were iron-poor augite and minor ilmenite. Settling of plagioclase or augite grains would not have carried iron downward, and although some unusually large ilmenite grains may have settled, there is no proof that such large grains existed. Moreover, if ilmenite grains had settled, the Romeral diorite would have been depleted in Ti as well as in iron, but it is not. In fact, both La Liga andesite and Romeral diorite contain more Ti than average
Arqueros andesites. Furthermore, this cannot be attributed to post-magmatic addition of Ti. Although both La Liga andesite and Romeral diorite contain deuteric sphene, the sphene almost invariably is in reaction rims around disseminated ilmenite grains. Thus the Ti in the sphene presumably came from the ilmenite grains.

It also is doubtful that magnetite grains sank. Some of the magnetite in Romeral diorite is intergrown with ilmenite in ilmenite-rich grains, but most is in dust-sized particles that are disseminated in interstitial actinolite, and some is in actinolite veinlets that cut the diorite. There are no known magnetite grains of magmatic origin that approach the 0.25 to 0.5 mm diameter that would have been necessary to cause settling of magnetite grains in a magma with a viscosity of $10^{4.5}$ poises or more (Table 14).

Furthermore, amphibole, which accounts for more than three times as much of the iron in Romeral diorite as ilmenite and magnetite together, almost certainly did not sink. Amphibole probably began to form at temperatures near 900°C, when the diorite was about 75 percent crystalline, and the viscosity of its interstitial residual magma was about $10^6$ poises or more (Fig. 77). With such a high proportion of crystals to impede sinking, and at such high magmatic viscosities, it is almost certain that amphibole did not sink. Textural relationships, moreover, indicate that amphibole partially replaced augite and plagioclase grains and filled the interstices between those grains. Thus, amphibole
probably acted as a cement, which bound the earlier crystals together and prevented them from sinking.

Still another flaw in the hypothesis of settling and partial resorption of iron-bearing mineral grains as a mechanism for producing iron-depleted and iron-enriched magmas, is that several factors would combine to limit the resorption of settling grains. According to Bowen (1928, p. 274-276), if crystal sinking is taking place in a body of magma, and the lower liquid is still somewhat above the temperature of saturation for the sinking mineral, some direct solution will occur. However, "as the concentration of the dissolving substance increases in the liquid, the temperature at which saturation will occur becomes higher. Moreover, in the act of solution, heat is absorbed, and the temperature of the liquid is lowered. Soon the temperature is reached at which saturation occurs for the new concentration, and simple solution ceases."

Finally, there are several ore-forming constituents, other than iron, that could not have been concentrated by crystal settling and partial resorption. For example, the density contrast between apatite grains and dioritic magma is too slight to allow settling of apatite grains of reasonable dimensions (Table 14). Furthermore, water, probably the most important iron-transporting agent, could not have been concentrated by settling and resorption.
Residual magmatic concentration of iron and other ore-forming constituents:

Progressive crystallization of early plagioclase, later augite, and minor ilmenite from Romeral diorite magma would have left the resulting residual magmas progressively enriched in iron, water, phosphorous, sulfur, copper, boron, and other ore-forming constituents.

Speed (1962) explained the concentration of iron in a scapolitized gabbroic complex in Nevada by suggesting that the iron-silicate solidus was depressed by high volatile content and that pO₂ stayed so low throughout the magmatic stage that little magnetite formed before the deuteric stage.

According to Osborn (1959, 1962) however, oxygen fugacities in water-rich, calc-alkaline magmas should be high-temperature dissociation of water. Further, according to Osborn's experiments, magnetite should crystallize at high temperatures from magmas with high oxygen fugacities. Thus, water-rich, calc-alkaline residual magmas should become progressively iron depleted, according to Osborn.

On the other hand, Eggler and Burnham (1973, p. 2528) have shown that "in a variety of geologically possible situations, O₂ in a magma within the crust is controlled largely by the initial FeO₁.₅/FeO or H₂O/H₂ ratio of the magma or by the O₂ buffering assemblage of the wallrocks." Moreover, "various geobarometers indicate that initial O₂ values in magmas and O₂ values in normal wallrocks are near NNO." Thus the high oxygen fugacities postulated by Osborn probably are not attained in natural magmas, and "neither
ilmenite nor magnetite is stable near the silicate liquidus” (Eggler and Burnham, 1973, p. 2524).

The textural characteristics of Romeral diorite suggest that the cores of plagioclase crystals were the earliest mineral grains to form, followed by plagioclase inner rims, augite, and minor ilmenite. Magnetite and amphibole apparently were late-magmatic to post-magmatic minerals. As shown in Figure 76, plagioclase probably began to form at 1175°C, whereas augite and ilmenite began to form at about 1075°C, and magnetite and amphibole began to form at still lower temperatures. Thus, with the progressive crystallization of anhydrous, non-ferrous plagioclase, the resulting residual magmas should have been progressively enriched in iron, water, and other ore-forming constituents.

In the subtraction diagram of Figure 80, it is assumed that the initial Romeral diorite magma had about the same composition as La Liga andesite, which contains about 5.4 percent of iron. If 80 percent of this magma had crystallized as Romeral diorite containing only 3.4 percent of iron, the remaining 20 percent of residual magma would have attained the composition of intramineral dioritite magma, with about 6 to 8 weight percent of iron and about 7 weight percent of water.

The next questions are; How large a body of intramineral dioritite magma would have been required to supply the iron that was added to the ores and actinolized rocks of El Romeral; and Is it geologically reasonable that a body of intramineral
Figure 80. Subtraction diagram showing how residual intramineral dioritite magma might have been derived by crystallization of Romeral diorite from an initial magma with the composition of La Liga andesite. About 80% of the initial magma would have had to crystallize as Romeral diorite to produce a residual magma with the composition of intramineral dioritite.
dioritite of that size existed? As shown in Table 16, approximately $1038 \times 10^5$ moles of iron were added to the rocks of the Romeral ore and actinolitic zones per vertical zone meter. Assuming that the vertical dimensions of those zones were about 500 m, then approximately $5.2 \times 10^{10}$ moles of Fe were added. Intramineral dioritite magma, containing 6.5 weight percent of iron and 7 weight percent of water, and with a density of 2.24 $\text{g cm}^{-3}$ would contain $10.9 \times 10^3$ moles of iron per cubic meter of magma. If iron were to be completely stripped from such a magma, it would require about $4.8 \times 10^6$ m$^3$ (or $4.8 \times 10^{-3}$ km$^3$) of the magma to supply as much iron as was added to the Romeral ores and actinolitic rocks. As suggested in Figure 80, the volume of the intramineral magma body was about 20 percent of the volume of the Romeral diorite pluton. Thus the horizontal area of the intramineral magma body may have been about 20 percent as large as that of the Romeral diorite pluton, or 20 percent of 11 km$^2$, or 2.2 km$^2$. Given this approximate size, only about the uppermost 2.2 meters of intramineral magma would have to be depleted in iron to supply all the iron that was added to the rocks of the Main and South Extension orebodies and their associated actinolitic zones. If only half of the magmatic iron were removed, the upper 4.4 meters of the intramineral magma would be required. If only half of the magmatic iron were removed from the upper magma in only one tenth of the 2.2 km area of intramineral magma, the upper 44 meters of magma would be required. In any case, the volumes of residual
magma required would be geologically reasonable.

Another question is: How was disequilibrium maintained between early-formed crystals and residual magmas, so that magmatic differentiation and residual enrichment of iron, water, and other ore-forming constituents could occur? According to Bowen's reaction principle, the minerals of the continuous and discontinuous reaction series tend to undergo changes in composition as the temperature drops, and they react with the residual magma. Thus, if equilibrium is maintained between the crystals and the residual magma, the resulting rock should have the same composition as the initial magma. However, if equilibrium is not maintained, then the composition of the residual magma should gradually change. The preservation of compositional zonation in the plagioclase crystals of Romeral diorite indicates that equilibrium was not maintained between such plagioclase crystals and their interstitial residual magmas. Moreover, if equilibrium was not maintained over small domains, it is almost certain that gross equilibrium was not attained throughout the crystallizing body of magma. Thus, magmatic differentiation could have occurred without physical separation of early-formed crystals from residual magmas by crystal settling. Ionic diffusion in the crystals and in the viscous residual magmas evidently was sufficiently slow that metastable disequilibrium was maintained until the late magmatic and post-magmatic stages, even though the tendency toward equilibrium would have become increasingly strong as the
compositional differences between the crystals and the residual magmas became increasingly extreme.

In the Romeral diorite, partial replacement of augite and plagioclase grains by amphibole and magnetite, probably resulted from a late approach toward equilibrium between early-formed crystals, interstitial residual magmas, and/or iron-bearing deuteritic fluids that separated from such magmas. Similarly, the replacement of andesitic and metasedimentary rocks by actinolite and magnetite to form the Romeral ore-bodies, probably resulted from an approach toward equilibrium between the wallrocks and iron-bearing hydrothermal fluids that emanated from subjacent, residual dioritic magmas.

The residual concentration hypothesis appears to offer the best explanation of how iron, water, and other ore-forming constituents were concentrated during the magmatic stage at El Romeral. One considerable advantage of the residual concentration hypothesis is that it can account for the concentration of magmatic water, magnesium, phosphorous, chlorine, sulfur, copper, and boron, as well as iron. All of these ore-forming constituents probably were present in the initial Romeral diorite magma, and they all probably were progressively enriched in residual magmas as a result of progressive crystallization of early plagioclase, later augite, and minor ilmenite.

Stage 4 - Separation of a Fluid Phase

According to Burnham (1967, p. 48), a water-rich fluid
phase will separate from a magma if the equilibrium pressure of water ($P_{H_2O}$) in the melt exceeds the total confining pressure ($P_t$). Furthermore, upon separation of an aqueous phase, the various constituents of the magma will tend to apportion themselves between the magma and the aqueous phase so as to equalize the chemical potential of each freely variable components in both phases. "Consequently, the partition coefficients of each component, and, hence the bulk composition of the aqueous phase is dependent upon temperature, pressure, bulk composition, and the nature of the coexisting phases" (Burnham, 1967, p. 49).

In the case of Romeral diorite, an aqueous fluid probably separated from interstitial residual magma when the diorite was largely crystalline. Iron, magnesium, and silica probably partitioned into this fluid, which reacted with magmatic plagioclase, augite, and probably hornblende, to form deuteric actinolite and magnetite.

In the case of the body of residual diorite magma that is inferred to have existed beneath El Romeral during ore deposition, water probably separated in response to pressure relief caused by strike-slip faulting in the overlying rocks. Residually concentrated magmatic iron probably partitioned favorably into the hot (about 1000°C), exsolving fluid, along with other residually concentrated ore-forming constituents such as Mg, P, Cl, S, Cu, and B.

According to Burnham (1967, p. 48), the rate of diffusion of dissolved magmatic water is very slow, but the
rate at which bubbles of a separate aqueous fluid would rise in a magma "appears to be sufficiently great that large scale transfer of water to the upper parts of a magma chamber could be readily effected in a short time." Thus, if iron-bearing fluids separated from residual dioritite magmas below El Romeral, those fluids would have migrated toward the top of their residual magma chamber and then would have escaped upward into fractures caused by faulting in the roof rocks.

Stage 5 - Ore-Forming Hydrothermal Processes

The final and most effective stage of iron concentration at El Romeral was the hydrothermal stage. During this stage, iron probably was taken from dioritite magma containing about 6.5 percent of iron, and it was deposited in the orebodies at concentrations ranging from 20 to 70 percent of iron. Thus, three- to ten-fold increases in iron concentration were achieved during the hydrothermal stage.

The ore-forming hydrothermal fluids of El Romeral probably separated from residual dioritite magma at about 1000°C and migrated upward into progressively cooler fractured roof rocks east of the Romeral fault (Fig. 82). At 550° and below, and at low f0₂ and low fS₂, magnesium, iron, and silicate ions in the fluid reacted with wallrock plagioclase to produce actinolite. This reaction consumed H⁺ ions, and, together with the predominant effect of decreasing temperature, promoted precipitation of magnetite and apatite.
Deposition of actinolite, magnetite, and apatite reduced concentrations of iron, magnesium, and phosphorous ions at sites of deposition. This established ionic concentration gradients and promoted ionic diffusion through the aqueous fluids from sites of solution, where the magma was exsolving water, to sites of deposition, where the orebodies were forming. There, chemical exchanges took place between the hydrothermal fluids and the wallrocks. As ore minerals replaced wallrock minerals, the replaced constituents were taken into solution and passed upward and outward by ordinary transport and by diffusion, probably toward cooler parts of the system, where they too probably were deposited in rocks that have now been eroded away.

With the passage of time, replacement fronts in the Main orebody moved northeastward, guided by northeast-trending shear zones related to left-lateral movement on the Romeral fault. Simultaneously, the hydrothermal fluids penetrated progressively smaller fractures throughout the ore zone, and wallrocks were progressively replaced along minor fractures, capillary fractures, mineral cleavages, grain boundaries, and pores. During later stages of mineralization, northwest-trending tensional gash fractures opened in response to more left-lateral movement on the Romeral fault. Many of these fractures were filled by intramineral dikes, and others served as channelways for hydrothermal fluids, such that northwest-trending high-grade zones were superimposed on earlier north- and northwest-
trending high-grade zones. Apatite, which is scarce in the early north-trending high-grade zones, was deposited in greater abundance in the later northwest-trending zones in the eastern part of the orebody.

During late stages of ore deposition, the temperature gradient began to decay, and replacement fronts in the Main orebody retreated inward, toward the central part of the orebody, and downward, toward its roots. For this reason, pyrite, a late mineral, is more abundant in the eastern and lower parts of the orebody than in its western and upper parts.

In the North orebody, fluid penetration probably was guided by differential permeability along foliation planes of phyllites and schists, and replacement reactions took place preferentially along biotite-rich layers, which were chemically more susceptible to replacement than feldspathic layers.

The important role ascribed to diffusive transmission of ions through a fluid network deserves to be emphasized, especially with regard to the mode of formation of the large bodies of metasomatic rocks at El Romeral. Such diffusion of ions through fluids would have allowed vastly greater ionic throughput than mass flow of fluid through microfractures and pores. On the other hand, mass fluid flow may have been more important than diffusion in the formation of the relatively open, late actinolite-magnetite-apatite veins that are scattered within a 3 to 4 km radius of El Romeral.
Stage 6 - Post-Ore Processes

**Quarzo-feldspathic alteration:** Quarzo-feldspathic alteration lagged behind actinolitization and Romeral ore deposition. This suggests that silica, aluminum, sodium, and potassium were soluble to lower temperatures than iron, magnesium, and phosphate. They were deposited later, at lower temperatures (about 475 ± 50°C), and at greater distances from a magmatic heat source subjacent to the Romeral magnetite deposits.

The final destination of the iron that was removed from the Romeral diorite during quartzo-feldspathic alteration is not known. However, such iron probably was not redeposited to form the Romeral magnetite deposits or the actinolitic rocks that surround them. Those rocks probably formed before most quartzo-feldspathic alteration occurred. Furthermore, at least eleven times as much iron probably was added to the rocks of the ore and actinolitic zones as was removed from the rocks of the quartzo-feldspathic zone.

**Iron-poor late dikes:** Tourmaline-bearing quartz-albite pegmatite dikes cut the Romeral orebodies, and this suggests that an iron-depleted body of magma, rich in water, silica, alumina, soda, and boron, existed below E1 Romeral after the ore was deposited. This further suggests that with progressive hydrothermal removal of iron from the upper parts of a
subjacent body of residual magma, the uppermost residual magma became depleted in iron and residually enriched in silica, alumina, soda, and boron. Such a magma may also have been the source of the silica-soda-bearing hydrothermal fluids that caused the aplitization of the Romeral diorite west of the orebodies as well as the feldspathization of the actinolized phyllites north and east of the orebodies (Fig. 83).

Iron-rich late dikes: Iron-rich biotite diorite dikes cut the Romeral orebodies and the aplitized rocks west of them. This suggests that iron-rich, water-rich residual magmas existed below El Romeral, even after the ore deposits and the aplitized rocks had formed. Such magmas probably existed in the deep portions of a subjacent residual-magma body. Furthermore, such magmas probably were the sources of the iron-bearing hydrothermal fluids from which the late magnetite-actinolite-apatite veins formed (Fig. 83).

Argillization and martitization: Argillization and martitization probably occurred during collapse of the Romeral hydrothermal system. The argillic assemblage suggests lower temperatures (350 ± 50°C), increased acidity, and increased pO₂, perhaps resulting from increasing admixture of oxygenated meteoric waters.

Weathering and erosion: Weathering of Romeral magnetite ore produced goethite along fractures. It is doubtful that much of the Romeral magnetite was removed by erosion in that
alluvial deposits near El Romeral only contain a few magnetite boulders, and those boulders are near their bedrock sources.

Summary of Magmatic and Hydrothermal Events and Processes

A graphic summary of the main igneous and hydrothermal events and processes of El Romeral is presented in a series of four diagrammatic, reconstructed sections (Figs. 81 through 84). Explanations of the events pictured in the sections are given in the accompanying captions.

Figure 81. Diagrammatic east-west section through El Romeral illustrating pre-ore igneous events, thermal metamorphism, contact metasomatism, and deuteric alteration. The Romeral diorite intruded metasedimentary rocks and La Liga andesite porphyry, both of which became dioritized near the contact. Hematite was hydrothermally removed from quartzites near the contact and redeposited in quartzites about 90 to 100 m from the contact. Thermal metamorphism resulted in the formation of diopside and oligoclase in the andesite, and of cordierite in phyllites and schists. Romeral diorite was deuteronically actinolitized and intruded by pre-ore dikes ranging in composition from iron-depleted dacite porphyry to iron-enriched porphyritic meladiorite. These dikes probably tapped different parts of a subjacent body of differentiated residual magma.
Figure 82. Diagrammatic east-west section through El Romeral illustrating ore deposition and ore-related igneous and hydrothermal events and processes. At this stage, iron-enriched residual magma probably became water-saturated. Aqueous fluid exsolved from the magma at about 1000°C, and iron partitioned into this fluid. Bubbles of the iron-bearing fluid rose through the magma and escaped into the roof rocks along the Romeral fault zone and its associated fractures. At 550 to 475°C, magnetite and actinolite were deposited from the fluid. This established concentration gradients within the fluid network and promoted ionic diffusion from sites of solution, where the magma was exsolving water, to sites of deposition in the growing orebodies. Actinolitic hornblende probably formed in rocks below the orebodies, and clinozoisite and chlorite probably formed in rocks along the outer and upper margins of the actinolitic zones. As ore deposition continued, tensional gash fractures opened in response to left-lateral movements on the Romeral fault. Intramineral dioritite magmas irrupted into such fractures and crystallized rapidly. The resulting intramineral dioritite dikes were then hydrothermally altered and locally replaced by magnetite.
Figure 83. Diagrammatic east-west section through El Romeral, illustrating late igneous and hydrothermal events and processes. At this stage the upper part of the inferred body of residual magma probably was hydrothermally depleted in mafic constituents and residually enriched in silica and soda, thus providing source magmas for the quartz-albite pegmatites that cut the orebodies, and for the silica-soda-bearing hydrothermal fluids that seeped upward and caused aplitization of Romeral diorite and feldspathization of actinolitized phyllites. At the same time, deeper magmas probably became increasingly enriched in mafic constituents and water, thus becoming source magmas for late, iron-rich, biotitic diorite dikes, and for the iron-bearing hydrothermal fluids from which the late actinolite-magnetite-apatite veins were deposited. In the orebodies, late actinolite, apatite, pyrite, sphene, clinozoisite, calcite, axinite, and chlorite were superimposed on earlier magnetite, actinolite, and scapolite. Meanwhile, clinozoisite, biotite, chlorite, and tourmaline formed in rocks peripheral to the orebodies and to the quartzo-feldspathic zones.
Quartz-feldspathic alteration is superimposed on dacitically and hydrothermally altered rocks.

