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MAGMATIC IRON ORES AND ASSOCIATED MINERALISATION: EXAMPLES FROM THE CHILEAN HIGH ANDES AND COASTAL CORDILLERA

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Abstract - The El Laco magnetite-apatite ore deposits in the Andean Cordillera of northern Chile occur as massive, tabular bodies, as stratified, pyroclastic ores, and as crosscutting dykes and vein complexes. The ore deposits and surrounding volcanic rocks, mainly andesites, are Plio-Pleistocene in age and preserve many of their original volcanic textures and structures. All the field and laboratory data are consistent with an origin by eruption and shallow intrusion of a high-temperature, volatile-rich, iron-oxide magma. A number of other iron-oxide-apatite deposits of Cenozoic age in the Andean Cordillera, and a belt of Cretaceous iron deposits in the Coastal Cordillera of Chile also have features that suggest a magmatic origin. Associated with these magmatic ore deposits are economic and sub-economic concentrations of Cu, Au, U, and REE.

Introduction

The discovery of the enormous Cu-Au-U-REE hematite breccias at Olympic Dam in Australia (Roberts and Hudson, 1983; Oreskes and Einaudi, 1990, 1992; Haynes et al., 1995) has made iron-oxide-apatite deposits of considerable economic interest for commodities other than iron. Indeed, many deposits are associated with nearby Cu, Au, U, or REE mineralisation (Bai and Yuan, 1983; Crocker, 1985; Einaudi and Oreskes, 1990; Hauck, 1990; Kisvarsanyi, 1990; Marikos et al., 1990; Vivallo et al., 1993a, 1994a, 1995a, 1995b; Espinoza et al., 1994; Förster and Jafarzadeh, 1994; Rojas and Henríquez, 1994; Foose and McLelland, 1995; Porter, 2000). Although iron-oxide-apatite deposits share many common features, it is clear that they did not all form by the same process or processes. Many deposits show evidence of formation in evolving magmatic systems: some deposits present clear magmatic features whereas others are predominantly hydrothermal in character; others still have such complicated histories or such poor exposures, that their origins are unclear.

Low-Ti, magnetite-apatite ore deposits occur associated with volcanic rocks in two belts in northern Chile (Figure 1); one of Cenozoic age in the High Andean Cordillera, and one of Cretaceous age in the Coastal Cordillera. The best preserved orebodies are the Plio-Pleistocene, El Laco deposits in the High Andes, which were first described by Park (1961), who reported iron-oxide orebodies that resembled lava flows. He suggested that the deposits formed from a volatile-rich, iron-rich magma which intruded the local volcanic sequence at shallow depth and in places erupted to the surface. This interpretation has been assumed by many subsequent investigators (Rogers, 1969; Haggerty, 1970; Frutos and Oyarzún, 1974; Henríquez and Martin, 1978; Frutos et al., 1990; Nyström and Henríquez, 1994, 1995; Naslund et al., 1997, 1998; Henríquez and Nyström, 1998; Broman et al., 1999; Nyström et al., 2002). Although not as well preserved, similar magmatic iron ore deposits occur in the High Andes of Chile at Incahuasi and Magetita Pedernales, and at numerous localities in the Cretaceous iron belt (Figure 1) (Grez et al., 1991; Henríquez et al., 1994; Travisany et al., 1995). A number of recent reports, however, have suggested that the El Laco iron-oxide deposits are hydrothermal in origin (Hildebrand, 1986; Hitzman et al., 1992; Larson and Oreskes, 1993; Larson, 1994; Rhodes and Oreskes, 1995a, 1995b, 1996, 1999; Oreskes et al., 1994, 1995; Rhodes, 1996; Sheets et al., 1997; Rhodes et al., 1997, 1999; Hitzman, 2000).

Editor's note: One of the aims of this two volume set has been to present the differing views on what constitutes a member of the "iron-oxide copper-gold" family of deposits, and how they were formed so that the reader can study the reasoning behind the varying opinions and make their own informed decision. There is a well reasoned argument that some of the important magnetite-apatite deposits are in fact magnetite lavas, including those found in districts that also host hydrothermal iron-oxide copper-gold ores. As a consequence, this group of authors, who through years of careful observation and research have supported this interpretation, were invited to contribute a paper outlining and arguing the case for their viewpoint. Their work is largely based on the El Laco deposits in Chile which are probably the best example available for such a study. The El Laco magnetite deposits fall within the Puna Belt, which also includes the Arizaro and Lindero Cu-Au Prospects described by Dow and Hitzman in this volume.

El Laco is, therefore, a critical locality. If the best-preserved, seemingly magmatic iron ore deposits in this district are hydrothermal in origin, then all similar deposits are likely to be hydrothermal as well. If, however, it can be demonstrated that iron-oxide magmas erupted at even one place on Earth, then a magmatic origin must be considered for other similar, less-well-preserved, deposits.

Although there are some features at El Laco that are clearly hydrothermal in origin, as one would expect at any volcanic centre, all of the available evidence indicates that the magnetite ore deposits were formed from iron-oxide magmas and magmatic fluids released by the magmas as they cooled and solidified. This paper will outline the field relationships at El Laco and the available analytical data, and discuss how they fit with both the magmatic and hydrothermal hypotheses. The paper will then describe the general geology of the Cretaceous iron belt, and the mineralisation associated with magmatic iron-oxide deposits in Chile.