Late quartz-albite pegmatite dikes cut the orebodies.

Late, silica-sodium-rich hydrothermal fluids

Late, iron-rich hydrothermal fluids

Iron-rich biotite diorite dikes cut the ore and quartz-feldspathic rocks.

Actinolite-magnetite-apatite veins cut quartz-feldspathic rock.

Actinolite-magnetite-apatite veins

Late quartz-albite pegmatite

o = oligoclase (An_{0-20})
g = quartz
m = microcline
c = chlorite
cs = clinopyroxene
ap = apatite (chlorapatite)
t = tourmaline
py = pyrite
sp = sphene
bi = biotite
ox = oxinitie
ac = actinolite
Figure 84. Diagrammatic east-west section through El Romeral illustrating post-ore structural, hydrothermal, and igneous events and processes. In the upper Arqueros formation, above El Romeral, the shift from gray-green andesites and marine limestones to red sandstones and purple andesites, suggests a shift from submarine volcanism and sedimentation under reducing conditions, to subaerial volcanism and deposition under oxidizing conditions. This shift may also be manifested in the martite and the advanced argillic minerals present in the upper parts of the Romeral orebodies. These minerals probably formed in response to oxidation and hydrogen-ion metasomatism that occurred at about 350 ± 50°C, as the Romeral thermal gradient decayed. The high oxidation potentials and strong acidities required for such alteration suggest deep convection of oxygenated meteoric waters, and mixing of such waters with late, \( H_2S \)-bearing magmatic emanations to form \( H_2SO_4 \).

After Romeral ore deposition and alteration, the Punta de Piedra granodiorite batholith was emplaced east of El Romeral. Epidote, pyrite, chalcopyrite, and montmorillonite formed adjacent to and within this batholith during deuteric and hydrothermal alteration, which closely followed its crystallization.
Summary of the Romeral Genetic Model

Stage 1 - partial melting: Iron was mobilized as a constituent of dioritic magmas generated at about 80 to 100 km depth by partial melting of basaltic and/or peridotitic materials that were subducted beneath the western edge of the South American plate.

Stage 2 - magmatic ascent: These dioritic magmas, being less dense than surrounding and overlying rocks, moved upward through the mantle and crust. However, they probably did not assimilate much iron from the rocks they traversed, for they are not iron-enriched with respect to average andesites.

Stage 3 - magmatic differentiation: As diorite crystallized (at about 7 ± 3.5 km depth) progressive crystallization of early plagioclase, later augite and minor ilmenite, resulted in residual concentration of water, iron, magnesium, phosphorous, chlorine, sulfur, copper, and boron in residual magmas represented by intramineral dioritite dikes.

Stage 4 - separation of an aqueous fluid phase: Residual water concentration, upward intrusion, and/or pressure release resulting from faulting and fracturing within the Romeral fault system caused water to separate from residual dioritite magma. At about 1000°C, iron, magnesium, phosphorous, and chlorine, along with minor boron, sulfur, copper, and carbon dioxide, partitioned into the aqueous phase.
Stage 5 - ore-forming hydrothermal processes: By upward movement of aqueous fluid through magma and through wallrock fractures and pores, a network of fluid was established through a vertical thermal gradient of about 500° in 2000 m. At 550°C and below, and at low fO₂ and low fS², magnesium, iron and silicate ions in the fluid reacted with wallrock plagioclase to produce actinolite. This reaction consumed H⁺ ions, and together with the predominant effect of decreasing temperature, promoted deposition of magnetite and apatite. Formation of actinolite, magnetite, and apatite lowered the concentrations of iron, magnesium, and phosphorous ions in hydrothermal fluids at sites of deposition. This established ionic concentration gradients within the fluid network, and promoted ionic diffusion through the aqueous fluids from sites of solution, where the magma was exsolving water, to sites of deposition within the growing orebody.

Stage 6 - Post-ore processes: Late ore deposition was accompanied and followed by quartzo-feldspathic alteration and followed by emplacement of contrasting sets of mafic and sodic-silicic dikes. This suggests that with progressive hydrothermal removal of iron, the upper parts of a subjacent body of residual magma became depleted in iron and residually enriched in alkalies and silica, whereas the lower parts of the residual magma body continued to become progressively enriched in iron.
Thermal gradients decayed with time as the subjacent magma crystallized and cooled. Argillic alteration probably occurred during collapse of the Romeral hydrothermal system. Dickite, kaolinite, pyrophyllite, and sericite probably developed in response to lower temperatures (near 350 ± 50°C) and increasing hydrothermal acidity, which may have resulted from increased admixture of oxygenated groundwaters with late, H₂S-bearing hydrothermal fluids of magmatic origin.
APPLICATIONS

Applications in Exploration

"My basic premise is that intelligent search for additional reserves must be based on a thorough knowledge of known deposits."

S. R. Wallace, 1974, p. 217

Much geologic exploration is guided by logical analogy between little-known prospects and well-known orebodies. Therefore, the better the known orebodies are understood, the more intelligent the search for new ones can become. However, since no two orebodies are the same, a geologic model should be genetic as well as descriptive. "It is only by understanding the origin that the geologist is able to interpret the differences between orebodies, and thereby intelligently apply a model in varied geologic settings" (S. R. Wallace, 1974, p. 225).

Although the foregoing descriptive and genetic model of El Romeral is far from complete, and although it probably contains errors of many kinds, hopefully it also contains descriptive information and conceptual tools that will be useful in exploration for other magnetite deposits of a generally similar nature.

Implications regarding other deposits: Many magnetite-actinolite deposits visited in Chile and in Nevada have much in common with El Romeral. They are adjacent to or near
plutons of intermediate composition (most commonly diorite), they have similar assemblages of ore and gangue minerals, and they evidently formed by fracture filling and wallrock replacement along fault zones and/or igneous contacts. The magmatic stage of iron concentration may have been more important in some than in others, and the hydrothermal stage in some may have overlapped with a stage of deuteric release of iron. Magnetite-garnet-pyroxene skarns in carbonate rocks probably form by a somewhat different process and commonly are related to more granodioritic and granitic plutons.

**Implications regarding metallogenic provinces:** Iron-bearing igneous rocks of intermediate composition are common around the Pacific Basin. They probably formed by partial melting of subducted basaltic and/or peridotitic materials, and they probably are the starting materials from which many metasomatic iron deposits formed. Groupings of iron deposits define metallogenic provinces such as the Chilean iron province, the California-Nevada iron province, and the Philippine iron province. Such provinces may represent areas in which conditions were favorable for magmatic concentration of iron and water, followed by fracturing, separation of water, and establishment of hydrothermal networks through steep temperature gradients. Such provinces may therefore represent areas that underwent similar processes, rather than areas that were underlain by particularly iron-rich rocks, either in their basement complexes, or in the
lower crust, or upper mantle.

**Specific prospecting suggestions:** Contact metasomatic magnetite deposits similar to El Romeral are likely to be found in belts of hypabysal intrusions of intermediate composition. Such belts are likely to be found parallel to active or dormant subduction zones, and they are likely to be nearer to their respective plate margins than are belts of copper and molybdenum deposits (Sillitoe, 1972). However, this is definitely not to say that magnetite deposits are found only in such environments.

Much emphasis has been placed on the importance of carbonate rocks as the most chemically favorable host rocks for replacement-type deposits, and on the importance of argillic alteration as a guide to porphyry-copper type mineralization. However, El Romeral and many other magnetite-amphibole deposits, some with abundant copper, serve to demonstrate that andesitic volcanic rocks can host replacement-type deposits, and that actinolitic and quartz-feldspathic rocks may serve as guides to ore. To the extent that magmatic iron retention was important in the Romeral process, iron-deficient diorites may be favorable indications of ore. To the extent that amphibole- and biotite-rich dioritite dikes might be exposed, they too would be favorable, as would dike sets of extremely divergent character that might indicate extreme magmatic differentiation. To the extent that actinolite-apatite-magnetite veins are scattered about massive
deposits like El Romeral, they too may be important guides to ore - in fact they may themselves be ore.

Applications to Mining

Several factors influence the quality and value of magnetite ore and concentrates. Among these are its iron content, its content of deleterious constituents such as phosphorous, titanium, and sulfur, and its particle size.

Continued mapping of intramineral and late dikes, together with relogging of core to locate, classify, and intercorrelate the dikes would aid in accurate prediction of ore grade. Because phosphorous, titanium, and sulfur are present in non-magnetic apatite, sphene, and pyrite, they can be magnetically separated from the ore if they are liberated by crushing. However, impurity content must be balanced against particle size to maximize ore value. Therefore, in addition to knowing grade distributions for iron and deleterious elements, it would be useful to learn more about the spacial and textural relationships between ore minerals, deleterious gangue minerals, and post-ore fracture zones. This type of information, together with studies of beneficiation characteristics of various types recognized, could be helpful in planning future mining, milling, and marketing procedures. A pilot study of this type by Antonio Pena was inconclusive, perhaps because it focused only on thin-section study rather than on combined
field, hand-sample, core-sample, and thin-section studies.

Continued mapping of faults, fractures, andesite-phyllite contacts, and argillic alteration zones could aid in prediction of slope stabilities and in future pit design.
SUGGESTIONS FOR FURTHER RESEARCH

The Paleozoic-Jurassic unconformity near La Escoba should be examined in greater detail and fully mapped, as should the west side of Cordón La Liga and the contact zone of the Juan Soldado batholith.

Jurassic volcanic sections should be measured north of El Romeral, and detailed comparisons should be made with the La Escoba andesites.

Detailed pit mapping and ore zoning should be continually updated as new exposures and new drill holes become available.

Relative ages suggested in this report should be checked and refined by continued mapping and by radiometric age dating methods if possible.

Actinolite from ore zones, alteration zones, and actinolite veins could be chemically analyzed and compared. Aluminum content of the actinolite should vary with temperatures of formation. Such analyses would provide checks and refinements of the temperature estimates given herein.

Trace-element contents of Romeral chlorapatite samples could be determined and compared to those of apatite-bearing deposits such as Kiruna, Iron Springs, and Lake LieuLieu.

Strontium-isotope work on Romeral diorite and intramineral dioritites might further indicate whether their source materials were crustal, or sub-crustal, or hybrid.

Oxygen-isotope work on primary igneous minerals might indicate how much groundwater diffused into and/or reacted with Romeral magmas and Romeral igneous rocks. Hydrogen-
oxygen isotope work on OH-bearing secondary minerals and on fluid inclusions might suggest the relative importances of magmatic, connate, and meteoric waters in the ore-bearing hydrothermal fluids.

Detailed work on the compositions of fluids trapped in fluid inclusions and on the compositions of their daughter minerals might further elucidate the chemical compositions of the ore-bearing fluids.

Additional experiments with fluid-phase transport of magnetite and apatite might simulate Romeral ore transport and deposition. Such experiments should be run with at least 7 weight percent of HCl-bearing water, and with temperatures near 1000°C at the hot end, and 500°C at the cool end of the thermal gradient.
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REFERENCES


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APPENDIX A
TABLES OF MINERALOGICAL COMPOSITIONS,
CHEMICAL COMPOSITIONS,
AND TEXTURAL CHARACTERISTICS
OF ROCKS
FROM EL ROMERAL
### TABLE A-1

**Schist**

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**Mineralogical Composition**

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### Mineralogical Composition

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La Escoba Andesite
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* See Appendix C for optical data and approximate chemical compositions of minerals.
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<td>random (0-50%)</td>
<td>replaces plagioclase and may replace biotite.</td>
<td>early magmatic</td>
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<tr>
<td>(groundmass)</td>
<td>.01-.08</td>
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<td>matrix, no preferred orientation.</td>
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<tr>
<td>biotite</td>
<td>.01-.03</td>
<td>subhedral</td>
<td>mostly in groundmass, but also in plagioclase phenocrysts and in clusters to .5 mm.</td>
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<td>post-magmatic</td>
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<td>actinolite</td>
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<td>clusters of needles, &amp; poikiloblastic knots.</td>
<td>random.</td>
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<td>along octahedral planes of magnetite</td>
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<td>equant grains</td>
<td>most abundant in and near veinlets</td>
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<td>hydrothermal</td>
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<tr>
<td>scapolite</td>
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<td>anhedral</td>
<td>most abundant near magnetite-actinolite-apatite veins.</td>
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<td>(marialitic)</td>
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<td>random</td>
<td>post-magmatic(?)</td>
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<td>Wt. %</td>
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<td>99.90</td>
<td>98.65*</td>
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* (Oyarzun and Villalobos, 1969).
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<th>Vol. % An</th>
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<td>(16.20) 27±2</td>
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<td>(outer rim)</td>
<td>(9.22) 21±3</td>
<td>(8.70) 19±1</td>
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<td>29.7</td>
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</tr>
<tr>
<td>apatite</td>
<td>tr</td>
<td>--</td>
</tr>
<tr>
<td>sphene</td>
<td>--</td>
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<td>--</td>
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<tr>
<td>chlorite</td>
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100.0 100.1
### TABLE A-7
Cerro Juan Soldado Quartz Monzonite
Textural Characteristics

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<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape and Character</th>
<th>Distribution</th>
<th>Interpretations: Processes, Reactions</th>
<th>Stage</th>
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<td>plagioclase</td>
<td>1-2</td>
<td>subhedral laths with oscillatory zoning (3-15 zones).</td>
<td>random, framework</td>
<td></td>
<td>early to late magmatic</td>
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<td>microcline</td>
<td>.3-1.5</td>
<td>subhedral to anhedral, micropertithic</td>
<td>interstitial</td>
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<td>mid to late magmatic</td>
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<td>anhedral</td>
<td>interstitial</td>
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<td>biotite 'c'</td>
<td>.1- .5</td>
<td>subhedral</td>
<td>-interstitial</td>
<td>-replaces hornblende</td>
<td>-late magmatic</td>
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<td>.01-.05</td>
<td>anhedral</td>
<td>-intergrown with hornblende</td>
<td>replacements plagioclase</td>
<td>-post-magmatic</td>
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<td>hornblende 'c'</td>
<td>.4-.7</td>
<td>subhedral prisms</td>
<td>poikiloblastic, with feldspar inclusions.</td>
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<td>late magmatic</td>
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<tr>
<td>(ferrohornblende)</td>
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<td>magnetite</td>
<td>.02-.2</td>
<td>subhedral to anhedral</td>
<td>interstitial and within grains of feldspar, quartz, biotite, and hornblende.</td>
<td>-appears to replace feldspar, quartz, biotite, and hornblende</td>
<td>-late magmatic to post-magmatic</td>
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<td>.05-.4</td>
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<tr>
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<td>thin lamellae</td>
<td>along (0001) planes in cores of ilmenite grains - no hematite in rims.</td>
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<td>anhedral</td>
<td>interstitial</td>
<td></td>
<td>late magmatic</td>
</tr>
<tr>
<td>zircon</td>
<td>.02</td>
<td>equant, rounded</td>
<td>inclusion in biotite, with pleochroic halo</td>
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<td>magmatic</td>
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<td>.05</td>
<td>equant</td>
<td>random</td>
<td></td>
<td>(?)</td>
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<td>sphen</td>
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<td>anhedral</td>
<td>intergrown with biotite</td>
<td>replaces biotite</td>
<td>post-magmatic</td>
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<td>chlorite</td>
<td>.01- .2</td>
<td>sheaves of plates</td>
<td>intergrown with biotite</td>
<td>replaces biotite</td>
<td>post-magmatic</td>
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<tr>
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<td>.1</td>
<td>anhedral</td>
<td>intergrown with plagioclase</td>
<td>replaces plagioclase</td>
<td>post-magmatic</td>
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### TABLE A-8
Cerro Juan Soldado Quartz Monzonites
Chemical Compositions

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<th>Marginal Facies, #1616</th>
<th>Central Facies, #1618</th>
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<td>SiO₂</td>
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<td>TiO₂</td>
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<td>14.78</td>
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<td>Fe₂O₃</td>
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<tr>
<td>MnO</td>
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<td>MgO</td>
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<tr>
<td>Cl</td>
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<td>CO₂</td>
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<tr>
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## TABLE A-9

### East Soldado Trondjemite

Mineralogical Composition and Calculated Chemical Composition

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### Mineralogical Composition

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### Calculated Chemical Composition

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\[ \text{H}_2\text{O}^+ \quad 0.008 \]

\[ 99.96 \]
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<td>23.22</td>
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<td>11.13</td>
<td>24.00</td>
<td>--</td>
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<td>0.96</td>
<td>0.41</td>
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<td>1.08</td>
<td>0.04</td>
<td>0.37</td>
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<td>0.31</td>
<td>0.01</td>
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<td>0-.31</td>
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<td>0.22</td>
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<td>0.53</td>
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<td></td>
<td>0.11</td>
<td>0-.53</td>
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<td>rutile</td>
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<td>--</td>
<td>--</td>
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<td>tr</td>
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<td>99.73</td>
<td>99.88</td>
<td>99.99</td>
<td>99.68</td>
<td>100.03</td>
<td>100.0</td>
<td>99.63</td>
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### TABLE A-12

La Liga Intrusive Andesite Porphyry and Andesites of the Neocomian Arqueros Formation

Mean Chemical Compositions

Mean of 7 calculated chemical compositions, La Liga andesite porphyry.

<table>
<thead>
<tr>
<th>Element</th>
<th>Mean Wt.%</th>
<th>S.D.</th>
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<tr>
<td>SiO₂</td>
<td>54.50 ±</td>
<td>1.93</td>
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<td>TiO₂</td>
<td>1.57 ±</td>
<td>0.92</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.2 ±</td>
<td>2.60</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.54 ±</td>
<td>0.93</td>
</tr>
<tr>
<td>FeO</td>
<td>3.83 ±</td>
<td>1.46</td>
</tr>
<tr>
<td>MnO</td>
<td>0.03 ±</td>
<td>0.03</td>
</tr>
<tr>
<td>MgO</td>
<td>5.52 ±</td>
<td>2.18</td>
</tr>
<tr>
<td>CaO</td>
<td>10.52 ±</td>
<td>1.15</td>
</tr>
<tr>
<td>Na₂O</td>
<td>4.07 ±</td>
<td>0.84</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.37 ±</td>
<td>0.08</td>
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<tr>
<td>P₂O₅</td>
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</tr>
<tr>
<td>Cl</td>
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<tr>
<td>CO₂</td>
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<tr>
<td>F</td>
<td>0.02</td>
<td></td>
</tr>
<tr>
<td>V₂O₃</td>
<td>0.001 ±</td>
<td>0.001</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.89 ±</td>
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<tr>
<td>H₂O⁻</td>
<td>0.06 ±</td>
<td>0.02</td>
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Sum: 100.18

Mean of 18 published chemical analyses (Oyarzun and Villalobos, 1969) Arqueros Fm. meta-andesites, Coquimbo Province.

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<th>Element</th>
<th>Mean Wt.%</th>
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<tbody>
<tr>
<td>SiO₂</td>
<td>51.47 ±</td>
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</tr>
<tr>
<td>TiO₂</td>
<td>1.06 ±</td>
<td>0.39</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.98 ±</td>
<td>1.87</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>4.75 ±</td>
<td>1.65</td>
</tr>
<tr>
<td>FeO</td>
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<td>MnO</td>
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<td>0.07</td>
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<tr>
<td>MgO</td>
<td>5.08 ±</td>
<td>2.53</td>
</tr>
<tr>
<td>CaO</td>
<td>7.58 ±</td>
<td>2.93</td>
</tr>
<tr>
<td>Na₂O</td>
<td>5.34 ±</td>
<td>1.88</td>
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<tr>
<td>K₂O</td>
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<td>0.90</td>
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<tr>
<td>P₂O₅</td>
<td>0.30 ±</td>
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Sum: 100.08

* lbc: lost by calcination
### TABLE A-13
La Liga Intrusive Andesite Porphyry
Textural Characteristics

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size(mm)</th>
<th>Shape and Character</th>
<th>Distribution</th>
<th>Processes, Reactions</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase phenocrysts</td>
<td>.8-3</td>
<td>subhedral laths, oscillatory zoning, 2-10 zones</td>
<td>random</td>
<td>partly replaced by actinolite</td>
<td>early magmatic</td>
</tr>
<tr>
<td>(11.6 mean vol.%)</td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>plagioclase groundmass</td>
<td>.1-.8</td>
<td>subhedral laths and irregular remnants</td>
<td>random</td>
<td>partly replaced by actinolite</td>
<td>late magmatic</td>
</tr>
<tr>
<td>(43.7 mean vol.%)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>augite</td>
<td>.05-1.5</td>
<td>subhedral phenocrysts, anhedral in groundmass</td>
<td>-random, -interstitial</td>
<td>partly replaced by actinolite</td>
<td>middle to late magmatic</td>
</tr>
<tr>
<td>actinolitic hornblende</td>
<td>.05-.2</td>
<td>anhedral</td>
<td>-interstitial, -intergrown with augite</td>
<td>replaces augite</td>
<td>late magmatic</td>
</tr>
<tr>
<td>quartz</td>
<td>.05-.2</td>
<td>anhedral</td>
<td>-intergrown with augite, hornblende, and plagioclase</td>
<td>product of reaction: late plagioclase + augite + water → hornblende + quartz</td>
<td>late magmatic</td>
</tr>
<tr>
<td>ilmenite</td>
<td>.05</td>
<td>amoeboid</td>
<td>random, interstitial</td>
<td>later than plag., augite, and some quartz and hornblende.</td>
<td>middle to late magmatic</td>
</tr>
<tr>
<td>Mineral</td>
<td>Size(mm)</td>
<td>Shape and Character</td>
<td>Distribution</td>
<td>Processes,Reactions</td>
<td>Stage</td>
</tr>
<tr>
<td>--------------</td>
<td>----------</td>
<td>---------------------</td>
<td>--------------</td>
<td>---------------------------------------------------------</td>
<td>-------------</td>
</tr>
<tr>
<td>actinolite</td>
<td>.05-.4</td>
<td>-prisms, -bunches of needles, -poikiloblastic clots.</td>
<td>-interstitial; -intergrown with augite, plagioclase; and opaques; in veinlets.</td>
<td>replaces augite, plagioclase, and hornblende.</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>sphene</td>
<td>.05-.3</td>
<td>-anhedral</td>
<td>-rims on ilmenite, -intergrown with ilmenite, -randomly distrib., -in and near veinlets.</td>
<td>-product of reaction between ilmenite and plagioclase; -introduced</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>magnetite</td>
<td>.02-.05</td>
<td>-equant, rounded -thin &quot;blades&quot; -sub-octahedral</td>
<td>-inclusions in augite and amphiboles -along augite and amphibole cleavages -in veinlets, with actinolite</td>
<td>replaces augite and amphiboles</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>.05</td>
<td>equant grains</td>
<td>in veinlets</td>
<td>replaces plagioclase and augite</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>.1-.1.0</td>
<td>-irregular shreds -groups of equant grains</td>
<td>-intergrown with plagioclase and augite.</td>
<td>replaces plagioclase and augite</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>penninite</td>
<td>.05-.4</td>
<td>-sheaves of plates</td>
<td>-intergrown with actinolite</td>
<td>replaces actinolite</td>
<td>post-magmatic</td>
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<tr>
<td>hematite</td>
<td>.00X</td>
<td>-thin plates</td>
<td>-on (0001), ilmenite on octahedral planes, magnetite</td>
<td>-exsolved from ilmenite -replaces magnetite (by oxidation)</td>
<td>late magmatic post-magmatic</td>
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Table A-14
Augite Microdiorite
Mineralogical Composition

<table>
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<tr>
<th>Sample #</th>
<th>Coordinates:</th>
<th>Elevation:</th>
<th>Relative Unaltered</th>
<th>Moderately Actinolitized</th>
</tr>
</thead>
<tbody>
<tr>
<td>1160</td>
<td>13435N 14940E</td>
<td>770m</td>
<td>Augite microdiorite</td>
<td>Augite microdiorite</td>
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<tr>
<td>1336</td>
<td>14130N 4465E</td>
<td>910m</td>
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<table>
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<th>Vol. %</th>
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<td>plagioclase</td>
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<tr>
<td>phenocrysts (about An31)</td>
<td>0.74</td>
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<td>groundmass (about An31)</td>
<td>48.76</td>
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<tr>
<td>actinolite</td>
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<td>magnetite</td>
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<td>clinozoisite</td>
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<td>1.06</td>
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<td>99.96</td>
<td>99.62</td>
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<tr>
<td>Mineral</td>
<td>Size (mm)</td>
<td>Shape, Character</td>
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<td>-----------</td>
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<td>.75-2.0</td>
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<td>subhedral laths</td>
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<td>augite</td>
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<td>-anhedral grains</td>
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<td>-equant grains</td>
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<tr>
<td>hornblende 'a'</td>
<td>.02-.2</td>
<td>-anhedral</td>
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<td>(actinolitic)</td>
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<td>-irregularly shaped groups</td>
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<td>of very fine prisms.</td>
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<td>anhedral</td>
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<td>magnetite</td>
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<td>equant</td>
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<tr>
<td>sphen</td>
<td>.02-.2</td>
<td>anhedral</td>
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<tr>
<td>clinoholite</td>
<td>.1-.3</td>
<td>groups of equant grains</td>
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<td>chlorite</td>
<td>.1-.3</td>
<td>sheaves of platelets</td>
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### TABLE A-16

**Augite Microdiorite**

Calculated Chemical Compositions

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<th>Sample #1160</th>
<th>Sample #1336</th>
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<td>relatively unaltered augite microdiorite</td>
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<tr>
<td><strong>SiO₂</strong></td>
<td>53.42</td>
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<td><strong>TiO₂</strong></td>
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<td><strong>Al₂O₃</strong></td>
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<td><strong>Fe₂O₃</strong></td>
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<td><strong>FeO</strong></td>
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<td><strong>MnO</strong></td>
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</tr>
<tr>
<td><strong>MgO</strong></td>
<td>8.13</td>
</tr>
<tr>
<td><strong>CaO</strong></td>
<td>13.68</td>
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<tr>
<td><strong>Na₂O</strong></td>
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</tr>
<tr>
<td><strong>K₂O</strong></td>
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<tr>
<td><strong>F</strong></td>
<td>0.003</td>
</tr>
<tr>
<td><strong>H₂O⁺</strong></td>
<td>0.92</td>
</tr>
<tr>
<td><strong>H₂O⁻</strong></td>
<td>0.08</td>
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<tr>
<td><strong>Total</strong></td>
<td>100.33</td>
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TABLE A-17
Ocoite
Mineralogical and Chemical Compositions

| Sample #: | 1503 |
| Coordinates: | 11210N 4320E |
| Elevation | 705m |

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<th>Vol.%</th>
<th>Constituent</th>
<th>Wt. %</th>
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<td>52.64</td>
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<td>18.55</td>
<td>TiO$_2$</td>
<td>0.91</td>
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<tr>
<td>(core, An53)</td>
<td>(13.15)</td>
<td>Al$_2$O$_3$</td>
<td>16.36</td>
</tr>
<tr>
<td>(rim, An38)</td>
<td>(5.40)</td>
<td>Fe$_2$O$_3$</td>
<td>0.82</td>
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<tr>
<td>groundmass</td>
<td>37.65</td>
<td>FeO</td>
<td>4.19</td>
</tr>
<tr>
<td>(An42)</td>
<td></td>
<td>MnO</td>
<td>0.13</td>
</tr>
<tr>
<td>Total plagioclase</td>
<td>56.20</td>
<td>MgO</td>
<td>6.98</td>
</tr>
<tr>
<td>augite (phenocrysts)</td>
<td>3.66</td>
<td>CaO</td>
<td>13.72</td>
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<td>c) in veinlets</td>
<td>c) in veinlets</td>
<td>c) added</td>
<td>post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>augite</td>
<td>.2-1.0</td>
<td>a) stubby prisms</td>
<td>a) random</td>
<td>a-b) eutectic crystallization</td>
<td>late magmatic</td>
<td>magmatic</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>b) anhedral networks</td>
<td>b) interstitial</td>
<td>b) plagioclase</td>
<td>post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>c) in veinlets</td>
<td>c) in actinolite</td>
<td>c) replaced by actinolite</td>
<td>post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>sphene</td>
<td>.01-.05</td>
<td>a) anhedral intergrowths with ilmenite</td>
<td>a-b) reaction product of ilmenite and plagioclase</td>
<td>late magmatic to sphene</td>
<td>late magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>b) rims between ilmenite and plagioclase</td>
<td>ilmenite and plagioclase</td>
<td></td>
<td>late magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>c) random, and in diamond-shaped grains</td>
<td>c) added</td>
<td></td>
<td>post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>quartz</td>
<td>.02-.07</td>
<td>anhedral</td>
<td>a) reaction product(?)</td>
<td>late to post-magmatic</td>
<td>late magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>b) random, near contacts with quartzite inclusions</td>
<td>partly digested quartzite inclusions</td>
<td></td>
<td>pre-magmatic</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>chlorapatite</td>
<td>.02</td>
<td>equant, rounded</td>
<td>random</td>
<td></td>
<td>post-magmatic(?)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mineral</td>
<td>Size(mm)</td>
<td>Shape and Character</td>
<td>Distribution</td>
<td>Processes, Reactions</td>
<td>Stage</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-----------------</td>
<td>----------</td>
<td>-----------------------------------------</td>
<td>--------------------------------------------------</td>
<td>----------------------------------------</td>
<td>------------------</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>biotite 'b'</td>
<td>.01-.03</td>
<td>small books, and anhedral clots</td>
<td>scattered through feldspars and assoc. with scapolite and actinolite</td>
<td>replaces plagioclase and actinolite</td>
<td>late post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>penninite</td>
<td>.1</td>
<td>sheaves of platelets</td>
<td>with actinolite</td>
<td>replaces actinolite</td>
<td>late post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>tourmaline</td>
<td>.05</td>
<td>irregularly shaped clots of prisms</td>
<td>in plagioclase, and associated with actinolite</td>
<td>replaces plagioclase and actinolite</td>
<td>late post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ilmenite</td>
<td>.01-.25</td>
<td>very irregularly shaped grains</td>
<td>randomly disseminated (interstitial)</td>
<td>replaces actinolite</td>
<td>late post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>magnetite</td>
<td>.01-.05</td>
<td>equant grains</td>
<td>mostly in actinolite (actinolite appears to be the most favorable host mineral for magnetite deposition)</td>
<td>replaces actinolite</td>
<td>post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>hematite</td>
<td>.00X-.25</td>
<td>a) thin lamellae</td>
<td>a) on(0001) ilmenite</td>
<td>a) exsolution</td>
<td>late post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>b) thicker lamellae</td>
<td>b) on octahedral planes</td>
<td>b) oxidation</td>
<td>post-magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>c) anhedral grains</td>
<td>c) in magnetite</td>
<td>c) oxidation</td>
<td>weathering</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>rutile</td>
<td>.00OX-.05</td>
<td>a) needles (rare)</td>
<td>a) inclusions(?) in plagioclase and quartz</td>
<td>a) exsolution</td>
<td>magmatic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>b) knee-shaped twins (very rare)</td>
<td></td>
<td></td>
<td>to</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

324
### TABLE A-22

Leuco-andesite Porphyry Dike

Mineralogical and Chemical Composition, and Textural Characteristics

| Sample #: | 179 |
| Location: | near east contact of East Mass, Romeral Diorite |
| Coordinates: | 11350N 7650E |
| Elevation: | 250m |
| Trend: | N 80 W, 90° |
| Width: | 5m |

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol.%</th>
<th>Size(mm)</th>
<th>Distribution</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase phenocrysts</td>
<td>13.75</td>
<td>2-4</td>
<td>laths parallel to dike walls</td>
<td></td>
</tr>
<tr>
<td>plagioclase groundmass</td>
<td>67.13</td>
<td>.02-.05</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total plagioclase (about An 37)</td>
<td>80.88</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>hornblende 'a'</td>
<td>13.65</td>
<td>.02-.4</td>
<td>disseminated</td>
<td></td>
</tr>
<tr>
<td>magnetite-hematite</td>
<td>0.55</td>
<td>.02-.05</td>
<td>disseminated</td>
<td></td>
</tr>
<tr>
<td>clinzoisite</td>
<td>3.06</td>
<td>.1</td>
<td>dissem. &amp; vnlt.</td>
<td></td>
</tr>
<tr>
<td>sphene</td>
<td>0.55</td>
<td>.1</td>
<td>disseminated</td>
<td></td>
</tr>
<tr>
<td>chlorapatite</td>
<td>1.09</td>
<td>.05</td>
<td>vnlt.</td>
<td></td>
</tr>
</tbody>
</table>

Calculated Chemical Composition

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>55.04</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.58</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>22.04</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.99</td>
</tr>
<tr>
<td>FeO</td>
<td>1.53</td>
</tr>
<tr>
<td>MnO</td>
<td>0.02</td>
</tr>
<tr>
<td>MgO</td>
<td>2.22</td>
</tr>
<tr>
<td>CaO</td>
<td>9.49</td>
</tr>
<tr>
<td>Na₂O</td>
<td>5.58</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.54</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.48</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.04</td>
</tr>
<tr>
<td>Cl</td>
<td>0.08</td>
</tr>
<tr>
<td>F</td>
<td>0.04</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.28</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Total: 99.78
TABLE A-23
Aplitic Quartz Diorite Dike

Sample #: 170
Coordinates: 11640N 7325E
Elevation: 325m

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol.%</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td></td>
<td></td>
</tr>
<tr>
<td>core</td>
<td></td>
<td></td>
</tr>
<tr>
<td>An 28 ± 1</td>
<td>25.36</td>
<td></td>
</tr>
<tr>
<td>rim</td>
<td></td>
<td></td>
</tr>
<tr>
<td>An 24 ± 1</td>
<td>10.27</td>
<td></td>
</tr>
<tr>
<td>unzoned</td>
<td></td>
<td></td>
</tr>
<tr>
<td>An 27 ± 1</td>
<td>28.98</td>
<td></td>
</tr>
<tr>
<td>Total plagioclase</td>
<td>64.61</td>
<td></td>
</tr>
<tr>
<td>quartz</td>
<td>9.60</td>
<td></td>
</tr>
<tr>
<td>actinolite</td>
<td>18.60</td>
<td></td>
</tr>
<tr>
<td>clinozoisite</td>
<td>4.20</td>
<td></td>
</tr>
<tr>
<td>sphene</td>
<td>3.20</td>
<td>100.21</td>
</tr>
</tbody>
</table>

Calculated Chemical Composition

| SiO₂   | 60.71 |
| TiO₂   | 2.29  |
| Al₂O₃  | 16.46 |
| Fe₂O₃  | 0.62  |
| FeO    | 2.43  |
| MnO    | 0.005 |
| MgO    | 3.12  |
| CaO    | 8.54  |
| Na₂O   | 5.00  |
| K₂O    | 0.42  |
| H₂O⁺   | 0.53  |
| H₂O⁻   | 0.04  |

100.17
TABLE A-24
Aplitic Quartz Diorite Dike
Textural Observations and Interpretations

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape, Character</th>
<th>Distribution</th>
<th>Processes, Reactions</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>.08-1.0</td>
<td>subhedral laths, weakly zoned, with albite and Carlsbad twinning.</td>
<td>random framework</td>
<td></td>
<td>magmatic</td>
</tr>
<tr>
<td>quartz</td>
<td>.1-.15</td>
<td>anhedral</td>
<td>interstitial, random</td>
<td></td>
<td>magmatic</td>
</tr>
<tr>
<td>actinolite</td>
<td>.02-.3</td>
<td>anhedral</td>
<td>interstitial, random</td>
<td></td>
<td>post-magmatic (?)</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>.07-.5</td>
<td>anhedral</td>
<td>random</td>
<td></td>
<td>post-magmatic</td>
</tr>
<tr>
<td>sphene</td>
<td>.02-.5</td>
<td>irregular to crudely diamond-shaped</td>
<td>random</td>
<td></td>
<td>post-magmatic</td>
</tr>
</tbody>
</table>
TABLE A-25
Albite-Actinolite Porphyry Dike
Mineralogical and Chemical Composition, and Textural Characteristics