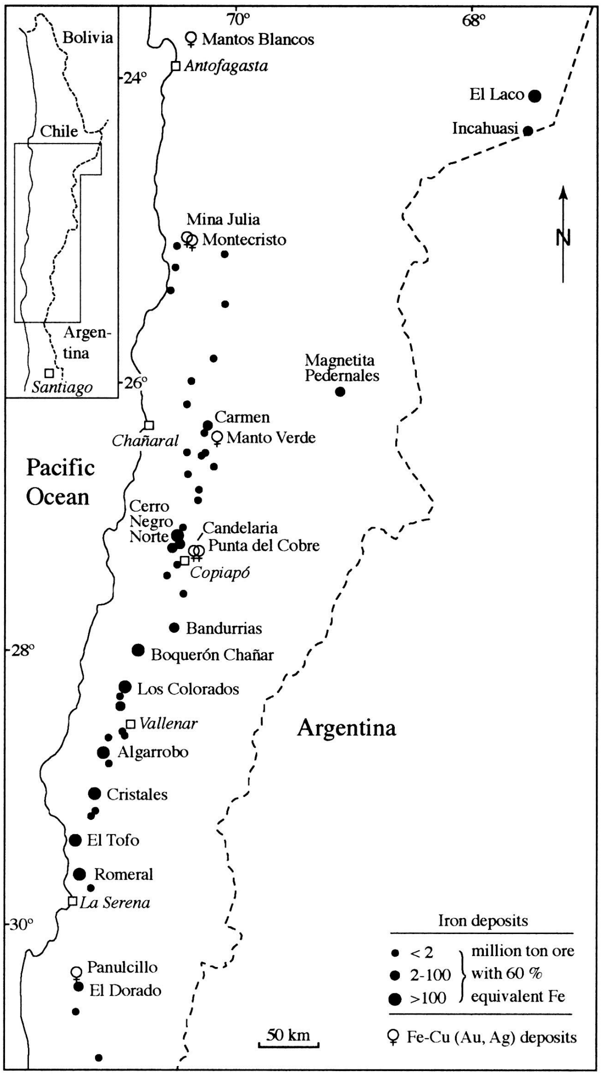

Figure 1: Magnetite-apatite ores and some associated Fe-Cu-(Au, Ag) mineralisation in northern Chile (the Toldo-Velarde deposit is situated 125 km north of Antofagasta.

Field Relationships at El Laco

Tectonic Setting

The El Laco iron deposits are exposed at altitudes between 4,700 to 5,300 m on the flanks of El Laco, a calc-alkaline volcanic complex in the Central Andes of northern Chile at 23°48' S latitude and 67°30' W longitude (Figures 1-2). Within an area of 30 km^2 there are seven deposits with total resources on the order of 500 million tons of high-grade, iron ore (Nyström and Henríquez, 1994). With the exception of the orebodies, El Laco is similar to other volcanic complexes in the area in terms of silicate lava compositions, phenocryst assemblages, general eruptive history, and degree and distribution of alteration. The ore deposits have been dated at 2.1 Ma by apatite fission track analysis (Maksaev et al., 1988). Andesitic to dacitic lavas predate and postdate the ore deposition. Fumarolic activity has continued at El Laco to the present day.

El Laco is located slightly behind the active volcanic arc, as defined by the volcanoes Láscar, Licancabur, and Putaria to the north, and Socompa, Lastarria, and Antofalla to the south. Crustal thickness beneath El Laco is ca. 60 km (Tassara, 1997), and the depth to the well-defined, Benioff zone beneath the volcano is approximately 125 km (Frutos, 1990). Calc-alkaline lavas within the Central Andes (21° to 26° S) have systematically higher $^{18}O/^{16}O$ and $^{87}Sr/^{86}Sr$ ratios than do lavas in the rest of the volcanic chain, which may be related to the thicker crust in this segment (Harmon and Hoefs, 1984). Tectonic reconstructions suggest that the crustal section below El Laco contains a Lower Paleozoic basement of marine sedimentary rocks, unconformably overlain by Mesozoic and Lower Tertiary continental sediments and Upper Tertiary volcanic rocks (Frutos, 1990). Small evaporitic basins are common in the high Altiplano around El Laco. Glacial deposits are present to the west of El Laco, and glacial erosion has removed the upper part of the San Vicente Bajo orebody.

General Features of the Deposits

The El Laco ore deposits, which are composed mostly of primary magnetite or martite pseudomorphs after magnetite, occur as pyroclastic material, lava flows and related dykes (Laco Norte, Laco Sur, and San Vicente Alto), subvolcanic bodies (San Vicente Bajo, Rodados Negros, and Laquito), and an intrusive dyke-vein network (Cristales Grandes). Due to its friable nature, the pyroclastic material is exposed only in mined areas (Laco Sur), road cuts and exploration pits. Apatite is locally abundant in the intrusions, but is only a minor accessory phase in the lavas. The ore contains variable amounts of partly to completely altered pyroxene. Hematite is a ubiquitous oxidation product after magnetite in intrusive and pyroclastic ore, and is in places primary. Dykes and veins of magnetite, ± pyroxene, and ± apatite are abundant in the massive units and in the surrounding host rocks, but comprise less than 1% of the ore volume. Many of the host rocks surrounding the deposits have been pervasively altered, and massive mounds of fumarolic sulphate and sulphur occur in the volcanic complex. Late-stage, low-temperature, quartz veins are common.