<table>
<thead>
<tr>
<th>Sample #:</th>
<th>Location:</th>
<th>Coordinates:</th>
<th>Elevation:</th>
<th>Trend:</th>
<th>Width:</th>
</tr>
</thead>
<tbody>
<tr>
<td>142</td>
<td>near west contact of East Mass, Romeral Diorite</td>
<td>11225N</td>
<td>6575E</td>
<td>340m</td>
<td>10m</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol.%</th>
<th>Size(mm)</th>
<th>Distribution</th>
<th>Wt.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase phenocrysts</td>
<td>24.51</td>
<td>2-4</td>
<td>random</td>
<td>SiO₂  62.90</td>
</tr>
<tr>
<td>groundmass</td>
<td>47.58</td>
<td>.02-.05</td>
<td>matrix</td>
<td>TiO₂  0.14</td>
</tr>
<tr>
<td>Total plagioclase (about An 6)</td>
<td>72.09</td>
<td></td>
<td></td>
<td>Al₂O₃ 14.85</td>
</tr>
<tr>
<td>hornblende 'a'</td>
<td>1.16</td>
<td>.2-.4</td>
<td>phenos</td>
<td>Fe₂O₃ 0.47</td>
</tr>
<tr>
<td>actinolite</td>
<td>26.75</td>
<td>.01-.02</td>
<td>dissem. &amp; vnlt.</td>
<td>FeO   3.66</td>
</tr>
<tr>
<td>magnetite</td>
<td>0.04</td>
<td>.02-.07</td>
<td>with actinolite</td>
<td>MnO   0.002</td>
</tr>
<tr>
<td>ilmenite</td>
<td>0.04</td>
<td>.05</td>
<td>random</td>
<td>MgO   4.79</td>
</tr>
<tr>
<td>sphene</td>
<td>tr</td>
<td>.00X</td>
<td>around ilmenite</td>
<td>CaO   4.69</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>tr</td>
<td>.03</td>
<td>with actinolite</td>
<td>Na₂O  7.40</td>
</tr>
<tr>
<td>biotite</td>
<td>tr</td>
<td>.01-.02</td>
<td>with actinolite</td>
<td>K₂O   0.62</td>
</tr>
</tbody>
</table>

Calculated Chemical Composition

<table>
<thead>
<tr>
<th></th>
<th>Wt.%</th>
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</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>62.90</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.14</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.85</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.47</td>
</tr>
<tr>
<td>FeO</td>
<td>3.66</td>
</tr>
<tr>
<td>MnO</td>
<td>0.002</td>
</tr>
<tr>
<td>MgO</td>
<td>4.79</td>
</tr>
<tr>
<td>CaO</td>
<td>4.69</td>
</tr>
<tr>
<td>Na₂O</td>
<td>7.40</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.62</td>
</tr>
<tr>
<td>F</td>
<td>0.003</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.63</td>
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<tr>
<td>H₂O⁻</td>
<td>0.06</td>
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</table>

100.215
TABLE A-26
Dacite Porphyry Dike
Mineralogical and Chemical Composition, and Textural Characteristics

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Shape and Distribution</th>
<th>Calculated Chemical Composition</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase phenocrysts (core, An 40)</td>
<td>subhedral laths random orient.</td>
<td>SiO₂</td>
<td>64.02</td>
</tr>
<tr>
<td></td>
<td></td>
<td>TiO₂</td>
<td>0.49</td>
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<tr>
<td></td>
<td></td>
<td>Al₂O₃</td>
<td>17.86</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Fe₂O₃</td>
<td>0.40</td>
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<tr>
<td></td>
<td></td>
<td>FeO</td>
<td>1.96</td>
</tr>
<tr>
<td></td>
<td></td>
<td>MnO</td>
<td>0.001</td>
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<td></td>
<td></td>
<td>MgO</td>
<td>2.41</td>
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<td>CaO</td>
<td>7.13</td>
</tr>
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<td></td>
<td></td>
<td>Na₂O</td>
<td>5.06</td>
</tr>
<tr>
<td></td>
<td></td>
<td>K₂O</td>
<td>0.42</td>
</tr>
<tr>
<td></td>
<td></td>
<td>F</td>
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</tr>
<tr>
<td></td>
<td></td>
<td>H₂O⁺</td>
<td>0.34</td>
</tr>
<tr>
<td></td>
<td></td>
<td>H₂O⁻</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>100.12</td>
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<td></td>
<td></td>
<td></td>
<td>99.30</td>
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</tbody>
</table>
TABLE A-27  
Diorite Pegmatite  
Mineralogical and Chemical Composition  

Sample #: 1272  
Coordinates: 12250N 4600E  
Elevation: 820m

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol. %</th>
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<tr>
<td>plagioclase</td>
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</tr>
<tr>
<td>core An 53</td>
<td>25.5</td>
</tr>
<tr>
<td>inner rim</td>
<td></td>
</tr>
<tr>
<td>An 52</td>
<td>3.0</td>
</tr>
<tr>
<td>outer rim</td>
<td></td>
</tr>
<tr>
<td>An 37</td>
<td>11.7</td>
</tr>
<tr>
<td>unzoned</td>
<td></td>
</tr>
<tr>
<td>An 48 ± 2</td>
<td>29.1</td>
</tr>
<tr>
<td>Total plagioclase</td>
<td>69.3</td>
</tr>
<tr>
<td>augite</td>
<td>12.8</td>
</tr>
<tr>
<td>actinolite</td>
<td>12.6</td>
</tr>
<tr>
<td>sphene</td>
<td>1.76</td>
</tr>
<tr>
<td>ilmenite</td>
<td>1.40</td>
</tr>
<tr>
<td>hematite</td>
<td>0.02</td>
</tr>
<tr>
<td>hematite</td>
<td></td>
</tr>
<tr>
<td>pseudobrookite</td>
<td>0.02</td>
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<tr>
<td>clinozoisite</td>
<td>1.5</td>
</tr>
<tr>
<td>scapolite</td>
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<tr>
<td>(marialitic)</td>
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Calculated Chemical Composition  

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>52.39</td>
</tr>
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<td>TiO₂</td>
<td>2.51</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>19.49</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.80</td>
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<tr>
<td>FeO</td>
<td>3.65</td>
</tr>
<tr>
<td>MnO</td>
<td>0.05</td>
</tr>
<tr>
<td>MgO</td>
<td>4.37</td>
</tr>
<tr>
<td>CaO</td>
<td>12.11</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.83</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.34</td>
</tr>
<tr>
<td>Cl</td>
<td>0.01</td>
</tr>
<tr>
<td>V₂O₃</td>
<td>0.004</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.58</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.05</td>
</tr>
</tbody>
</table>

100.18
<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape and Character</th>
<th>Distribution</th>
<th>Processes, Reactions</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>1.0-10</td>
<td>subhedral laths, normally zoned, and polysynthetically twinned.</td>
<td>framework of randomly oriented laths</td>
<td>began crystallizing before augite</td>
<td>early magmatic to late magmatic (rims)</td>
</tr>
<tr>
<td>augite</td>
<td>.07-1.2</td>
<td>(1) equant grains (2) anhedral</td>
<td>(1) single and in clusters (2) interstitial</td>
<td>eutectic crystallisation with plagioclase</td>
<td>middle to late magmatic</td>
</tr>
<tr>
<td>actinolite</td>
<td>.01-4.0</td>
<td>poikiloblastic</td>
<td>intergrown mostly with augite, but some with plagioclase</td>
<td>replaces augite and plagioclase</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>ilmenite</td>
<td>.05-3.0</td>
<td>anhedral, convoluted grains</td>
<td>interstitial to plagioclase intergrown with sphene, and associated with actinolite</td>
<td></td>
<td>middle magmatic to post-magmatic</td>
</tr>
<tr>
<td>sphene</td>
<td>.02-.3</td>
<td>anhedral</td>
<td>(1) intergrown with ilmenite (2) rims on ilmenite (3) in veinlets (4) disseminated</td>
<td>(1-2) product of ilmenite-plagioclase reaction (3-4) added</td>
<td>late magmatic to post-magmatic</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>.01-1.3</td>
<td>anhedral scraps and poikiloblastic clots</td>
<td>in plagioclases, and associated with actinolite</td>
<td>replaces plagioclase</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>scapolite (marialitic)</td>
<td>.5</td>
<td>irregular patches</td>
<td>in plagioclase, and associated with clinozoisite</td>
<td>replaces plagioclase</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>hematite and pseudobrookite</td>
<td>.02</td>
<td>extremely fine, vermicular intergrowths in sub-octahedral grains</td>
<td>inclusions in augite</td>
<td>oxidation of titaniferous magnetite</td>
<td>inclusions oxidation late to post-magmatic</td>
</tr>
</tbody>
</table>
# Table A-29

Porphyritic Meladiorite

Mineralogical and Chemical Composition

<table>
<thead>
<tr>
<th>Sample #: 1194</th>
<th>Coordinates: 13000N 5380E</th>
<th>Elevation: 575m</th>
</tr>
</thead>
</table>

## Porphyritic Meladiorite

### Mineral and Composition

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol.%</th>
<th>Wt. %</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>phenocrysts (core, An 64)</td>
<td>14.19</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(rim, An 48)</td>
<td>5.79</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total phenos</td>
<td>19.98</td>
<td></td>
<td></td>
</tr>
<tr>
<td>plagioclase matrix (An 64)</td>
<td>36.98</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total plagioclase</td>
<td>56.96</td>
<td></td>
<td></td>
</tr>
<tr>
<td>orthoclase</td>
<td>tr</td>
<td></td>
<td></td>
</tr>
<tr>
<td>augite</td>
<td>8.45</td>
<td></td>
<td></td>
</tr>
<tr>
<td>hornblende 'a'</td>
<td>27.30</td>
<td></td>
<td></td>
</tr>
<tr>
<td>biotite 'b'</td>
<td>5.17</td>
<td></td>
<td></td>
</tr>
<tr>
<td>chlorapatite</td>
<td>0.52</td>
<td></td>
<td></td>
</tr>
<tr>
<td>magnetite</td>
<td>1.25</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ilmenite</td>
<td>0.56</td>
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</tr>
<tr>
<td>hematite</td>
<td>0.10</td>
<td></td>
<td></td>
</tr>
<tr>
<td>sphene</td>
<td>tr</td>
<td></td>
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</tr>
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### Average Composition

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<thead>
<tr>
<th>Diorite</th>
<th>Romeral Diorite</th>
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<tr>
<td>SiO₂</td>
<td>48.25</td>
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<tr>
<td>TiO₂</td>
<td>1.04</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>18.99</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>3.98</td>
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<tr>
<td>FeO</td>
<td>4.85</td>
</tr>
<tr>
<td>MnO</td>
<td>0.08</td>
</tr>
<tr>
<td>MgO</td>
<td>6.18</td>
</tr>
<tr>
<td>CaO</td>
<td>12.27</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.53</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.78</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.21</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.02</td>
</tr>
<tr>
<td>Cl</td>
<td>0.04</td>
</tr>
<tr>
<td>F</td>
<td>0.08</td>
</tr>
<tr>
<td>V₂O₃</td>
<td>--</td>
</tr>
<tr>
<td>B₂O₃</td>
<td>--</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.64</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.07</td>
</tr>
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</table>

| Wt. % | 100.012 | 100.219 |

<p>| Wt. % | 100.012 | 100.219 |</p>
<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape and Character</th>
<th>Distribution</th>
<th>Processes, Reactions</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>1-4</td>
<td>subhedral laths (inner rim zones contain inclusions of augite and magnetite)</td>
<td>random</td>
<td>co-crystallization of plagioclase, augite, and magnetite</td>
<td>magmatic</td>
</tr>
<tr>
<td>phenocrysts</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>plagioclase</td>
<td>.1-.6</td>
<td>subhedral laths</td>
<td>matrix, random</td>
<td>exsolution</td>
<td>magmatic</td>
</tr>
<tr>
<td>matrix</td>
<td></td>
<td></td>
<td>orientations</td>
<td></td>
<td></td>
</tr>
<tr>
<td>orthoclase</td>
<td>.005-.08</td>
<td>lamellae</td>
<td>in plagioclases, along (010) planes</td>
<td>late or post-magmatic</td>
<td></td>
</tr>
<tr>
<td>augite</td>
<td>.3-.7</td>
<td>a) phenocrysts (stubby prisms)</td>
<td>random</td>
<td>middle to late magmatic</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>b) anhedral</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>.01-.02</td>
<td>c) equant, rounded</td>
<td>interstitial (between plagioclase grains)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>hornblende (actinolitic)</td>
<td>.01-.7</td>
<td>mostly anhedral</td>
<td>inclusions in plagioclase phenocrysts</td>
<td>replaces augite</td>
<td>late magmatic</td>
</tr>
<tr>
<td>biotite &quot;b&quot;</td>
<td>.02-.2</td>
<td>subhedral booklets</td>
<td>-interstitial</td>
<td></td>
<td></td>
</tr>
<tr>
<td>chlorapatite</td>
<td>.08-.2</td>
<td>rounded hexagonal prisms</td>
<td>-polikliloblastic inter-growths with augite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>magnetite</td>
<td>.1 -.9</td>
<td>rounded, angular, irregular</td>
<td>associated with hornblende</td>
<td>may replace hornblende</td>
<td>late magmatic</td>
</tr>
<tr>
<td>ilmenite</td>
<td>.1 -.9</td>
<td>-tabular lamellae -anhedral grains</td>
<td>random</td>
<td></td>
<td>magmatic(?)</td>
</tr>
<tr>
<td>hematite</td>
<td>.00X</td>
<td>thin lamellae</td>
<td>- in magnetite grains</td>
<td>exsolution</td>
<td>late magmatic(?)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>along basal planes in ilmenite and octahedral planes in magnetite</td>
<td>exsolution and oxidation</td>
<td>late magmatic to post magmatic</td>
</tr>
</tbody>
</table>
Before argillization, Sample #711 contained about 14% plagioclase phenocrysts and about 33% plagioclase groundmass.

### TABLE A-31

**Intramineral (IM-1) Diorite Porphyry**

**Mineralogical Composition**

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Location</th>
<th>Coordinates</th>
<th>Elevation</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
</tr>
</thead>
<tbody>
<tr>
<td>1416</td>
<td>SW Main Pit (cuts ore)</td>
<td>11570N 6018E 310m.</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>592</td>
<td>NE Main Pit (in ore)</td>
<td>11710N 6153E 310m.</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>690</td>
<td>120 m. SE of ore limit, Main Orebody</td>
<td>11420N 6200E 360m.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>711</td>
<td>300 m. SE of ore limit, Main Orebody</td>
<td>12238N 6190E 360m.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Minerals**

- plagioclase phenocrysts
  - core
  - inner rim
  - rim
  - unzoned

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase phenocrysts</td>
<td>26.00</td>
<td>20.00</td>
<td>14.36</td>
<td>2.69</td>
</tr>
<tr>
<td>actinolite</td>
<td>21.5</td>
<td>23.68</td>
<td>37.88</td>
<td>38.64</td>
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<tr>
<td>hornblende 'a'</td>
<td>--</td>
<td>1.28</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>penninite</td>
<td>18.50</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>epidote</td>
<td>tr</td>
<td>0.22</td>
<td>12.46</td>
<td>4.78</td>
</tr>
<tr>
<td>biotite 'g'</td>
<td>--</td>
<td>3.38</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>sphene</td>
<td>tr</td>
<td>--</td>
<td>0.80</td>
<td>--</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>tr</td>
<td>--</td>
<td>0.36</td>
<td>1.19</td>
</tr>
<tr>
<td>tourmaline</td>
<td>--</td>
<td>tr</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>scapolite</td>
<td>--</td>
<td>tr</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>nontronite</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>0.85</td>
</tr>
<tr>
<td>montmorillonite</td>
<td>7.50</td>
<td>3.62</td>
<td>--</td>
<td>42.33 (after plag.)</td>
</tr>
<tr>
<td>ilmenite</td>
<td>tr</td>
<td>0.57</td>
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<td>--</td>
</tr>
<tr>
<td>titanomagnetite</td>
<td>--</td>
<td>2.18</td>
<td>0.69</td>
<td>2.05</td>
</tr>
<tr>
<td>magnetite</td>
<td>--</td>
<td>1.15</td>
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<td>--</td>
</tr>
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<td>hematite</td>
<td>--</td>
<td>--</td>
<td>0.01</td>
<td>--</td>
</tr>
<tr>
<td>chalcopyrite</td>
<td>--</td>
<td>--</td>
<td>tr</td>
<td>--</td>
</tr>
<tr>
<td>bornite</td>
<td>--</td>
<td>--</td>
<td>tr</td>
<td>--</td>
</tr>
</tbody>
</table>

**Total**

| Total | 100.00 | 100.00 | 100.17 | 100.30 |

* Before argillization, Sample #711 contained about 14% plagioclase phenocrysts and about 38% plagioclase groundmass.
### TABLE A-32
Intramineral (IV-1) Diorite Porphyry
Textural Characteristics

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Observations</th>
<th>Distribution</th>
<th>Interpretations:</th>
<th>Reactions, Processes</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase phenocrystals</td>
<td>1-5 subhedral laths, relatively unzoned</td>
<td>random</td>
<td>replaces hornblende</td>
<td>post-magmatic</td>
<td>early magmatic</td>
</tr>
<tr>
<td>groundmass</td>
<td>.05-.15 subhedral laths and anhedral grains</td>
<td>matrix</td>
<td></td>
<td>magmatic</td>
<td></td>
</tr>
<tr>
<td>hornblende</td>
<td>.05-.4 subhedral prisms, remnants of phenocrystals</td>
<td></td>
<td></td>
<td>magmatic</td>
<td></td>
</tr>
<tr>
<td>actinolite</td>
<td>.01-2.0 -single needles</td>
<td>random, in veinlets, and</td>
<td>replaces hornblende</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td></td>
<td>-radiating groups</td>
<td>with hornblende</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>-rims on hornblende</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>-bundles of parallel needles pseudo-domorph after hornblende phenocrysts</td>
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<td></td>
</tr>
<tr>
<td>biotite 'g'</td>
<td>.04 subhedral booklets and clots</td>
<td>associated with amphiboles</td>
<td>replaces amphiboles</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>sphene</td>
<td>.03-.15 anhedral to subhedral diamond shaped grains</td>
<td>random</td>
<td></td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>chlorapatite</td>
<td>.08-.15 anhedral and subhedral hexagonal prisms</td>
<td>associated with epidote and</td>
<td>alteration of plag.</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td></td>
<td>submicroscopic platelets</td>
<td>actinolite, and in veinlets</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>montmorillonite</td>
<td>.05 clots of radially arranged platelets</td>
<td>in plagioclases</td>
<td>alteration</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>nontronite (?)</td>
<td></td>
<td>clots .01-.05 mm., randomly located in groundmass plag.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>tourmaline</td>
<td>.05 anhedral</td>
<td>random</td>
<td>replaces rock</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>ilmenite</td>
<td>.1 anhedral, convoluted grains with sphenic rims</td>
<td>disseminated</td>
<td></td>
<td>magmatic</td>
<td></td>
</tr>
<tr>
<td>titanomagnetite</td>
<td>.05 suboctahedral</td>
<td>disseminated</td>
<td></td>
<td>magmatic</td>
<td></td>
</tr>
<tr>
<td>titanomaghemite-pseudobrookite</td>
<td>cloudlike &quot;myrmekitic&quot; intergrowths</td>
<td>in titanomagnetite</td>
<td>oxidation of titanomagnetite</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>magnetite</td>
<td>.2 spongelike skeletal grains</td>
<td>random</td>
<td>inclusions(?)</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>hematite</td>
<td>.05-.1 lamellae</td>
<td>in magnetite</td>
<td>oxidation(?)</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>chalcopyrite-bornite</td>
<td>anhedral intergrowths</td>
<td>in plagioclase phenocrystal</td>
<td>replacement</td>
<td>post-magmatic</td>
<td></td>
</tr>
<tr>
<td>Sample #</td>
<td>Location</td>
<td>Univ.</td>
<td>Calc.</td>
<td>Mexican</td>
<td>Chem.</td>
</tr>
<tr>
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<td>----------</td>
<td>-------</td>
<td>-------</td>
<td>---------</td>
<td>-------</td>
</tr>
<tr>
<td>1416</td>
<td>SW Main Pit (cuts ore)</td>
<td>SW Main Pit (in ore)</td>
<td>NE Main Pit (in ore)</td>
<td>711</td>
<td>300 m. SE of ore limit</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Control</td>
<td></td>
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</table>

<table>
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<tr>
<th>constituent</th>
<th>Wt.%</th>
<th>Wt.%</th>
<th>Wt.%</th>
<th>Wt.%</th>
<th>Wt.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>52.69</td>
<td>53.40</td>
<td>53.72</td>
<td>53.70</td>
<td>50.26</td>
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<tr>
<td>TiO₂</td>
<td>0.08</td>
<td>0.65</td>
<td>0.72</td>
<td>0.72</td>
<td>0.33</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.39</td>
<td>16.30</td>
<td>15.50</td>
<td>16.70</td>
<td>11.40</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.84</td>
<td>2.74</td>
<td>2.42</td>
<td>4.89</td>
<td>4.67</td>
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<tr>
<td>FeO</td>
<td>3.47</td>
<td>4.17</td>
<td>5.08</td>
<td>5.56</td>
<td>6.86</td>
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<tr>
<td>MnO</td>
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<td>0.08</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
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<tr>
<td>MgO</td>
<td>10.52</td>
<td>5.87</td>
<td>6.11</td>
<td>4.57</td>
<td>8.04</td>
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<tr>
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<td>7.14</td>
<td>10.38</td>
<td>11.46</td>
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<td>3.59</td>
<td>4.90</td>
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<tr>
<td>K₂O</td>
<td>0.40</td>
<td>0.31</td>
<td>0.32</td>
<td>0.70</td>
<td>0.63</td>
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<td>CO₂</td>
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<td>Cu</td>
<td>--</td>
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<tr>
<td>V₂O₅</td>
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<td>--</td>
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<td>H₂O⁺</td>
<td>3.30</td>
<td>1.65</td>
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<td>0.86</td>
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<tr>
<td>H₂O⁻</td>
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<td>0.08</td>
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<td>Sum</td>
<td>100.18</td>
<td>100.30</td>
<td>100.28</td>
<td>100.19</td>
<td>100.31</td>
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</table>

TABLE A-33
Intramineral (IM-1) Diorite Porphyry
Chemical Compositions
### TABLE A-34
Intramineral (IM-2) Dioritite
Mineralogical Composition

<table>
<thead>
<tr>
<th>Sample #:</th>
<th>Location:</th>
<th>Coordinates:</th>
<th>Elevation:</th>
<th>Mineral Composition</th>
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<td>West Side</td>
<td>11793N 11762N 6017E</td>
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<tr>
<td></td>
<td></td>
<td>536</td>
<td>West Side</td>
<td>11762N 6024E</td>
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<td>1419</td>
<td>East Side</td>
<td>11631N 6114E</td>
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<td>1424</td>
<td>East Side</td>
<td>11602N 6135E</td>
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<td></td>
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<td>1419</td>
<td>Main Pit</td>
<td>310m.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1424</td>
<td>Main Pit</td>
<td>310m.</td>
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<table>
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<tr>
<th>Mineral</th>
<th>Vol.% An</th>
<th>Vol.% An</th>
<th>Vol.% An</th>
<th>Vol.% An</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase (fine-grained to microcrystalline)</td>
<td>26.95 52*</td>
<td>9.86 30</td>
<td>30.20 29</td>
<td>--</td>
</tr>
<tr>
<td>actinolite</td>
<td>72.21</td>
<td>38.00</td>
<td>49.20</td>
<td>7.50</td>
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<tr>
<td>penninite</td>
<td>--</td>
<td>6.34</td>
<td>0.17</td>
<td>17.30</td>
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<tr>
<td>scapolite (marialitic)</td>
<td>--</td>
<td>32.00</td>
<td>--</td>
<td>8.80</td>
</tr>
<tr>
<td>epidote (low-iron)</td>
<td>--</td>
<td>--</td>
<td>0.17</td>
<td>43.00</td>
</tr>
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<td>biotite (green)</td>
<td>0.67</td>
<td>6.84</td>
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<td>sphene</td>
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<td>--</td>
<td>--</td>
</tr>
<tr>
<td>chlorapatite</td>
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<td>--</td>
<td>--</td>
</tr>
<tr>
<td>calcite</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>0.30</td>
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<td>montmorillonite</td>
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<td>--</td>
</tr>
<tr>
<td>ilmenite</td>
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<td>0.14</td>
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<td>titanomagnetite</td>
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<td>1.0</td>
<td>18.94</td>
<td>20.61</td>
</tr>
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<td>hematite</td>
<td>tr</td>
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<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>pyrite</td>
<td>--</td>
<td>--</td>
<td>1.35</td>
<td>1.56</td>
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<tr>
<td>chalcopyrite</td>
<td>--</td>
<td>--</td>
<td>0.06</td>
<td>--</td>
</tr>
<tr>
<td>Total:</td>
<td>100.00</td>
<td>99.99</td>
<td>100.11</td>
<td>99.07</td>
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</table>

* Although this sample comes from a dike which cuts magnetite and is on the projection of the intramineral dike from which sample 536 was taken, the high An content of its plagioclase suggests it may be a different rock type.
<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape, Character</th>
<th>Distribution</th>
<th>Interpreations: Reactions, Processes</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>mostly .06-.2 some to .9</td>
<td>subhedral laths, polycrystalline twinned</td>
<td>evenly distributed in unaltered rocks</td>
<td>recrystallization without much change in composition</td>
<td>magmatic</td>
</tr>
<tr>
<td>plagioclase</td>
<td>.05-.1</td>
<td>anhedral, unzoned, untwinned</td>
<td>random</td>
<td>recrystallization without much change in composition</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>actinolite</td>
<td>.01-.4.0</td>
<td>-parallel clusters, -radial clusters, -single needles</td>
<td>disseminated, in veinlets, in clots.</td>
<td>-replaces plagioclase -may have replaced hornblende, but left no remnants</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>penninite</td>
<td>.1-.4.0</td>
<td>pseudomorphous</td>
<td>after actinolite</td>
<td>replaces actinolite</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>epidote</td>
<td>.05-.1</td>
<td>-single grains -clasts of equant grains to 5 cm. across</td>
<td>-in plagioclase</td>
<td>-replaces plagioclase -replaces diorite</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>biotite 'g'</td>
<td>.02-.08</td>
<td>single booklets, and clots of booklets</td>
<td>with actinolite</td>
<td>replaces actinolite(?)</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>sphene</td>
<td>.01-.05</td>
<td>anhedral</td>
<td>disseminated, and in rims on ilmenite</td>
<td>replaces rock, reaction product of plagioclase and ilmenite</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>scapolite</td>
<td>.01-.15</td>
<td>clots of equigranular polyhedra</td>
<td>in veinlets with epidote</td>
<td>replaces plagioclase</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>calcite</td>
<td>.05-.1</td>
<td>anhedral</td>
<td>in plagioclase</td>
<td>replaces plagioclase</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>montmorillonite</td>
<td>.000X</td>
<td>sub-microscopic</td>
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**TABLE A-35 (cont.)**

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<th>Mineral</th>
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<th>Shape, Character</th>
<th>Distribution</th>
<th>Reactions, Processes</th>
<th>Interpretations</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>ilmenite</td>
<td>.05</td>
<td>irregular, convoluted, with sphenes</td>
<td>-random in titanomagnetite</td>
<td>exsolution</td>
<td>magmatic</td>
<td>magmatic</td>
</tr>
<tr>
<td>titanomagnetite*</td>
<td>.02-.08</td>
<td>suboctahedral</td>
<td>random with titanomaghemite pseudo-brookite intergrowths***</td>
<td>-intergrowths by oxidation</td>
<td>magmatic</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>magnetite**</td>
<td>.05-.7</td>
<td>skeletal, partial octahedra</td>
<td>disseminated, and in vein-lets, with actinolite</td>
<td>added</td>
<td>post-magmatic</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>pyrite</td>
<td>.02-.2.5</td>
<td>cubes and angular, anhedral grains</td>
<td>with epidote</td>
<td>replaces rock, especially actinolite</td>
<td>post-magmatic</td>
<td>post-magmatic</td>
</tr>
<tr>
<td>chalcopyrite</td>
<td>.01-.015</td>
<td>rounded blebs</td>
<td>in pyrite</td>
<td>exsolution</td>
<td>post-magmatic</td>
<td>post-magmatic</td>
</tr>
</tbody>
</table>

* Ilmenite and pale brownish pink, slightly anisotropic suboctahedral grains of titaniferous magnetite are present in western IM-2 dikes but are absent in eastern IM-2 dikes. Ilmenite and titaniferous magnetite are probably magmatic.

** Skeletal suboctahedral and "spongy" anhedral, isotropic, low-titanium magnetite grains are present in IM-2 dikes on both sides of the orebody, and they are abundant in eastern IM-2 dikes. They probably formed by replacement.

*** Uytenbagaart and Burke, 1971, p. 168.
### TABLE A-36

Intramineral (IM-2) Diorite Dikes
Calculated Chemical Compositions

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Location:</th>
<th>630 West Side Main Pit</th>
<th>536 West Side Main Pit</th>
<th>1419 East Side Main Pit</th>
<th>1424 East Side Main Pit</th>
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<tr>
<td>Constituent</td>
<td>Wt.%</td>
<td>Wt.%</td>
<td>Wt.%</td>
<td>Wt.%</td>
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<tr>
<td>SiO₂</td>
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<td>51.14</td>
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<tr>
<td>TiO₂</td>
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<td>0.44</td>
<td>0.13</td>
<td>0.08</td>
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<td>Al₂O₃</td>
<td>9.07</td>
<td>11.48</td>
<td>7.55</td>
<td>14.58</td>
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<tr>
<td>Fe₂O₃</td>
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<td>5.06</td>
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<td>24.19</td>
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<td>0.03</td>
<td>--</td>
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<td>0.09</td>
<td>0.01</td>
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### TABLE A-37
Intramineral (IM-3) Dioritite Dikes
Mineralogical Compositions

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<th>Location</th>
<th>Coordinates</th>
<th>Elevation</th>
<th>Mineral</th>
<th>Vol.% An</th>
<th>Vol.% An</th>
<th>Vol.% An</th>
<th>Vol.% An</th>
</tr>
</thead>
<tbody>
<tr>
<td>594</td>
<td>West Side Main Pit (in ore)</td>
<td>11726N 6020E</td>
<td>310 m.</td>
<td>plagioclase</td>
<td>44.39</td>
<td>28</td>
<td>29 ±2</td>
<td>28.40</td>
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<tr>
<td>803</td>
<td>East Side Main Pit (in ore)</td>
<td>11624N 6083E</td>
<td>310 m.</td>
<td>hornblende 'a'</td>
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<td>--</td>
<td>24.20</td>
<td>15.18</td>
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<tr>
<td>738</td>
<td>East Side Main Pit (edge of ore)</td>
<td>11469N 6160E</td>
<td>350 m.</td>
<td>actinolite</td>
<td>0.83</td>
<td>27.24</td>
<td>25.63</td>
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</tr>
<tr>
<td>708</td>
<td>Southeast of Main Pit (outside ore)</td>
<td>11237N 6190E</td>
<td>360 m.</td>
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<td>15.94</td>
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</tr>
<tr>
<td></td>
<td></td>
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<td></td>
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<td>--</td>
<td>0.29</td>
<td>1.42</td>
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<tr>
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<td></td>
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<td>epidote</td>
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<td>18.20</td>
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<td>montmorillonite</td>
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</tr>
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<td></td>
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<td>ilmenite</td>
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<td>0.04</td>
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<td>magnetite</td>
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<td>hematite</td>
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<td>tr</td>
<td>tr</td>
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</tr>
<tr>
<td></td>
<td></td>
<td></td>
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<td>chalcopyrite</td>
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<td>tr</td>
<td>--</td>
<td>--</td>
</tr>
</tbody>
</table>

<p>|          |          |             |           |               | 100.00   | 99.98    | 100.18   | 100.59   |</p>
<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape, Character</th>
<th>Distribution</th>
<th>Interpretations, Processes</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>.02-.3</td>
<td>subhedral laths, relatively unzoned, Carlsbad and albite twinning.</td>
<td>evenly distributed, weakly aligned, parallel to dike walls.</td>
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<tr>
<td>plagioclase</td>
<td>.05-.3</td>
<td>anhedral</td>
<td>random</td>
<td>recrystallization with little change in composition</td>
<td>post-magmatic</td>
</tr>
<tr>
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<td>.05-.4</td>
<td>subhedral prisms and groups of prisms</td>
<td>evenly disseminated</td>
<td></td>
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<tr>
<td>actinolite</td>
<td>.002-.05</td>
<td>1. rims</td>
<td>1. on hornblende</td>
<td>1. replaces hornblende</td>
<td>post-magmatic</td>
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<td>2. pseudomorphs</td>
<td>2. of hornblende</td>
<td>2. replaces hornblende</td>
<td>post-magmatic</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3. needles and radial clusters</td>
<td>3. in plagioclase</td>
<td>3. replaces plagioclase</td>
<td>post-magmatic</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4. slender prisms</td>
<td>4. in veinlets</td>
<td>4. fracture filling</td>
<td>post-magmatic</td>
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<td>penninite</td>
<td>.05-.25</td>
<td>sheaves of platelets</td>
<td>intergrows with amphiboles</td>
<td>replaces amphiboles</td>
<td>post-magmatic</td>
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<td>biotite (green)</td>
<td>.02-.07</td>
<td>single booklets and clusters to .25 mm.</td>
<td>with amphiboles</td>
<td>replaces amphiboles(?)</td>
<td>late-mag.(?)</td>
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<td>.05</td>
<td>subhedral hexagonal prisms and anhedral grains</td>
<td>random, and in veinlets</td>
<td></td>
<td>post-magatic(?)</td>
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<tr>
<td>sphene</td>
<td>.03-.5</td>
<td>anhedral, very irregular shapes</td>
<td>1. rims on ilmenite</td>
<td>1. ilmen.-plagioclase reaction</td>
<td>post-magmatic</td>
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<td>2. replaces rock</td>
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<td>Size (mm)</td>
<td>Shape, Character</td>
<td>Distribution</td>
<td>Interpretations:</td>
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<td>epidote</td>
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<td>1. replaces plagioclase</td>
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<td>2. groups of polyhedral grains</td>
<td>2. random distribution</td>
<td>2. replaces rock</td>
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<td>suboctahedral and anhedral grains (slightly anisotropic)</td>
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<td>3. fracture filling</td>
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<td>.02-.1</td>
<td>suboctahedral</td>
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<td>1. on octahedral planes in magnetite</td>
<td>oxidation</td>
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<td>spongelike &quot;myrmekitic&quot; intergrowths</td>
<td>in titaniferous magnetites</td>
<td>oxidation</td>
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<td>random</td>
<td>replaces rock</td>
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<td>montmorillonite</td>
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<td>in plagioclase</td>
<td>alteration</td>
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<td>anhedral</td>
<td>in veinlets with actinolite and magnetite</td>
<td>fracture filling</td>
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<td>goethite</td>
<td>.5-2.0</td>
<td>sub-cubic aggregates</td>
<td>pseudomorph after randomly distributed pyrite</td>
<td>weathering of post-magmatic pyrite</td>
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**Stage**

- post-magmatic
- magmatic
- late-magmatic
- post-magmatic
### TABLE A-39
Intramineral (IM-3) Dioritite Dikes
Chemical Compositions

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<th>Wt. %</th>
<th>Wt. %</th>
<th>Wt. %</th>
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<td>TiO₂</td>
<td>Al₂O₃</td>
<td>Fe₂O₃</td>
<td>FeO</td>
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<td>17.09</td>
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<td>MnO</td>
<td>MgO</td>
<td>CaO</td>
<td>Na₂O</td>
<td>K₂O</td>
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<td>3.37</td>
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<td>8.47</td>
<td>4.18</td>
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<td>6.64</td>
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<td>P₂O₅</td>
<td>CO₂</td>
<td>Cl₂</td>
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<td>H₂O⁺</td>
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<td>0.06</td>
<td>0.11</td>
<td>0.04</td>
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Total: 100.04 99.65 100.23 100.20 100.23
TABLE A-40
Intramineral (IM-4) Porphyritic Dioritite
Mineralogical Compositions

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<thead>
<tr>
<th>Sample #:</th>
<th>Location:</th>
<th>Coordinates:</th>
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<td>540</td>
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<td>310 m.</td>
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<tr>
<td>508</td>
<td>East Side Main Pit</td>
<td>11515N, 6148E</td>
<td>340 m.</td>
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<table>
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<th>Mineral</th>
<th>Vol. % An</th>
<th>Vol. % An</th>
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<tbody>
<tr>
<td>plagioclase microphenocrysts</td>
<td>7.00 27</td>
<td>17.00 29</td>
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<tr>
<td>groundmass plagioclase</td>
<td>48.45 27</td>
<td>52.13 29</td>
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<tr>
<td>secondary plagioclase</td>
<td>3.00</td>
<td>--</td>
</tr>
<tr>
<td>Total plagioclase</td>
<td>58.45</td>
<td>69.13</td>
</tr>
<tr>
<td>actinolite</td>
<td>--</td>
<td>7.74 (yellow-green, iron poor)</td>
</tr>
<tr>
<td>penninite</td>
<td>27.31</td>
<td>18.42</td>
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<tr>
<td>titanomagnetite</td>
<td>3.90</td>
<td>4.17</td>
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<tr>
<td>biotite 'g'</td>
<td>7.11</td>
<td>0.53</td>
</tr>
<tr>
<td>nontronite</td>
<td>2.99</td>
<td>--</td>
</tr>
<tr>
<td>calcite</td>
<td>0.17</td>
<td>--</td>
</tr>
<tr>
<td>chalcopyrite</td>
<td>tr</td>
<td>--</td>
</tr>
<tr>
<td>chalcocite-covellite</td>
<td>tr</td>
<td>--</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>tr</td>
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<tr>
<td>magnetite</td>
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<td>tr</td>
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<td>ilmenite</td>
<td>tr</td>
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<tr>
<td>Total</td>
<td>99.99</td>
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TABLE A-41
Intramineral (IM-4) Porphyritic Diorite
Textural Characteristics