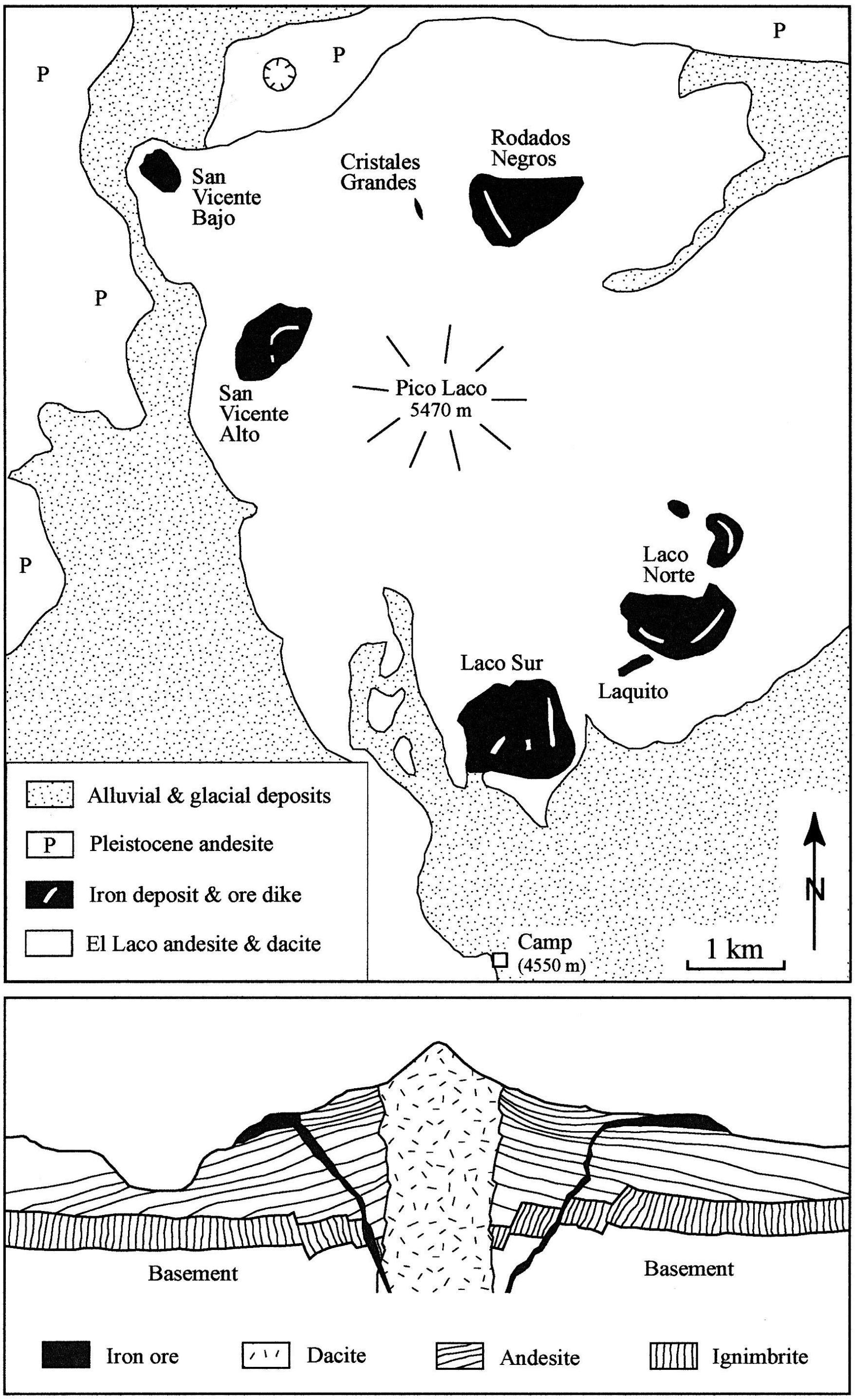

Figure 2: Geological sketch map and profile of the El volcanic complex and iron deposits (modified after Nyström and Henríquez, 1994 and Frutos and Oyarzún, 1974).

Massive Iron Ores

The term massive is used here for ore without bedding and includes ore lava, pyroclastic material and intrusive ore. Massive orebodies in excess of 100 m thick, composed of >98% iron-oxide, are present at Laco Norte, Laco Sur, San Vicente Alto, and San Vicente Bajo. Evidence for a direct magmatic origin for these massive units includes: 1) pahoehoe and aa flow surfaces; 2) contorted flow layering; 3) razor-sharp, chilled contacts; 4) vesicular textures with open vesicles; 5) friable, porous aggregates of euhedral magnetite crystals which locally pass into stratified ore ± apatite; 6) abundant gas escape tubes; and 7) coarse, spherulitic and dendritic magnetite crystals (Henríquez and Martin, 1978; Nyström and Henríquez, 1994; Naslund et al., 1997, 1998). It has been argued that these textures are relicts of an original silicate rock replaced by hydrothermal iron-oxide. Such a process would require the gradual replacement of a silicate lava or tuff while retaining their fine-scale textures. No samples showing a progression from altered andesite to massive iron-oxide (rocks with 20%, 40%, or 60% iron-oxide, for example) have been reported by any investigators. The only partial "replacement" by iron-oxide reported, is where thin (1 mm to 1 cm) crosscutting veins cement fractured and brecciated host andesites. Such vein "replacement", however, does not preserve original textures, nor is it likely to result in rock bodies 100 m thick with >98% iron-oxide. The proposed hydrothermal replacement process for El Laco (Rhodes et al., 1999), involves the replacement of initial silicate lavas by smectite, scapolite, and feldspar, followed by a second stage in which the altered lavas are replaced with coarse, massive pyroxene, followed by a final stage in which the pyroxene rock is replaced by magnetite. Such a multistage process would certainly obliterate any original textures. In fact, although in some places rocks adjacent to the massive orebodies have been completely replaced by smectite, scapolite, and feldspar, and in other places rocks have been replaced by coarse, massive pyroxene, none of the replaced rock preserves any of the volcanic textures observed in the massive iron ores.

It has been suggested that the contacts of the massive magnetite units at El Laco are gradational (Larson, 1994; Rhodes and Oreskes, 1995a; Oreskes et al., 1995) based on the fact that altered andesite talus gradually changes to iron-ore talus below the iron-ore outcrops. Where exposed, however, the contacts of the massive magnetite units are razor-sharp, and in many cases a fine-grained chilled margin is present in the magnetite adjacent to the contact (Figure 3). Within the Laco Sur deposit there are a number of large (>1m diameter) xenoliths of altered andesite. There is no iron mineralisation within the xenoliths, the surrounding massive iron-oxide ore is chilled against the xenoliths, and the contact between rock with <5% iron-oxide and rock with >99% iron-oxide is razor-sharp. Similar razor-sharp, chilled contacts are present in vesicular dykes below the Laco Sur orebody. At the base of San Vicente Bajo thin veins of magnetite crosscut the host andesite for tens of metres below the contact of the orebody, and the adjacent andesite is badly altered, and in places brecciated. The contact, however, is marked by a change from altered andesite to massive iron ore over a distance of <1mm (see Figures 13B-C in Rhodes et al., 1999). Semi-horizontal fracture patterns in the andesite below the contact and in the ore above the contact are similar, and give the rocks a crude layered appearance. It has been suggested that this "layered morphology" in the ore was inherited from an original andesite protolith (Rhodes, 1996; Rhodes et al., 1999). Regardless of how the ore deposit formed, however, these fractures are clearly a post-mineralisation feature. Had they been present at the time of mineralisation, whether by iron-oxide magmas or by hydrothermal fluids, they would have acted as channel ways for the mineralising liquids. The fractures are unmineralised.