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<tr>
<th>Mineral</th>
<th>Size (mm)</th>
<th>Shape, Character</th>
<th>Distribution</th>
<th>Interpretations: Reactions, Processes</th>
<th>Stage</th>
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<tr>
<td>plagioclase</td>
<td>.5-1.5</td>
<td>subhedral laths with poly-</td>
<td>random</td>
<td>feldspathization</td>
<td>early magmatic</td>
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<tr>
<td>microphenos</td>
<td></td>
<td>synthetic twinning</td>
<td></td>
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<tr>
<td>groundmass</td>
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<tr>
<td>plagioclase</td>
<td>.03-.15</td>
<td>equant grains and a few twinned</td>
<td>matrix</td>
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<td>magmatic</td>
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<tr>
<td></td>
<td></td>
<td>laths</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>secondary</td>
<td>.1-.3</td>
<td>anhedral sponge-like clots</td>
<td>-random</td>
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<td>post-magmatic</td>
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<tr>
<td>plagioclase</td>
<td></td>
<td></td>
<td>-in veinlets with chlorite and magnetite</td>
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<td>actinolite</td>
<td>.02 -.8</td>
<td>1. bundles of parallel needles</td>
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<td>replaces hornblende plagioclase</td>
<td>post-magmatic</td>
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<td></td>
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<td>2. single needles</td>
<td>2.3. in plagioclase</td>
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<td>3. radial clusters</td>
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<td>-parallel and radiating clusters</td>
<td>-pseudomorphous after actinolite</td>
<td>alteration of actinolite and plagioclase</td>
<td>post-magmatic</td>
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<td>of platelets</td>
<td>-in veinlets with magnetite and secondary plagioclase</td>
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<td>biotite 'g'</td>
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<td>single booklets and diffuse</td>
<td>disseminated</td>
<td>replacement</td>
<td>post-magmatic</td>
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<td>with chlorite and minor quartz</td>
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<td>anhedral shreds</td>
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<td>replacement, fracture filling</td>
<td>hydrothermal</td>
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TABLE A-42
Intramineral (IM-4) Porphyritic Dioritite
Calculated Chemical Compositions

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### Mineralogical Compositions

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<th>nepheline</th>
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<td>5980E</td>
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<td>11825N</td>
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<td>N 73 E</td>
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Late Dikes
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**Control:**

**Univ. Mexico Chemical Analysis**

**Calculated Chemical Composition**
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<th>Interpretations, Processes</th>
<th>Stage</th>
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TABLE A-47
La Totora (?) Quartz-bearing Diorite
Chemical Composition

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<td>crystallization from magma; myrmekitic rims and mottled cores probably indicate late stage water saturation</td>
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### Punta de Piedra Granodiorite

#### Chemical Compositions

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TABLE A-51
Granite Aplite
Mineralogical and Chemical Composition

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### TABLE A-53
Actinolite-Plagioclase Assemblage
Mineralogical Compositions

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TABLE A-54
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TABLE A-55
Actinolite-Sphene Assemblage
Mineralogical Composition

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<th>Vol. %</th>
<th>Vol. %</th>
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TABLE A-57
Aplitic Oligoclase-Albite-Quartz Rock
Mineralogical Composition

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<th>Vol.% An</th>
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mean
TABLE A-58
Aplitic Oligoclase-Albite-Quartz Rock
Chemical Composition

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TABLE A-59
Aplitic Quartz-Oligoclase-Albite Rock
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**Aplitic Oligoclase-Quartz Rock**  
**Mineralogical and Chemical Composition**

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**Coordinates:** 10540 N. 5600 E.  
**Elevation:** 330 m.

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<td>0.88</td>
</tr>
<tr>
<td>tourmaline</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>tr.</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>biotite</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>augite</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Total</td>
<td>99.94</td>
<td>100.06</td>
<td>100.05</td>
<td>99.94</td>
<td>100.05</td>
<td>99.88</td>
<td>99.74</td>
<td>99.65</td>
<td>99.78</td>
</tr>
</tbody>
</table>

Mean Romeral Diorite

0.98\(\pm\) 2.93

7.98\(\pm\) 0.67

17.76\(\pm\) 1.54

4.04\(\pm\) 2.26 (An 36)

25.55\(\pm\) 4.67

0.11

1.99\(\pm\) 1.23

1.20 (An 27)

1.86\(\pm\) 1.71

0.13\(\pm\) 0.12

0.04

0.14

2.30

2.05

0.18 apatite

0.10 penninite

0.01 hematite

tr. pyrite

tr. chalcopyrite

100.18
APPENDIX B

ANALYTICAL METHODS
TABLE A-64
Typical Romeral Ores
Approximate Mineralogical Compositions
(as reconstructed from assays and mineralogical observations)

<table>
<thead>
<tr>
<th>Mineral</th>
<th>high grade Vol.%</th>
<th>medium grade Vol.%</th>
<th>low grade Vol.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>magnetite</td>
<td>88.0</td>
<td>61.0</td>
<td>40.5</td>
</tr>
<tr>
<td>actinolite</td>
<td>12.0</td>
<td>37.0</td>
<td>55.0</td>
</tr>
<tr>
<td>chlorapatite</td>
<td>tr</td>
<td>1.0</td>
<td>3.0</td>
</tr>
<tr>
<td>pyrite</td>
<td>tr</td>
<td>0.2</td>
<td>0.5</td>
</tr>
<tr>
<td>sphene</td>
<td>tr</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>plagioclase (An27±5)</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>scapolite</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>clinozoisite</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>chlorite</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>100.0</strong></td>
<td><strong>100.0</strong></td>
<td><strong>100.0</strong></td>
</tr>
</tbody>
</table>
TABLE A-65
Typical Romeral Ores
Approximate Chemical Compositions
(as reconstructed from assays and mineralogical observations)

<table>
<thead>
<tr>
<th>Constituent</th>
<th>high grade Wt.%</th>
<th>medium grade Wt.%</th>
<th>low grade Wt.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>4.2</td>
<td>14.6</td>
<td>23.8</td>
</tr>
<tr>
<td>TiO₂</td>
<td>tr</td>
<td>0.5</td>
<td>0.6</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.2</td>
<td>0.7</td>
<td>1.2</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>63.7</td>
<td>49.7</td>
<td>36.8</td>
</tr>
<tr>
<td>FeO</td>
<td>29.5</td>
<td>25.3</td>
<td>21.4</td>
</tr>
<tr>
<td>MgO</td>
<td>1.2</td>
<td>4.0</td>
<td>6.6</td>
</tr>
<tr>
<td>CaO</td>
<td>1.0</td>
<td>3.9</td>
<td>7.0</td>
</tr>
<tr>
<td>Na₂O</td>
<td>tr</td>
<td>0.1</td>
<td>0.2</td>
</tr>
<tr>
<td>K₂O</td>
<td>tr</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.2</td>
<td>.5</td>
<td>0.9</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.2</td>
<td>tr</td>
<td>tr</td>
</tr>
<tr>
<td>Cl</td>
<td>tr</td>
<td></td>
<td>0.2</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.3</td>
<td></td>
<td>0.9</td>
</tr>
<tr>
<td>S</td>
<td>0.1</td>
<td></td>
<td>0.3</td>
</tr>
<tr>
<td>sulfide Fe</td>
<td></td>
<td>0.1</td>
<td>0.3</td>
</tr>
<tr>
<td>Total</td>
<td>100.2</td>
<td>99.8</td>
<td>100.2</td>
</tr>
</tbody>
</table>
Wholesale-rock chemical analyses were performed at the Laboratorios de Química (Ing. Alberto Obregón, chief analyst) of Instituto de Geología, Universidad Nacional Autónoma de México (Mexico City).

Briefly, the methods used were as follows: (Salas, 1972): SiO₂, colorimetric; Al₂O₃, precipitation of ammonia-insoluble hydroxides and ignition to R₂O₃; TiO₂, colorimetric using special acid solution of Ti(SO₄)₂; Fe₂O₃, calculation from total Fe and FeO; FeO, titration of ferrous sulfate with potassium dichromate; CaO, colorimetric by means of its precipitation as an oxalate and gravimetrically titrated with permanganate; MgO, titration using EDTA; Na₂O, flame photometric; MnO, colorimetric using potassium metaperiodate; P₂O₅, colorimetric using citromolybdate solution; SΟ₃, using the ascorbic acid-barium chloride procedure; CO₂, by means of HCl, caustic soda solution and bromophenol blue. The lower limits of detection are as follows: SiO₂, 0.3; TiO₂, 0.05; Al₂O₃, Fe₂O₃, FeO, MgO, Na₂O, K₂O, 0.2; CaO, 0.05; MnO, 0.03; P₂O₅, 0.02.
PROCEDURE FOR IRON ANALYSIS BY TITRATION

1. Pulverize the sample.
2. Weigh out a 1 gram sample.
3. Boil in hot HCl for 10 minutes to dissolve iron.
4. Reduce Fe$^{3+}$ to Fe$^{2+}$. Add SnCl$_2$, drop by drop until solution loses its color. Then add one more drop.
5. Boil again, and if color returns, repeat steps 4 and 5.
6. Add 15 ml. of HgCl$_2$ while solution is hot.
7. Cool to room temperature.
8. Add 25 ml. sodium diphenyl sulfonate indicator, and dilute to 200 ml. with H$_2$O.
9. Add KCr$_2$O$_7$ from pipette and stir as it is added. When color goes to purple and stays, solution is titrated. (When all Fe$^{2+}$ is oxidized to Fe$^{3+}$, the indicator oxidizes and becomes purple.)
10. Read the pipette. For a one gram sample, each ml. indicates about 0.5 grams of iron.

This procedure was suggested by Peter Avotins, a Stanford graduate student in metallurgical engineering. He also supplied the reagents and helped with the first several analyses. The procedure is the iron assaying procedure of the Cleveland Cliffs Mining Company.

Results were repeatable to within 0.1% iron.
APPENDIX C

MINERALS OF EL ROMERAL:

Diagnostic properties determined, and chemical compositions used in calculation of rock compositions.
ACTINOLITE

| SiO₂ | 54.33 |
| TiO₂ | 0.29 |
| Al₂O₃ | 2.68 |
| Fe₂O₃ | 1.09 |
| FeO | 11.68 |
| MnO | -- |
| MgO | 15.31 |
| CaO | 12.46 |
| Na₂O | 0.54 |
| K₂O | 0.12 |
| H₂O⁺ | 2.04 |
| H₂O⁻ | 0.20 |

D = 3.11

Nₓ = 1.627
Nᵧ = 1.641
Nᵣ = 1.650
Nᵣ-Nₓ = .023
2Vₓ = --
ZᵣC = 20°

El Romeral #962

Nₓ = 1.628
Nᵧ = 1.641
Nᵣ = 1.652
Nᵣ-Nₓ = .024
2Vₓ = 80°
ZᵣC = 10°

X = pale greenish yellow
Y = bluish green
Z = deep blue green

Z > Y > X

Assoc: actinolite vein cutting magnetite ore

Assoc: actin.-hb.-ep.-schist
Loc: Start Dist., So. Devon
Ref: Deer, Howie, and Zussman 1962, Vol. II, p. 252, #9
ACTINOLITIC HORNBLende

<table>
<thead>
<tr>
<th></th>
<th>Wt. %</th>
<th>Wt. %</th>
<th>Mean Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>51.90</td>
<td>48.92</td>
<td>50.41</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>3.11</td>
<td>5.88</td>
<td>4.49</td>
</tr>
<tr>
<td>TiO₂</td>
<td>.16</td>
<td>1.21</td>
<td>.69</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>4.48</td>
<td>6.50</td>
<td>5.49</td>
</tr>
<tr>
<td>FeO</td>
<td>8.13</td>
<td>7.79</td>
<td>7.96</td>
</tr>
<tr>
<td>MgO</td>
<td>16.10</td>
<td>14.32</td>
<td>15.21</td>
</tr>
<tr>
<td>CaO</td>
<td>12.58</td>
<td>11.37</td>
<td>11.98</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.53</td>
<td>1.20</td>
<td>1.36</td>
</tr>
<tr>
<td>K₂O</td>
<td>.09</td>
<td>.71</td>
<td>.40</td>
</tr>
<tr>
<td>MnO</td>
<td>.17</td>
<td>.17</td>
<td>.17</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>1.83</td>
<td>1.37</td>
<td>1.60</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>.18</td>
<td>.18</td>
<td>.18</td>
</tr>
<tr>
<td>F</td>
<td>--</td>
<td>.27</td>
<td>.14</td>
</tr>
</tbody>
</table>

El Romeral #1504

D = 3.14  3.15  3.145
Nₓ = 1.637 1.644
Nᵧ = 1.655
Nₓ = 1.665
Nₓ-Nₓ = .021  .021
2Vₓ = 75  75
ZxC = 20.5

X=blue-green  pale grn-gold  light olive green
Y=olive green  light olive grn.  olive brn.-grn.
Z=greenish gold  dark olive green  dusky green

Assoc: after augite in diorite  Assoc: diorite  Assoc: Romeral Diorite
Loc: Norway  (some after augite)

ARFVEDSONITE

<table>
<thead>
<tr>
<th>Element</th>
<th>Formula</th>
<th>Weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>47.63</td>
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</tr>
<tr>
<td>TiO₂</td>
<td>1.22</td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.45</td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>14.73</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>22.34</td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.44</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>0.25</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>2.8</td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
<td>7.34</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>2.11</td>
<td></td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>1.26</td>
<td></td>
</tr>
</tbody>
</table>

El Romeral #1432

OAP ⊥ (010), ǁ (100)

D = 3.4
N_X = 1.692
N_Y = 1.696
N_Z = 1.704
N_Z - N_X = 0.012
2V =
ZAC = 6-9°

2V_Z = 10° (+)
ZAC = 11°

length fast
X = "dirty" blue green
Y = gold-green
contains inclusions with pleochroic haloes

Assoc: nepheline syenite
Loc: Kola Penninsula

Assoc: late biotite-arfvedsonite-nepheline-bearing dioritite
APATITE (CHLORAPATITE)
\[ \text{Ca}_5(\text{PO}_4,\text{CO}_3)_3(\text{Cl}) \]

Diagnostic properties:

Sample #1082: apatite from magnetite-apatite cataclasite
\[ N_o = 1.6529 \] relatively high indices
\[ N_e = 1.6493 \] suggest chlorapatite
\[ N_o - N_e = 0.0036 \]
\[ 2V \approx 2-5^\circ \] (not quite uniaxial, which suggests \( \text{CO}_3^- \))
-bubbles slowly but steadily in nitric acid
\((\text{CO}_2 \text{ is evolved})\)
-addition of silver nitrate to the dissolved apatite causes immediate precipitation of white silver chloride, confirming the presence of \( \text{Cl}^- \).

Sample #643: from actinolite-sphene zone
\[ N_o = 1.6536 \]
\[ N_e = 1.6499 \]
\[ N_o - N_e = 0.0037 \]
\[ 2V \approx 2-5^\circ \]

Sample #115: actinolite-apatite-magnetite vein
\[ N_o = 1.6511 \]
\[ N_e = 1.6471 \]
\[ N_o - N_e = 0.0040 \]
\[ 2V \approx 2-5^\circ \]

Sample #476: apatite fragment in Romeral Fault breccia
\[ N_o = 1.6577 \]
\[ N_e = 1.6552 \]
\[ N_o - N_e = 0.0025 \]
\[ 2V \approx 2-5^\circ \]

Sample #557: from magnetite ore, Main Orebody
\[ N_o = 1.6528 \]
\[ N_e = 1.6489 \]
\[ N_o - N_e = 0.0039 \]
\[ 2V \approx 2-5^\circ \]
APATITE (CHLORAPATITE)

Diagnostic properties (cont.):

Sample #1241: from apatite-actinolite vein

\[
N_o = 1.6535 \\
N_e = 1.6498 \\
N_o - N_e = 0.0037 \\
2V \sim 2.5^\circ
\]

Composition diagram for apatite, using optical properties for chlorapatite, fluorapatite, and carbonate apatite given by Deer, Howie, and Zussman (1966, p. 504-509), and assuming linear changes in optical properties with changes in composition.

Composition used in calculations:

- \( P_2O_5 \) 38.1
- \( CaO \) 53.8
- Cl 6.8
- \( CO_2 \) 2.9*

*According to Deer, Howie, and Zussman (1962) \( CO_2^- \) can substitute for a maximum of 10 wt.% of the \( PO_4^- \). Birefringence suggests about .4 of the possible 10 wt.% substitution.
APATITE (FLUORAPATITE)  
$\text{Ca}_5(\text{PO}_4)_3(\text{F})$

Diagnostic properties:

Sample #258: La Totora quartz-bearing diorite  
$N_0 = 1.636$  
$N_e = 1.632$  
$N_o - N_e = .004$

Sample #268: La Totora quartz-bearing diorite  
$N_0 = 1.636 \pm .003$  
$N_e = 1.632 \pm .003$  
$N_o - N_e = .004$

Composition used in calculations:

<table>
<thead>
<tr>
<th>Element</th>
<th>Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>CaO</td>
<td>55.5</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>42.3</td>
</tr>
<tr>
<td>F</td>
<td>3.8</td>
</tr>
<tr>
<td>D</td>
<td>3.2</td>
</tr>
</tbody>
</table>

**AUGITE**

<table>
<thead>
<tr>
<th>Element</th>
<th>Formula</th>
<th>Content</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>48.50</td>
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</tr>
<tr>
<td>TiO₂</td>
<td>1.07</td>
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</tr>
<tr>
<td>Al₂O₃</td>
<td>4.50</td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.36</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>8.18</td>
<td></td>
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<tr>
<td>MnO</td>
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<tr>
<td>CaO</td>
<td>19.32</td>
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</tr>
<tr>
<td>Na₂O</td>
<td>0.32</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>0.17</td>
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</tr>
<tr>
<td>H₂O⁺</td>
<td>1.70</td>
<td></td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.15</td>
<td></td>
</tr>
</tbody>
</table>

D = 3.35  

**El Romeral**

| Value |  
|-------|-------|
| Nₓ    | 1.673 |
| Nᵧ    | 1.678 |
| N₂    | 1.705 |
| N₂-Nₓ | 0.032 |
| 2Vₓ   | 48-49° |
| Z:C   | 42° |

pale green  

Assoc: andesite porphyry  
diorite

BIOTITE "A"

\[
\begin{align*}
\text{SiO}_2 & \quad 36.50 \\
\text{TiO}_2 & \quad 3.52 \\
\text{Al}_2\text{O}_3 & \quad 16.70 \\
\text{Fe}_2\text{O}_3 & \quad 3.27 \\
\text{FeO} & \quad 16.83 \\
\text{MnO} & \quad 0.22 \\
\text{MgO} & \quad 10.47 \\
\text{CaO} & \quad 1.32 \\
\text{Na}_2\text{O} & \quad 0.55 \\
\text{K}_2\text{O} & \quad 7.68 \\
\text{F} & \quad 0.62 \\
\text{H}_2\text{O}^+ & \quad 2.28
\end{align*}
\]

D = 3.0

\[
\begin{align*}
\text{El Romeral} \\
N_x & \quad N_x \\
N_y & \quad N_y = 1.647 \\
N_z & \quad N_z \\
N_z - N_x & \quad N_z - N_x = 0.055 \\
2V & \quad 2V \\
X & \quad X = \text{straw yellow} \\
Y & \quad Y = \text{olive brown} \\
Z & \quad Z = \text{dark olive gray}
\end{align*}
\]

Assoc: La Escoba andesite

BIOTITE "B"

<table>
<thead>
<tr>
<th>Element</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>38.04</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>3.05</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>18.11</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>3.64</td>
</tr>
<tr>
<td>FeO</td>
<td>15.36</td>
</tr>
<tr>
<td>MnO</td>
<td>0.05</td>
</tr>
<tr>
<td>MgO</td>
<td>10.52</td>
</tr>
<tr>
<td>CaO</td>
<td>1.52</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.70</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>7.36</td>
</tr>
<tr>
<td>F</td>
<td>0.07</td>
</tr>
<tr>
<td>H$_2$O$^+$</td>
<td>1.48</td>
</tr>
</tbody>
</table>

El Romeral

<table>
<thead>
<tr>
<th>Property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td>3.0</td>
</tr>
<tr>
<td>$N_x$</td>
<td>1.588</td>
</tr>
<tr>
<td>$N_y$</td>
<td>1.648</td>
</tr>
<tr>
<td>$N_z$</td>
<td>1.648</td>
</tr>
<tr>
<td>$N_z - N_x$</td>
<td>0.06</td>
</tr>
<tr>
<td>2$V_x$</td>
<td>0-5°</td>
</tr>
<tr>
<td>X</td>
<td>pale yellow brown</td>
</tr>
<tr>
<td>Y</td>
<td></td>
</tr>
<tr>
<td>Z</td>
<td>dark brown</td>
</tr>
</tbody>
</table>

Assoc: biotite dioritite

Loc: So. Sierra Nevada Batholith

BIOTITE "C"

<table>
<thead>
<tr>
<th>Component</th>
<th>Percentage</th>
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<tbody>
<tr>
<td>SiO₂</td>
<td>34.96</td>
</tr>
<tr>
<td>TiO₂</td>
<td>3.99</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>15.29</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>2.90</td>
</tr>
<tr>
<td>FeO</td>
<td>18.30</td>
</tr>
<tr>
<td>MnO</td>
<td>0.39</td>
</tr>
<tr>
<td>MgO</td>
<td>10.42</td>
</tr>
<tr>
<td>CaO</td>
<td>1.11</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.41</td>
</tr>
<tr>
<td>K₂O</td>
<td>7.99</td>
</tr>
<tr>
<td>F</td>
<td>1.17</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>3.07</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.41</td>
</tr>
</tbody>
</table>

El Romeral

D = 3.0

\[ N_x = 1.596 \]
\[ N_y = 1.646 \]
\[ N_z = 1.650 \]
\[ N_z - N_x = 0.054 \]

2V = 5°

X = pale yellow
Y = yellow brown
Z = reddish brown

Assoc: Punta de Piedra
Granodiorite

**BIOTITE "G"**

<table>
<thead>
<tr>
<th>Element</th>
<th>Formula</th>
<th>Weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>38.42</td>
<td></td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.49</td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>18.26</td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>9.69</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>6.93</td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.36</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>9.46</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>0.59</td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.67</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>7.17</td>
<td></td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>4.58</td>
<td></td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.64</td>
<td></td>
</tr>
</tbody>
</table>

**El Romeral**

- \( D = 3.0 \)
- \( N_x = 1.649 \)
- \( N_y = 1.65 \)
- \( N_z = 0^\circ \)
- \( X = \text{yellowish olive green} \)
- \( Y = \text{green} \)
- \( Z = \text{dark brownish green} \)

**Assoc:** schist

**Loc:** Japan

BIOTITE "S"

<table>
<thead>
<tr>
<th>Element</th>
<th>Formula</th>
<th>Weight (wt%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>36.22</td>
<td></td>
</tr>
<tr>
<td>TiO₂</td>
<td>3.02</td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.39</td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>3.92</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>14.41</td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.16</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>11.11</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>0.00</td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.37</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>8.97</td>
<td></td>
</tr>
<tr>
<td>F</td>
<td>0.08</td>
<td></td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>4.00</td>
<td></td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.75</td>
<td></td>
</tr>
</tbody>
</table>

El Romeral

D = 2.98
Nₓ = 1.586
Nᵧ = 1.643
N₂ = 1.643
N₂-Nᵧ = 0.057
2V = 0°

X = golden brown
Y
Z = dark brown

Assoc: oligocl.-qtz.-biotite schigerneiss
Assoc: oligoclase-biotite schigneiss

CLINOZOISITE

<table>
<thead>
<tr>
<th>Element</th>
<th>Formula</th>
<th>Weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>38.64</td>
<td></td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.14</td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>27.69</td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>7.57</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>0.14</td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.04</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>0.11</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>23.79</td>
<td></td>
</tr>
<tr>
<td>H₂O</td>
<td>1.84</td>
<td></td>
</tr>
</tbody>
</table>

El Romeral

\[ D = 3.39 \]
\[ N_x = 1.7158-1.7202 \]
\[ N_y = 1.7196-1.7329 \]
\[ N_z - N_x = 0.001 \text{ to } 0.017 \]
\[ 2V_z = 85-95^\circ(+) \]
\[ Z:C = 0-36 \]

pale bronze-yellow

Assoc. with actinolite in and around Romeral magnetite deposits.

DIOPSIDE

\[
\begin{align*}
\text{SiO}_2 & \quad 53.65 \\
\text{TiO}_2 & \quad .04 \\
\text{Al}_2\text{O}_3 & \quad 1.89 \\
\text{Fe}_2\text{O}_3 & \quad .58 \\
\text{FeO} & \quad 1.53 \\
\text{MnO} & \quad .09 \\
\text{MgO} & \quad 16.23 \\
\text{CaO} & \quad 25.30 \\
\text{Na}_2\text{O} & \quad .19 \\
\text{H}_2\text{O}^+ & \quad .47 \\
\text{H}_2\text{O}^- & \quad .07
\end{align*}
\]

\[
\begin{align*}
\text{El Romeral} \\
D & = 3.28 \\
N_x & = 1.665 \\
N_y & = 1.672 \\
N_z & = 1.695 \\
N_z - N_x & = .03 \\
2V_z & = 57^\circ \\
Z:C & = 39
\end{align*}
\]

very pale yellowish green

\#195 dioritized quartzite
\#69 actinolized andesite porph.

EPIDOTE

\[\begin{align*}
\text{SiO}_2 & \quad 37.60 \\
\text{TiO}_2 & \quad .05 \\
\text{Al}_2\text{O}_3 & \quad 25.28 \\
\text{Fe}_2\text{O}_3 & \quad 10.83 \\
\text{FeO} & \quad 1.55 \\
\text{MnO} & \quad .07 \\
\text{MgO} & \quad .29 \\
\text{CaO} & \quad 23.44 \\
\text{H}_2\text{O}^{+} & \quad 1.88 \\
\text{H}_2\text{O}^{-} & \quad .03 \\
\end{align*}\]

El Romeral

\[\begin{align*}
D & = 3.37 \\
N_x & = 1.723 \\
N_y & = 1.739 \\
N_z & = 1.751 \\
N_z - N_x & = .029 \\
2V_x & = 60^\circ (-) \\
N_x & = 1.72 \\
N_z & = 1.75 \\
N_z - N_x & = .026-.035 \\
2V_x & = 70^\circ (-) \\
\end{align*}\]

gold-green to yellow-green

\#1335 epidote vein in andesite porphyry.

\#69 epidote veinlet cutting actinolite and clinozoisite veinlets in actinolized andesite porphyry.

average of \#6, \#7, and \#8.
FERROHORNBLende
(hornblende b)

\[
\begin{align*}
\text{SiO}_2 & \quad 48.71 \\
\text{TiO}_2 & \quad 0.32 \\
\text{Al}_2\text{O}_3 & \quad 9.48 \\
\text{Fe}_2\text{O}_3 & \quad 2.33 \\
\text{FeO} & \quad 9.12 \\
\text{MnO} & \quad 0.23 \\
\text{MgO} & \quad 14.43 \\
\text{CaO} & \quad 11.93 \\
\text{Na}_2\text{O} & \quad 1.16 \\
\text{K}_2\text{O} & \quad 0.15 \\
\text{H}_2\text{O}^+ & \quad 1.83 \\
\text{F} & \quad 0.23
\end{align*}
\]

El Romeral #280, #1023

\[
\begin{align*}
D & \quad = 3.12 \\
N_x & \quad = 1.638 \\
N_y & \quad = 1.652 \\
N_z & \quad = 1.661 \\
N_z - N_x & \quad = 0.023 \\
2V_x & \quad = 78^\circ \\
Z:C & \quad = 18^\circ
\end{align*}
\]

\[
\begin{align*}
N_x & \quad = 1.640 \\
N_y & \quad = 1.651 \\
N_z & \quad = 1.660 \\
N_z - N_x & \quad = 0.020 \\
2V_x & \quad = 75^\circ \\
Z:C & \quad = 17.5^\circ
\end{align*}
\]

X = light greenish yellow  
Y = medium olive green  
Z = dark yellowish green

Assoc: Punta de Piedra granodiorite

<table>
<thead>
<tr>
<th>Component</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>42.89</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>39.99</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.36</td>
</tr>
<tr>
<td>CaO</td>
<td>0.23</td>
</tr>
<tr>
<td>Na₂O</td>
<td>16.27</td>
</tr>
<tr>
<td>K₂O</td>
<td>6.57</td>
</tr>
</tbody>
</table>

El Romeral Sample #1432

\[ D = 2.6 \]
\[ N_0 = 1.545 \]
\[ N_e = 1.541 \]
\[ N_0 - N_e = 0.004 \]

Associated Minerals:
- biotite nepheline syenite
- late biotite-arf edsonite-nepheline-bearing dioritite

Location:
Blue Mtn., Metheun Township

Reference:
Deer, Howie, and Zussman, 1962
Vol. IV, p. 243. (analysis #12, optics from #10).
PENNINITE

<table>
<thead>
<tr>
<th>Element</th>
<th>Composition</th>
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</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>33.83</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>12.95</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>2.25</td>
</tr>
<tr>
<td>FeO</td>
<td>3.02</td>
</tr>
<tr>
<td>MgO</td>
<td>34.94</td>
</tr>
<tr>
<td>H₂O</td>
<td>13.11</td>
</tr>
</tbody>
</table>

\[ D = 2.8 \]

El Romeral

\[ N_y = 1.56-1.59 \]

\[ N_z - N_x = 0.002-0.004 \]

(+ and (-) varieties,

color: green

extinction: some anomalous blue; some anomalous red-brown

Reference:

Deer, Howie, and Zussman, 1962,
**SALITE**

\[
\begin{align*}
\text{SiO}_2 & \quad 50.74 \\
\text{TiO}_2 & \quad 0.94 \\
\text{Al}_2\text{O}_3 & \quad 2.98 \\
\text{Fe}_2\text{O}_3 & \quad 2.37 \\
\text{FeO} & \quad 10.04 \\
\text{MnO} & \quad 0.17 \\
\text{MgO} & \quad 14.24 \\
\text{CaO} & \quad 17.88 \\
\text{Na}_2\text{O} & \quad 0.67 \\
\text{H}_2\text{O}^+ & \quad 0.17 \\
\text{H}_2\text{O}^- & \quad 0.03 \\
\end{align*}
\]

El Romeral

\[
\begin{align*}
D & \quad = \ 3.4 \\
N_y & \quad = \ 1.704 \\
N_y & \quad = \ 1.70 \\
2V_z & \quad = \ 56.5^\circ \\
Z:C & \quad = \ 44^\circ \\
2V_z & \quad = \ 60^\circ \\
Z:C & \quad = \ 42^\circ \\
\end{align*}
\]

Assoc: Punta de Piedra granodiorite

Reference:

SCAPOLITE
(marialite)

<table>
<thead>
<tr>
<th>Element</th>
<th>Composition</th>
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<tbody>
<tr>
<td>SiO₂</td>
<td>59.60</td>
</tr>
<tr>
<td>TiO₂</td>
<td>--</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>19.87</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.01</td>
</tr>
<tr>
<td>MgO</td>
<td>0.21</td>
</tr>
<tr>
<td>CaO</td>
<td>3.42</td>
</tr>
<tr>
<td>Na₂O</td>
<td>10.66</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.47</td>
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<tr>
<td>H₂O⁺</td>
<td>0.66</td>
</tr>
<tr>
<td>Cl⁻</td>
<td>3.23</td>
</tr>
</tbody>
</table>

El Romeral

<table>
<thead>
<tr>
<th>Property</th>
<th>Value</th>
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</thead>
<tbody>
<tr>
<td>D</td>
<td>2.6</td>
</tr>
<tr>
<td>N₀</td>
<td>1.537</td>
</tr>
<tr>
<td>Nₑ</td>
<td>1.530</td>
</tr>
<tr>
<td>N₀-Nₑ</td>
<td>0.007</td>
</tr>
<tr>
<td>uniax</td>
<td>(-)</td>
</tr>
</tbody>
</table>

Assoc: #69 actinolitized andesite porphyry

Reference:
Deer, Howie, and Zussman,
TOURMALINE
(schorlize)

<table>
<thead>
<tr>
<th>Chemical</th>
<th>Composition</th>
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</thead>
<tbody>
<tr>
<td>( \text{SiO}_2 )</td>
<td>34.91</td>
</tr>
<tr>
<td>( \text{TiO}_2 )</td>
<td>0.17</td>
</tr>
<tr>
<td>( \text{B}_2\text{O}_3 )</td>
<td>9.53</td>
</tr>
<tr>
<td>( \text{Al}_2\text{O}_3 )</td>
<td>32.29</td>
</tr>
<tr>
<td>( \text{Fe}_2\text{O}_3 )</td>
<td>0.64</td>
</tr>
<tr>
<td>( \text{FeO} )</td>
<td>12.57</td>
</tr>
<tr>
<td>( \text{MnO} )</td>
<td>0.28</td>
</tr>
<tr>
<td>( \text{MgO} )</td>
<td>2.30</td>
</tr>
<tr>
<td>( \text{CaO} )</td>
<td>1.15</td>
</tr>
<tr>
<td>( \text{Na}_2\text{O} )</td>
<td>2.29</td>
</tr>
<tr>
<td>( \text{K}_2\text{O} )</td>
<td>0.16</td>
</tr>
<tr>
<td>( \text{Li}_2\text{O} )</td>
<td>0.23</td>
</tr>
<tr>
<td>( \text{F} )</td>
<td>0.03</td>
</tr>
<tr>
<td>( \text{H}_2\text{O}^+ )</td>
<td>3.67</td>
</tr>
<tr>
<td>( \text{H}_2\text{O}^- )</td>
<td>0.02</td>
</tr>
</tbody>
</table>

Average of #6, #7, and #8.

\( \text{N}_0 = 1.676 \)
\( \text{N}_e = 1.648 \)
\( \text{N}_0 - \text{N}_e = 0.028 \)

El Romeral

\( \text{N}_0 = 1.677-1.678 \)
\( \text{N}_e = 1.648-1.650 \)
\( \text{N}_0 - \text{N}_e = 0.028-0.029 \)

\( \text{X} = \) pale brown (with bluish, pinkish, and greenish patches).
\( \text{Z} = \) dark gray (with bluish, greenish, pinkish, and brownish patches).
<table>
<thead>
<tr>
<th>Mineral</th>
<th>Identification Method</th>
<th>Reference for Identification and/or Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>calcite:</td>
<td>optical;</td>
<td>Heinrich, Tröger, ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>dickite:</td>
<td>optical, X-ray;</td>
<td>D, H, and Z*, V. 5, p. 119, #5 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>goethite:</td>
<td>optical;</td>
<td>Uytenbogaardt and Burke (1971) composition from formula ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>hematite:</td>
<td>optical, X-ray;</td>
<td>D, H, and Z, V. 5, p. 29, #1 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>ilmenite:</td>
<td>optical, X-ray;</td>
<td>D, H, and Z, V. 3, p. 232, #5 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>kaolinite:</td>
<td>X-ray;</td>
<td>ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>magnetite:</td>
<td>optical, X-ray;</td>
<td>Uytenbogaardt and Burke (1971) ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>montmorillonite:</td>
<td>X-ray;</td>
<td>D, H, and Z, V. 3, p. 29, #1 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>muscovite:</td>
<td>optical, X-ray;</td>
<td>D, H, and Z, V. 3, p. 233, #10 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>nontronite:</td>
<td>optical;</td>
<td>D, H, and Z, V. 3, p. 16, #1 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>orthoclase:</td>
<td>optical;</td>
<td>D, H, and Z, V. 3, p. 16, #1 ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>plagioclases:</td>
<td>optical;</td>
<td>chemical compositions calculated from end-member compositions and optically determined An content</td>
</tr>
<tr>
<td>pseudobrookite:</td>
<td>optical;</td>
<td>composition from formula</td>
</tr>
<tr>
<td>pyrophyllite:</td>
<td>X-ray;</td>
<td>ASTM catalogue of diffraction patterns</td>
</tr>
<tr>
<td>quartz:</td>
<td>optical;</td>
<td>composition from formula</td>
</tr>
<tr>
<td>talc:</td>
<td>optical;</td>
<td>Heinrich, Tröger, Winchell and Winchell</td>
</tr>
<tr>
<td>titanomagnetite:</td>
<td>optical, X-ray;</td>
<td>Uytenbogaardt and Burke (1961), Lindsley (1965)</td>
</tr>
</tbody>
</table>

*Deer, Howie, and Zussman (1962) is abbreviated D, H, and Z.*
APPENDIX D

COMPUTER PROGRAM

by JAMES L. KRUMHANSL
Appendix D

A Computer Program to Calculate Chemical Compositions of Rocks

by James L. Krumhansl
Graduate Student
School of Earth Sciences
Department of Geology
Stanford University
1972

This computer program is written in ALGOL W computer language.