The vesicular textures common in many of the massive magnetite units (Figure 4) are hard to reconcile with an origin by hydrothermal alteration or replacement which usually fills pores and vesicles during the initial stages of alteration. Vesicular textures are ubiquitous in the massive ores, and almost all of the vesicles are unfilled, open pore spaces. Some vesicular dykes that crosscut the massive iron-oxide bodies are so rich in open spherical vesicles that they have densities of 1.9 g/cm^3 and yet are almost pure iron-oxide. It has been suggested that the porosity in the magnetite ores is a relict of angular pore spaces in brecciated andesite protolith or of pore spaces between coarse pyroxene crystals formed during replacement of andesite by pyroxene (Rhodes et al., 1999). The iron-oxide orebodies contain both angular and rounded vesicles, but neither resemble the angular pore spaces in breccias or in the pyroxene rocks adjacent to the deposits.

Figure 3: Contact between iron ore (below) and altered andesite (above) with a fine-grained chilled margin in the ore, Laco Sur. Knife handle is 6.0 cm long.

Open, vertical pipes metres to tens of metres long, and centimetres to tens of centimetres in diameter, with circular to oval cross sections are common in the massive orebodies (Figure 5). Such features are common in cooling silicate lava flows and ignimbrites and represent internal degassing conduits. At El Laco, these degassing pipes are typically lined with coarse magnetite octahedra. In some tubes the magnetite is intergrown with coarse pyroxene and apatite, and other tubes contain fine-grained ore with vesicles and locally even bedding. The tubes appear to represent conduits along which high-temperature gases escaped from the crystallising, volatile-rich, iron-oxide magma (Henríquez and Martin, 1978). Externally derived, hydrothermal fluids would be more likely to have been transported along planar fractures than along open, vertical tubes.

Stratified and Scoriaceous Ores

Mining of the Laco Sur deposit has uncovered bedded ash layers beneath the main orebody (Figure 6). The ash which is almost pure iron-oxide in composition, is crystalline, friable, size-sorted, and contains numerous magnetite spindle bombs up to 0.5 m in length. In some layers, rounded masses of diadochite ($Fe_2PO_4SO_4OH{\cdot}6H_2O$), 2 to 10 cm in diameter are present in the bedded ash, and appear to be a primary feature. These deposits are fresh, and have not been altered by any later folding, metamorphism, or fluid interaction. Bedded ore can also be found at several places within the massive orebody. The bedding within the more massive ore bodies, is best seen where it is marked by thin layers of apatite needles (see Figures 4A-B in Nyström and Henríquez, 1994). The textural features and friable nature of the bedded ash along with the solid magnetite spindle bombs, conclusively demonstrate that it is a near-vent, volcanic, pyroclastic deposit. An origin by hydrothermal replacement of a "normal" silicate ash is precluded by the friable, porous nature of the deposit, and the lack of veining or alteration along the upper or lower contacts of the unit. While the surface venting of a hydrothermal system could produce bedded hematite fragments, the magnetite spindle bombs require the eruption of an iron-oxide magma (Henríquez and Nyström, 1998). It is extremely difficult to imagine how a hydrothermal replacement process could convert an originally silicate, near-vent facies deposit to iron-oxide and leave the bedded ash and lapilli as unconsolidated magnetite and hematite fragments, while replacing the enclosed spindle bombs with solid (but vesicular) magnetite.

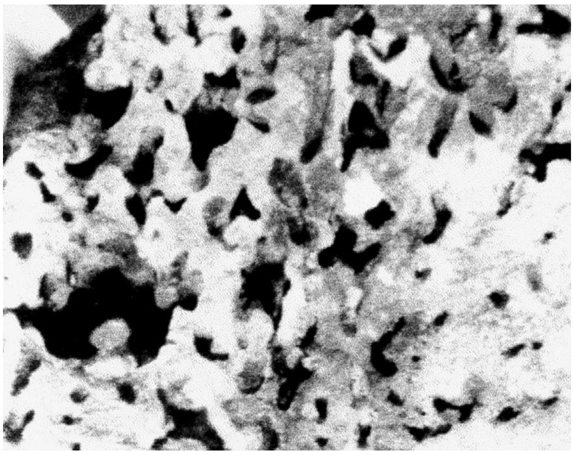

Figure 4: Vesicular texture in iron ore, Laco Sur. Width of view is ~3.3 cm.

At Laco Norte, pyroclastic iron ore is present in drill core from beneath and within the main orebody and is exposed in exploration pits north of the orebody. In the exploration pits, loose, vesicular iron-ore fragments centimetres to tens of centimetres in diameter with the appearance of basaltic scoria overlie an unaltered andesite lava flow and are overlain by a second unaltered lava flow. This scoriaceous ore appears to be formed by a strombolian-like eruption of iron-oxide magma. An origin by hydrothermal replacement would require the complete replacement of individual fragments while leaving their vesicles unfilled, the fragments un-cemented, and the lavas above and below the fragmental ore unit unaltered.

Dykes and Veins

Some of the smaller orebodies at El Laco are dykes (Rodados Negros and Laquito) or dyke-vein systems with prominent ore breccia and aggregates of magnetite, pyroxene and apatite in open spaces (Cristales Grandes). Crosscutting veins or dykes of magnetite, with pyroxene and apatite are common within and around many of the massive iron orebodies. Some veins 1 to 10 cm wide have open centres lined with magnetite crystal terminations showing clear evidence that magnetite crystals grew from the walls of the vein inward. In a few of these open veins, the central part has been filled with later, lower-temperature quartz or quartz plus apatite.

It has been suggested that the pyroxene crystallisation preceded magnetite crystallisation (Rhodes et al., 1999). Although in some places, pyroxene rock near the margins of the massive orebodies is cut by later magnetite veins, in most places pyroxene grew simultaneously with magnetite. Veins of intergrown magnetite and pyroxene cut the massive magnetite orebody at Laco Sur demonstrating that at least some of the pyroxene crystallisation postdates the main ore forming process (Broman et al., 1999). In some veins parallel layers of elongate pyroxene crystals are completely surrounded by coarse magnetite, demonstrating that in these veins magnetite crystallisation proceeded, occurred simultaneously with, and followed pyroxene crystallisation. In some other veins, magnetite ± pyroxene crystallisation alternated with magnetite ± apatite crystallisation, and in a few veins magnetite, apatite, and pyroxene clearly crystallised simultaneously. Some open veins contain perfectly formed crystal terminations of octahedral magnetite, prismatic pyroxene, and hexagonal apatite indicating that all three were in equilibrium with the liquid that filled the open space during the final stage of crystallisation. The idea that such veins represent the initial crystallisation of a layer of elongate pyroxene crystals, that the space between the pyroxene crystals was later "filled in" with magnetite, and that the parallel, elongate apatite crystals represent original pyroxene laths that were

Figure 5: Gas escape tube in iron ore, Laco Sur.

Figure 6: Magnetite spindle bomb in pyroclastic iron ore with bedding, Laco Sur.

completely replaced by apatite (Rhodes et al., 1999) is clearly incorrect.

It is suggested that "hydrothermal magnetite-quartz veins" are common in the host rocks (Larson and Oreskes, 1993; Larson, 1994; Rhodes and Oreskes, 1995a, 1995b, 1996; Oreskes et al., 1995; Rhodes et al., 1997, 1999). There are numerous veins cutting the host rock that contain an early magnetite selvage and an open or quartz-filled interior. These are, however, not "magnetite-quartz" veins. The magnetite is clearly of high-temperature origin, and the quartz is a later, lower-temperature, fracture-filling mineral. No data has been reported to suggest that magnetite and quartz were deposited in equilibrium; in fact, all the reported data indicates that they were not (see discussion of oxygen isotopes below). The texture of the quartz in hand specimen and thin section, indicates that it is a late, low-temperature mineral, unrelated to the crystallisation of the coarse magnetite crystals in the veins or to the massive magnetite orebodies. Calcite, which is present in some samples, is also a late, low-temperature mineral unrelated to the iron mineralisation.

At San Vicente Bajo, abundant magnetite dykes one to several metres wide cut sub-horizontal flow layering in the massive magnetite orebody (see Figure 13A in Rhodes et al., 1999). Previous workers attributed the close proximity of subvertical and sub-horizontal "layers" in the ore to replacement of an andesite protolith that had undergone "normal faulting" (Rhodes et al., 1999). There is no indication of any faulting at El Laco that could have rotated originally horizontal layers into a vertical position on such a large scale and over such short horizontal distances.

Alteration

Two types of alteration patterns are recognised at El Laco. The first type is restricted to the margins of the intrusive iron orebodies, and occurs as halos around them and as minerals deposited in open spaces within the extrusive ores. This first type is overprinted by hydrothermal alteration affecting large areas in the district.

Alteration assemblages seemingly related to the emplacement of the ore (Vivallo et al., 1993b, 1994b; Nyström and Henríquez ,1994; Broman et al., 1999) form metasomatic aureoles (halos) around the ore dykes and the San Vicente Bajo intrusion. The width of the aureoles (up to 20 m) varies with the size of the orebody and the permeability of the host rock. The aureoles consist of pyroxene, magnetite, quartz, and scapolite (marialite) in altered andesite lava and tuff. The contacts between the orebodies and the altered host rock are sharp at the few places where they are exposed, and in drill cores. Quartz is common in partly filled cavities and as veins within or outside the magnetite bodies, but textural relations show that the quartz invariably is younger than the iron-oxide.

Metasomatic aureoles like those around the intrusive orebodies and dyke-vein systems have not been observed at the margins of the extrusive deposits. In general, the andesitic lavas and tuffs immediately below the magnetite flows at Laco Sur and Laco Norte are hydrothermally altered. Illite, mixed-layer clays (smectite-illite) and pyrophyllite are present in the drill core, in addition to quartz, feldspar, and a few irregular magnetite veins without pyroxene fringes. In one trench on the northwest edge of the Laco Norte deposit coarse pyroxene rock is present adjacent to magnetite ore, but these samples appear to be associated with a feeder dyke for the Laco Norte flows.

Open spaces in the extrusive ore contain tridymite, quartz, amorphous silica, hematite, and small amounts or traces of Na-alunite, Fe-bearing phosphates, plagioclase (labradorite), sanidine, and rutile. These minerals (not listed according to temperature of formation) were deposited during degassing and subsequent stages of hydrothermal activity.

The El Laco deposits and associated metasomatic alteration are overprinted by hydrothermal alteration formed at low temperature that has affected large areas of the volcanic complex without a clear zonation pattern (Vivallo et al., 1993b, 1994b). It is expressed as a pervasive, strong silicification and advanced argillic alteration with destruction of the original textures, grading outwards to rocks with moderate or weak alteration. The degree and extent of the hydrothermal alteration show a good correlation with lithology and permeability, the most strongly affected being pyroclastic rocks. The silicified rocks consist of quartz, tridymite, cristobalite, alunite and jarosite, and minor amounts of K-feldspar, anatase and some unidentified iron sulphates and phosphates. Rocks affected by argillic alteration are characterised by quartz, alunite, kaolinite, gypsum, cristobalite, smectite, bassanite, pyrite and pyrophyllite.

In the less altered volcanic rocks there is only a partial replacement of pyroxene by chlorite and plagioclase by sericite and clay minerals; a hematite-epidote association is common. Locally, there are fumarolic deposits dominated by gypsum with subordinate alunite, jarosite, sulphur, cristobalite, tridymite and calcite. These exhalative deposits are restricted in extent and thickness and were deposited around several discrete centres in the district. The orebodies are crosscut by numerous veins of gypsum, which clearly postdate the ore forming process.

Analytical Data from El Laco

General Features

In terms of the reported analytical data, there do not appear to be any serious disagreements between those advocating a magmatic origin and those advocating a hydrothermal origin for El Laco. Where data have been reported by both groups, for example oxygen isotope analyses and fluid inclusion homogenisation temperatures, the results are similar. The interpretations of the results, however, are very different.

The main orebodies at El Laco are composed of almost pure magnetite or magnetite partly replaced with hematite pseudomorphs. Most ore samples are >98% iron-oxide.