The data required for the calculation are:

1. chemical compositions of the constituent minerals, and albite contents of the plagioclases,
2. volume percentages of each of the constituent minerals.

The program is listed on pages 398 through 400.

Sample mineral-composition data cards are shown on page 401.
Sample modal rock-composition data cards are shown on page 402.
A sample calculation is shown on pages 403 through 408.

The readout lists the following information:

1. chemical composition of each mineral read in,
2. wt. % of each mineral,
3. wt. % of each oxide in the rock due to each mineral,
4. total wt. % of each oxide,
5. computed density,
6. grams of each oxide due to each mineral per cc. of rock,
7. total grams of each oxide per cc. of rock.
BEGIN
BEGINT
INTEGER NOM, NCC, CT1, CI2, CTAB;
REAL WSI, TOTALW, PCTAB, CAO, NA2O, AL2O3, SI2O5, COMPAb, K2O, RHOJ;
REAL ARRAY SGAB(I..11);
STRING(IO) WS2;
REAL ARRAY NAME(I..NOM);
REAL ARRAY COMPDI(I..NOM, I..NOC);
REAL ARRAY PCT(I..NOM, I..NOC);
FOR I:=1 UNTIL NOM DO FOR J:=1 UNTIL NCC DO READONICOMPD(I, J);
FOR I:=1 UNTIL NCC DO WRITE(NAME(I), COMPDI(I, J));
FOR I:=1 UNTIL NOM DO BEGIN WRITE(NAME(I)); FOR J:=1 UNTIL NCC DO WRITE(PCT(I, J)); END;
READON(SAMPLENC);
WHILE SAMPLENO(C|7) NOT FINSH DO BEGIN
READON(LITHOLOGY);
FOR I:=1 UNTIL NOM DO BEGIN
WRITE(PCT(I, J)); FOR J:=1 UNTIL NCC DO WRITEONICOMPD(I, J); END;
READON(SAMPLENC);
END;
FOR I:=1 UNTIL NOM DO BEGIN WTHODE(1) := 0;
FOR J:=1 UNTIL NCC DO ANS(I, J) := 0; END;

WHILE [SAMPLENO[0,16]="SAMPLE"] AND [SAMPLENO[0,16]="FINS"] DO BEGIN
CT1 := 1; REAON(W51);
COMMENT WSI CONTAINS THE VOLUME % OF THE MINERAL IN THE ROCK
CT1 CONTAINS THE MINERAL ADDRESS;
WHILE [CT1 <= NOM+1] AND [SAMPLENO =~ NAME(CT1)] DO CT1 := CT1+1;

IF SAMPLENO[0,11] = "PLAGIOCLASE" THEN BEGIN
READON(PCTAB);
CTAB := 1;
WHILE [PCTAB > 10] AND (CTAB < 11) DO BEGIN
PCTAB := PCTAB-10; CTAB := CTAB+1; END;

DENSIY(CT1) := ([SGAB(CTAB) - SGAB(CTAB+1)]* [PCTAB]/100) + SGAB(CTAB);

COMMENT DENSITY IS NOW COMPUTED;
PCTAB := ([10*(CTAB-.95)*PCTAB]/100];

COMPAB := ([22.99+.96+39.10*0.05+23.925]*[PCTAB] + ([27.822]*[1-PCTAB]);
K2O := ([1.95]*[30.99]*PCTAB)/COMPAB;
AL2O3 := ([50.9*][1-PCTAB])/(COMPAB);
SIO2 := ([180.27]*[1-PCTAB] + [120.18]*[1-PCTAB])/(COMPAB);
NA2O := NA2O*100;
K2O := K2O*100;
AL2O3 := AL2O3*100;
SIO2 := SIO2*100;
CTAB := 1;
WHILE [CTAB < NOM+1] AND [COMP(CTAB) =~ "NA2O"] DO CTAB := CTAB+1;
PCT(CT1,CTAB) := NA2O;
CTAB := 1;
WHILE [CTAB < NOM+1] AND [COMP(CTAB) =~ "K2O"] DO CTAB := CTAB+1;
PCT(CT1,CTAB) := K2O;

CTAB := 1;
WHILE [CTAB < NOM+1] AND [COMP(CTAB) =~ "AL2O3"] DO CTAB := CTAB+1;
PCT(CT1,CTAB) := AL2O3;

CTAB := 1;
WHILE [CTAB < NOM+1] AND [COMP(CTAB) =~ "CAO"] DO CTAB := CTAB+1;
PCT(CT1,CTAB) := CAO;

CTAB := 1;
WHILE [CTAB < NOM+1] AND [COMP(CTAB) =~ "SIO2"] DO CTAB := CTAB+1;
PCT(CT1,CTAB) := SIO2;

END;
IF CT1 = NOM+1 THEN GO TO GOOF ELSE
WTMODE(CT1) := [WSI] * [DENSIY(CT1)];

COMMENT WTMODE CONTAINS [XOF矿物] [DENSIY];
READON(SAMPLENO);
TOTALW := TOTALW + WTMODE(CT1); END;

COMMENT NOW WE HAVE [XOF EACH MINERAL] [DENSIY] IN WTMODE AND THE
NEXT SAMPLE NO IN SAMPLENO AND SUM*DENSIY IN TOTALW. THE NEXT STEP...

The format requires that these constituets be designated in exactly this way:

"NA2O\]^"  
"AL2O3\]^"  
"K2O\]^"  
"SIO2\]^"  
"CAO\]^"

Other constituents can be designated by any symbol convenient to the user.
Puts wt % listed each mineral;

For i := 1 until ncm do begin
  if wtmode(i) > 10^-6 then begin
    wtmode[i] := (wtmode[i])/(totalw);
    write(name[i],"wt-percent",wtmode[i]*100);
  end;
end;

For i := 1 until ncm do begin
  if wtmode(i) > 10^-6 then begin
    if pct(i,j) > 10^-6 then ans(i,j) := pct(i,j)*wtmode(i)/100;
  end;
end;

Comment this creates a matrix of % oxide in mineral wt & mineral;

ctl := 0;

while ctl < ncm do begin
  ctl := ctl+1;
end;

if totalw > 10^-6 then
  write("total of each oxide");

write("gram of each oxide due to each mineral");

write("compute rock density is", rho);

write("total of each oxide per cc of rock");

write("total grams of each oxide per cc of rock");

write("total grams of each oxide per cc or rock");

for j := 1 until ncc do begin
  totalw := 0;
  for i := 1 until ncm do totalw := totalw + ans(i,j);
  write(compd(j), totalw*100);
end;

write("compute rock density is", rho);

write("total of each oxide per cc or rock");

write("total grams of each oxide per cc or rock");

write("total grams of each oxide per cc or rock");

for j := 1 until ncc do begin
  totalw := 0;
  for i := 1 until ncm do totalw := totalw + ans(i,j);
  write(compd(i), totalw);
end;

end;

execute options: debug,1 time=15 seconds pages=40

001.32 seconds in compilation, (06704, 02044) bytes of code generated.
Sample modal rock composition data cards.
<p>| QTZ  | 100.0000 | SiO₂  | 0.0000 | TiO₂  | 0.0000 | Al₂O₃   | 0.0000 | V₂O₅  | 0.0000 | Fe₂O₃  | 0.0000 | FeO    | 0.0000 | MnO    | 0.0000 | MgO    | 0.0000 |
|------|----------|-------|--------|-------|--------|---------|--------|-------|--------|---------|--------|--------|--------|--------|--------|--------|
| MT   | 0        | 0     | 0      | 0     | 0      | 68.97000 | 31.03000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| HEM  | 0        | 0     | 0      | 0     | 0      | 100.0000 | 0.0000  | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| ILM  | 0        | 48.10001 | 0     | 0.180000 | 0.085000 | 36.92000 | 0.240000 | 0.420000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| TIMT | 0        | 3.00000 | 0     | 0.0000  | 0.0000  | 65.97000 | 31.03000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| PSDOB| 0        | 33.35001 | 0     | 0.0000  | 0.0000  | 66.64999 | 0.0000  | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| SPN  | 31.27000 | 40.42999 | 0     | 0.510000 | 0.350000 | 0.840000 | 0.0500000 | 0.07999998 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| RTL  | 0        | 100.000 | 0     | 0.0000  | 0.0000  | 0.0000  | 0.0000  | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| PY   | 0        | 0      | 0     | 0.0000  | 0.0000  | 0.0000  | 0.0000  | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |
| CP   | 0        | 0      | 0     | 0.0000  | 0.0000  | 46.49001 | 53.49001 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 | 0.0000 |</p>
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**Note:** The table contains values that are likely numerical data, possibly representing measurements or financial figures.
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WEIGHT % OF EACH OXIDE IN ROCK DUE TO EACH MINERAL

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**Plagioclase IR**

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**Plagioclase R**

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<tr>
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<tr>
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<tr>
<td>K2</td>
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**Plagioclase U**

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</tr>
<tr>
<td>CA</td>
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<tr>
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<tr>
<td>K2</td>
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**Total of Each Oxide**

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<tr>
<td>V2</td>
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<tr>
<td>FE2</td>
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<tr>
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**Computed Rock Density**

- 2.854624/100.02
- Sum

**Grams of Each Oxide Due to Each Mineral**

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<td>Substance</td>
<td>Total Grams of Each Cxide per cc or Rock</td>
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<td>-----------------------------------------</td>
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APPENDIX E

ERROR ANALYSIS,
CALCULATED CHEMICAL COMPOSITIONS
ERROR ANALYSIS

CALCULATED CHEMICAL COMPOSITIONS

<table>
<thead>
<tr>
<th>Element</th>
<th>Wt. %*</th>
<th>Mean difference between calculated compositions and chemical analyses (ten samples)</th>
<th>Relative Error**</th>
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<tbody>
<tr>
<td>SiO₂</td>
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<td>+ 0.3%</td>
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<td>TiO₂</td>
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<tr>
<td>Al₂O₃</td>
<td>-0.40</td>
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<td>- 3.0%</td>
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<tr>
<td>Fe₂O₃</td>
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<td></td>
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<tr>
<td>FeO</td>
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<td>ΣFe</td>
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<td>Cu</td>
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<td>+</td>
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<td>H₂O⁻</td>
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<td>-167%</td>
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* Positive sign indicates calculated value is larger than analytical value.

** Percent relative error is the difference between the calculated and "true" values divided by the "true" value. This column gives the mean of 10 such relative error values for each constituent.
APPENDIX F

SPECTROGRAPHIC ANALYSES AND
UNIT-CELL MEASUREMENTS ON
Fe-Ti OXIDE MINERALS OF EL ROMERAL
SPECTROGRAPHIC ANALYSES OF ROMERAL MAGNETITES

Values reported in parts per million, except where noted otherwise, to the nearest number in the series, 1, 1.5, 2, 3, 5, 7, etc.

<table>
<thead>
<tr>
<th>Element</th>
<th>#1420 Magnetite from IM-3 Intramineral Dioritite</th>
<th>#521 Ore magnetite from the west side of the Main Orebody</th>
<th>#962 Ore magnetite from the northeast part of the Main Orebody</th>
<th>#1139 Vein magnetite from an actinolite-magnetite-apatite vein</th>
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</thead>
<tbody>
<tr>
<td>Fe</td>
<td>&gt;20%</td>
<td>&gt;20%</td>
<td>&gt;20%</td>
<td>&gt;20%</td>
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<td>1%</td>
<td>0.5%</td>
<td>1%</td>
</tr>
<tr>
<td>Mg</td>
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<td>0.1%</td>
<td>0.2%</td>
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<td>&lt;500</td>
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</table>

These analyses were done by Skyline Labs, Inc., 12090 W. 50th Place, Wheat Ridge, Colorado.
Charles E. Thompson, Chief Chemist
APPENDIX G

MICROPROBE ANALYSIS OF AMPHIBOLE

by Jane E. Pike
Figure F-1

Unit-cell dimensions of magnetites from El Romeral and El Laco, plotted on curves for variation of unit-cell dimensions with ulvöspinel content by Lindsley (1965). The 15 mole percent ulvöspinel content suggested for magnetite from IM-3 diorite is much higher than that indicated by the 1500–2000 ppm Ti detected by spectrographic analysis. The IM-3 magnetite lattice may be expanded by substitution of Ca$^{2+}$ for Fe$^{2+}$. Unit cells of ore and vein magnetites are in generally good agreement with low Ti contents as measured by spectrographic analysis.
Figure F-2

Unit-cell dimensions of ilmenite-hematite solid solution from Romeral porphyritic meladiorite plotted on curves for variation of unit-cell dimension with variation in solid-solution composition (after Lindsley, 1965).
Jane E. Pike generously took the time and trouble to do six microprobe analyses of amphiboles from actinolitized phyllites west of the Main Romeral orebody. She did the analyses on the microprobe at the Stanford School of Earth Sciences, using a chemically analyzed amphibole as a standard. Mean results of the six analyses on sample #335 are given in Table G-1.

**TABLE G-1**

Mean Composition of Actinolite from Felty Actinolitized Phyllite (sample #335)

<table>
<thead>
<tr>
<th>Element</th>
<th>Weight Percent</th>
<th>Atomic Percent</th>
<th>Atoms per Formula Weight Assuming 24O</th>
<th>Oxide</th>
<th>Weight Percent</th>
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<tbody>
<tr>
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<td>54.6</td>
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<td>0.1</td>
<td>tr (0.02)</td>
<td>TiO₂</td>
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</tr>
<tr>
<td>Al</td>
<td>1.9</td>
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<td>0.5</td>
<td>Al₂O₃</td>
<td>3.6</td>
</tr>
<tr>
<td>Fe</td>
<td>6.9</td>
<td>2.6</td>
<td>1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mn</td>
<td>0.1</td>
<td>tr (0.03)</td>
<td>tr (0.01)</td>
<td>MnO</td>
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<tr>
<td>Mg</td>
<td>5.3</td>
<td>5.2</td>
<td>1.9</td>
<td>MgO</td>
<td>8.8</td>
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<td>Ca</td>
<td>8.8</td>
<td>4.6</td>
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<td>CaO</td>
<td>12.3</td>
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<tr>
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<td>0.5</td>
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<tr>
<td>O</td>
<td>50.1</td>
<td>66.4</td>
<td>24</td>
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</tbody>
</table>

(0 determined by difference)
Discussion: Atomic percent oxygen in a "stoichiometric" amphibole should be about 57 rather than 66 percent. Because oxygen is determined by difference, the fact that oxygen is high implies that some other element or elements are low.
Discussion: Atomic percent oxygen in a "stoichiometric" amphibole should be about 57 rather than 66 percent, (Jane E. Pike, personal communication, 1973). Because oxygen is determined by difference, the fact that oxygen is high implies that some other element or elements are low.

The general structural formula for amphiboles is:

$$W_{0.1}X_2Y_5(Z_8O_{22})(O,H,F)_2.$$ 

Substituting the microprobe data into the structural formula gives:

$$(K_{0.2})(Ca_{1.7}Na_{0.2}O_{0.1})(Mg_{1.9}Ti_{0.02}Mn_{0.01}Fe_{1.0}O_{2.07})$$

$$(Si_{7.0}Al_{0.5}O_{0.5})_{22}(O,H,F)_2.$$ 

Because the microprobe does not distinguish between the valence states of multivalent elements, the structural formula cannot be electrically balanced with the available data. Because the only large shortage of atoms is in the Y group, which should total 5 but only totals 2.93, and because Mg and Fe are the major constituents of the Y group, it would appear that the microprobe values for Mg and Fe are low. Comparison of the structural formula based on microprobe data with published structural formulae for chemically analyzed amphiboles with similar optical properties suggests that the microprobe values for Ca and Si may also be somewhat low, as shown in Table G-2.
### TABLE G-2

Numbers of Ions Based on $2_4$ (O, OH, F)

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Si</strong></td>
<td>7.741</td>
<td>7.826</td>
<td>7.0</td>
</tr>
<tr>
<td><strong>Al</strong></td>
<td>0.259</td>
<td>0.174</td>
<td>0.5</td>
</tr>
<tr>
<td><strong>Al</strong></td>
<td>0.191</td>
<td>0.068</td>
<td>0.02</td>
</tr>
<tr>
<td><strong>Ti</strong></td>
<td>0.031</td>
<td>--</td>
<td>0.02</td>
</tr>
<tr>
<td><strong>Fe$^{3+}$</strong></td>
<td>0.118</td>
<td>0.256</td>
<td>2.93</td>
</tr>
<tr>
<td><strong>Fe$^{2+}$</strong></td>
<td>1.394</td>
<td>4.051</td>
<td>2.93</td>
</tr>
<tr>
<td><strong>$#Fe$</strong></td>
<td>1.512</td>
<td>4.307</td>
<td>2.93</td>
</tr>
<tr>
<td><strong>Cr</strong></td>
<td>--</td>
<td>--</td>
<td>1.0</td>
</tr>
<tr>
<td><strong>Mg</strong></td>
<td>3.250</td>
<td>0.156</td>
<td>1.9</td>
</tr>
<tr>
<td><strong>Mn</strong></td>
<td>--</td>
<td>0.468</td>
<td>0.01</td>
</tr>
<tr>
<td><strong>Na</strong></td>
<td>0.149</td>
<td>--</td>
<td>0.2</td>
</tr>
<tr>
<td><strong>Ca</strong></td>
<td>1.902</td>
<td>1.825</td>
<td>2.1</td>
</tr>
<tr>
<td><strong>K</strong></td>
<td>0.022</td>
<td>--</td>
<td>0.2</td>
</tr>
<tr>
<td><strong>OH</strong></td>
<td>1.938</td>
<td>2.255</td>
<td>2.26</td>
</tr>
<tr>
<td><strong>F</strong></td>
<td>--</td>
<td>--</td>
<td>2.1</td>
</tr>
</tbody>
</table>
Assuming that the microprobe value for Al is accurate (0.5 atoms of Al per formula weight), then the felty green amphibolitized phyllite contains amphibole that is transitional between actinolite and hornblende. According to Deer, Howie, and Zussman (1962) the division between actinolite and hornblende-series amphiboles is arbitrarily taken at Si\(\text{\(\rightarrow\)}\)Al replacement of 0.5 atoms per formula weight.

According to experimental work by Liou (1972), amphibole compositions near the actinolite-hornblende transition imply temperatures of formation near 550°C.