The bulk ores are rich in REE with some samples containing in excess of 1% total REE-oxides. An average of 13 samples of randomly selected bulk ore from Laco Sur contained 0.33% REE-oxides (Heriberto Fortín, Compañía Minera del Pacífico internal report, 2001). The REE in the bulk ores are concentrated in apatite and Fe-phosphates. Apatite is abundant in veins, gas escape tubes, and locally in the ore, but on average makes up a very small percentage of the deposit (<1%). Pyroxene is only abundant in veins, gas escape tubes, and some contact rocks, and in most samples where it would be expected, it is only present as badly altered relict crystals. In some cases the former presence of pyroxene is indicated by empty, elongate, prismatic holes in coarse magnetite crystals lining gas escape tubes. Near the margin of some of the massive magnetite bodies, the host rock has been completely converted to a coarse pyroxene rock completely destroying any original features so that the protolith is unknown. The volume of such pyroxene rock is very small relative to the orebodies (<0.01%).

Oxygen Isotopes

It has been suggested that since the oxygen isotopes in magnetite from veins and massive units are very similar, both must be hydrothermal (Larson, 1994; Rhodes and Oreskes, 1995a, 1995b, 1996, 1999). If the veins and massive units both had hydrothermal isotope signatures, a logical conclusion might be that both were hydrothermal. However, both the veins and the massive units have igneous oxygen isotope signatures (Vivallo et al., 1994b; Rhodes and Oreskes, 1999; Nyström et al., 2002). An alternative explanation is that the veins have similar isotopic signatures as the massive units, because the veins were deposited at near-magmatic temperatures from fluids exsolved from iron-oxide magmas as they solidified. In fact, all of the available oxygen isotopic data suggest that the iron-oxide deposits are unrelated to the hydrothermal alteration at El Laco (Figure 7). Unaltered andesites and dacites at El Laco have a narrow range of oxygen isotope values ($\delta^{18}O = 7.9$, Vivallo et al., 1994b; $\delta^{18}O = 7.2$ to 7.9, Rhodes and Oreskes, 1999), while all of the hydrothermally altered samples analysed have much higher values ($\delta^{18}O = 11.8$ to 35.3, Vivallo et al., 1994b; $\delta^{18}O = 8.5$ to 24.2, Rhodes and Oreskes, 1999). Altered magnetite, on the other hand, has lower oxygen isotope values ($\delta^{18}O = 2.3$ to 3.5 in samples with 7 to 75% hematite, Nyström et al., 2002; $\delta^{18}O = -8.9$ to 3.5 in samples with 6 to 96% hematite, Rhodes and Oreskes, 1999) than does unaltered magnetite ($\delta^{18}O = 3.1$ to 4.2, Nyström et al., 2002; $\delta^{18}O = 3.5$ to 5.2, Rhodes and Oreskes, 1999). This indicates that the processes responsible for the hydrothermal alteration of the silicate lavas are unrelated to those responsible for the formation of the ore deposits or for the alteration of the ore deposits.

As described above, numerous veins cutting the host rock contain an early magnetite selvage and an open or quartz-filled interior. In contrast to the narrow range of values for magnetite ($\delta^{18}O = 3.1$ to 5.2; Rhodes and Oreskes, 1999; Nyström et al., 2002), quartz displays a wide range of oxygen isotope values ($\delta^{18}O = 11.9$ to 25.0, Vivallo et al., 1994b; $\delta^{18}O = 6.0$ to 27.9, Rhodes and Oreskes, 1999). Moreover, quartz shows wide $\delta^{18}O$ variations within individual veins (Larson, 1994; Vivallo et al., 1994b; Rhodes and Oreskes, 1995b, 1999), confirming that the magnetite and quartz are not in equilibrium, and did not precipitate from the same liquids.

The proposed hydrothermal model (Rhodes and Oreskes, 1999; Rhodes et al., 1999) would require a series of fluids with a very complicated history of oxygen isotope changes. The unaltered andesites with a narrow range of $\delta^{18}O$ (7.2 to 7.9) would be first converted to altered andesites with a range of higher $\delta^{18}O$ during propylitic alteration and Na-K metasomatism ($\delta^{18}O$ values = 8.5 to 10.1), followed by high-temperature replacement and Ca metasomatism to produce pyroxene with a narrow range of low $\delta^{18}O$ (7.1 to 8.9), followed by low-temperature replacement to produce magnetite ores with a narrow range of still lower $\delta^{18}O$ (3.5 to 5.2), followed by quartz precipitation and argillic alteration at low temperature with a wide range of higher $\delta^{18}O$ (6.0 to 27.9) and apatite replacement at low temperature with a narrow range of $\delta^{18}O$ (7.8 to 8.0), followed finally by hematite alteration at low temperature to produce very low $\delta^{18}O$ altered ores (-8.9 to 3.5) (all data from Rhodes and Oreskes, 1999). A simpler model would be to produce the andesites, magnetites, apatites, and pyroxenes at high temperature with all four related to magmas or magmatic fluids with narrow ranges of $\delta^{18}O$, to then produce the altered andesites and quartz veins during a single, post Fe-mineralisation, hydrothermal, alteration event from fluids with a range of high $\delta^{18}O$ values, and to finally produce the hematite by alteration of the magnetite by low $\delta^{18}O$ meteoric water.

The difference in oxygen isotopes between unaltered iron ores and unaltered silicate lavas is similar to what would be predicted for immiscible liquids of these compositions. Experimental studies of immiscible silicate liquids have shown that iron-rich liquids have lower oxygen isotope ratios than do iron-poor liquids (Kyser et al., 1998). A difference of 0.6 per mil was observed for immiscible liquids which differed in iron content by only 22% (42% vs. 20% Fe-oxides). For liquids that differ in iron content by 90% (98% vs. 8% Fe-oxides, for example), a difference of ~2.5 per mil might be predicted, similar to that observed for the El Laco iron ores and silicate lavas. The difference in oxygen isotopes between magnetite and pyroxene is consistent with the crystallisation of these minerals at high temperatures (Nyström et al., 2002). The observed $\Delta\,\delta^{18}O$ for pyroxene-magnetite is 1.9 to 5.4 per mil (Rhodes et al., 1999) which corresponds to crystallisation temperatures of 1090 to 540 ºC (Chiba et al., 1989).

Fluid Inclusions

Magmatic to near-magmatic fluid inclusion homogenisation temperatures have been reported for pyroxenes associated with magnetite in dykes and veins at El Laco (710 to 840 ºC, Sheets et al., 1997; Rhodes et al., 1997; Rhodes and Oreskes, 1999; >800 ºC, Broman et al., 1999). Lower fluid inclusion homogenisation temperatures have been reported for El Laco apatites (219 to 360 ºC, Oreskes et al., 1994; 250 to 350 ºC, Sheets et al., 1997, Rhodes et al., 1997; 150 to >600 ºC, Broman et al., 1999). The low apatite temperatures may represent: 1) a low-temperature origin for at least some of the apatite; 2) the post-crystallisation trapping of later, low-temperature fluids; or 3) annealing and/or re-equilibration of apatite as it cooled. Fission tracks in apatite anneal at 100 ºC, so it is possible that fluid inclusions in apatite also anneal at low temperatures during slow cooling. The fluid inclusions in apatite that yield the low homogenisation temperatures contain a wide range of salinities (0.2 to 59 wt% NaCl equivalent, Sheets et al., 1997, Rhodes et al., 1997) suggesting that these inclusions probably do not accurately preserve the fluids from which they formed. In some El Laco ore samples, apatite fluid inclusions (hydrous saline Na-K chloride melt inclusions) failed to homogenise at 600 ºC, and in other samples the apatite host decrepitated before fluid inclusion homogenisation was achieved (Broman et al., 1999), demonstrating that some of the fluid inclusions in apatite formed at higher temperatures. Some of the apatite in veins is intergrown with quartz and is clearly a low-temperature phase. In most other samples from El Laco, however, apatite has grown simultaneously with magnetite and pyroxene and must be of higher temperature origin. The high fluid inclusion temperatures reported for pyroxene are in agreement with the temperatures obtained from pyroxene geothermometry (see below). The temperatures reported for the pyroxene are too high for hydrothermal systems *sensu stricto*, and imply crystallisation from a magma or from magmatic fluids. The minor amounts of magnetite deposited as magnetite veins with pyroxene and/or apatite near the contacts of massive iron-oxide bodies, are therefore, best explained as the product of the degassing of a volatile-rich magma during crystallisation. The most likely source for these magmatic fluids, are the magmas that formed the adjacent iron-oxide lavas and intrusive bodies.

Mineral Compositions

Analyses of pyroxene phenocrysts in unaltered El Laco silicate lavas and pyroxenes in magnetite-pyroxene dykes and veins demonstrate that they are almost identical in composition (Figure 8) and have similar crystallisation temperatures based on the Lindsley (1983) pyroxene geothermometer (silicate phenocrysts = 770 to 950 ºC; "vein" pyroxene = 500 to 850 ºC). The silicate lavas contain both Ca-rich and Ca-poor pyroxenes, and as a result, the pyroxene geothermometer is directly applicable. In the ore, however, unaltered low-Ca pyroxene is rare, and therefore, the pyroxene geothermometer will yield only minimum crystallisation temperatures. The similarity in major and minor elements between the pyroxenes in rock and ore suggests that both have "magmatic" compositions. In addition, the El Laco pyroxenes are very dry (<0.001 % OH, Broman et al., 1999), confirming that they are most likely of magmatic and not hydrothermal origin. If the centimetre- to metre-wide dykes and veins cutting andesite host rock tens of metres away from the massive orebodies record temperatures in excess of 800 ºC, it is likely that the 100 m thick massive orebodies were formed at even higher temperatures. Pyroxene associated with the El Laco orebodies is REE-rich with La values 280 to 1240 times chondritic abundances and Lu values >100 times chondrites (Rhodes et al., 1999).

Apatite is locally abundant (>20%) in some of the dykes and veins, but is much less abundant in the massive bodies and stratified ores. Apatite is also a common inclusion in phenocrysts in unaltered El Laco silicate lavas, suggesting that the silicate magmas were saturated or nearly saturated in apatite. Analysed apatites from El Laco andesites, ores and ore veins are almost identical in terms of F, Cl, SO_3, and OH (Figure 9) suggesting that all have magmatic compositions. Hydrothermal transport of iron by NaCl-rich brines, as has been proposed for other iron-oxide-apatite orebodies (Barton and Johnson, 1996), would be

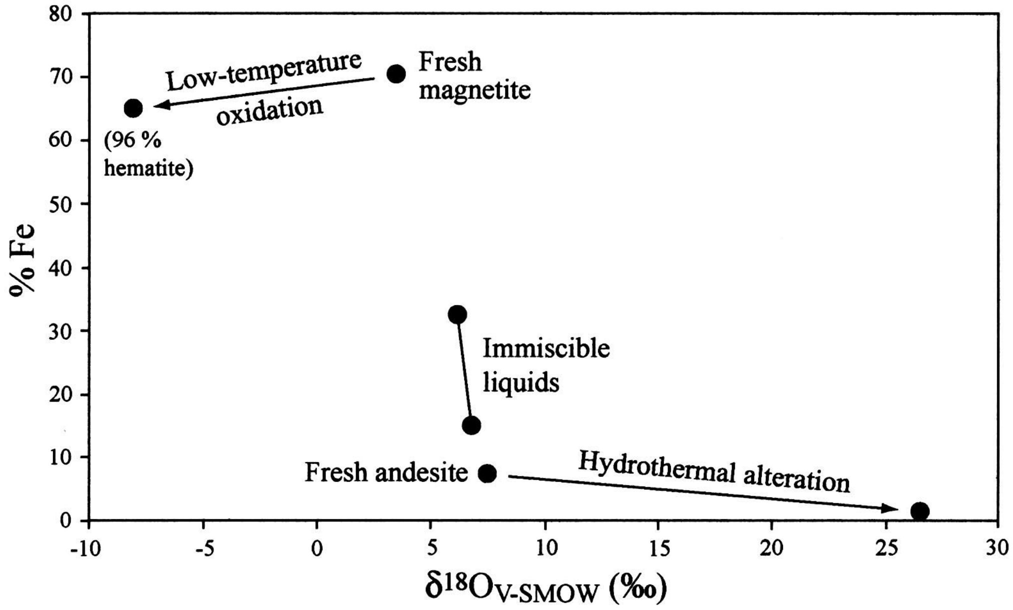

Figure 7: Oxygen isotope composition of iron-oxide-apatite ores and host rocks at El Laco, and immiscible liquids (Kyser et al., 1998).

expected to produce Cl-rich apatites. In fact, Cl-rich apatite is generally cited as evidence for hydrothermal deposition, while F-rich apatite is cited as an indication of magmatic crystallisation (Boudrou, 1995). Apatite associated with the El Laco orebodies is REE-rich with La values >10,000 times chondrites and average Lu values >200 times chondrites (Rhodes et al., 1999).

El Laco magnetite samples are rich in V (900 to 2500 ppm), Ni (90 to 360 ppm), and Mg (3000 to 8000 ppm) and poor in Cr (generally <10ppm), Ti (<5000 ppm), Co (<150 ppm), Zn (<150 ppm), Cu (<50 ppm), and Mn (<750 ppm), and they are similar in composition to magnetite in ores from the Cretaceous iron belt in Chile and the Kiruna iron ores in Sweden (Nyström and Henríquez, 1994). Average REE contents in El Laco magnetite samples range from 150 times chondrites for La to 15 times chondrites for Lu (Rhodes et al., 1999). Some magnetite samples in the El Laco deposits are rich in SiO_2 with contents up to 0.7 % (Nyström and Henríquez, 1994). XRD analysis shows distortion of the unit cell, suggesting that the SiO_2 is structurally incorporated in the magnetite.

Discussion of Magmatic and Hydrothermal Hypotheses for El Laco

Proposed Models for the Development of Massive Magnetite Ores

The origin of massive magnetite-apatite deposits is not well understood, and in fact, although they share a number of similarities, it is not likely that all of these deposits are produced by the same mechanisms. Proposed origins include: 1) separation of an immiscible iron-rich melt from a silica-rich melt during cooling; 2) partial or complete melting of iron-rich crustal rocks; 3) contact metamorphism and/or replacement of wall rocks by high-temperature, iron-rich magmatic fluids; 4) hydrothermal deposition or replacement; and (5) sedimentary-exhalative deposition. Processes 1 and 2, described below under "a magmatic model", both involve the transport of iron as an extremely iron-rich magma; they differ primarily in how that magma was formed. Processes 3, 4, and 5, described below under "a hydrothermal model", all involve the transport of iron as a dissolved component in a fluid, and differ mainly in where the fluid originates and in how and where the fluid interacts with the surrounding rocks to release the dissolved iron.

Agreement Between Magmatic and Hydrothermal Models

In addition to the analytical data described above there are a number of other features at El Laco that are agreed upon by both those advocating a hydrothermal origin and those advocating a magmatic origin for the deposits. The general geology of the El Laco deposits, including the composition and distribution of the main orebodies, vein complexes, and alteration zones is not in dispute. The textures within the deposits, including the size, shape, and distribution of magnetite, pyroxene, apatite, and quartz are features that have been observed and similarly reported by both groups. The preservation of volcanic features within the orebodies and in the surrounding friable, pyroclastic, iron-oxide deposits are also not in question. Although the two groups disagree about many of the interpretations, they agree that most of the hematite is secondary and formed by reaction between original magnetite and meteoric water and that this process is unrelated to the main alteration events. Both groups agree that the large, fumarolic sulphate mounds, generally located inside of the "ring" of magnetite orebodies, were deposited after the main ore forming event. Both groups also agree that in places, in particular at Cristales Grandes, magnetite-pyroxene and magnetite-apatite veins cut altered andesite host rock, and that these veins appear to have been precipitated from high-temperature fluids.

A Magmatic Model for El Laco

Experimental studies (Philpotts, 1967; Naslund, 1983) suggest that iron will dissolve readily in silica-rich melts at high temperatures, but upon cooling, immiscible silica-rich and iron-rich liquids will separate. In an experimental study of the effect of fO_2 on the development of immiscible iron-rich melts, the most iron-rich compositions (>78% Fe-oxides), occurred in melts at 1465 ºC formed in equilibrium with magnetite and hematite at the highest oxygen partial

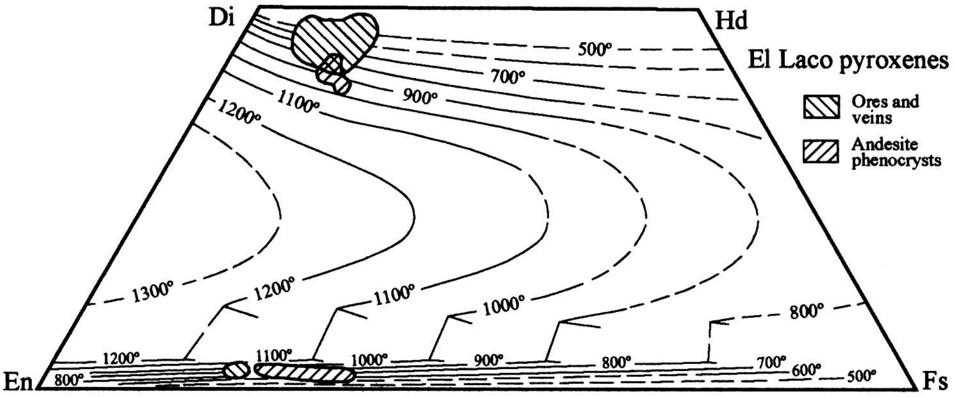

Figure 8: Pyroxene composition at El Laco and inferred crystallisation temperatures based on the Lindsley (1983) geothermometer.

pressures (Naslund, 1983). The addition of volatiles such as CO_2, H_2O, P, F, or SO_3 can lower the temperature of these iron-rich melts to more geologically reasonable values (< 1200ºC), and expand the immiscibility field to more iron-rich compositions (Lester, 2002). Weidner (1982) produced iron-oxide melts in CO_2 saturated experimental runs at 880 ºC, and Gibbon and Tuttle (1967) produced immiscible iron-oxide-rich melt in H_2O saturated experiments at 1060 ºC. The immiscible relationship between silicate liquids and FeS liquids has been long recognised (Naldrett, 1989). Recent field and laboratory research has suggested that there is a continuum between FeS liquids and Fe-S-O liquids that extends from pure FeS liquids to liquids with O/S > 1.3 (Larocque et al., 2000; Rose and Brenan, 2001). Such liquids are immiscible with silicate melts, and are characterised by high Fe and low Ti contents. The abundance of vesicles and gas escape tubes in the El Laco flows suggest that they contained considerable amounts of dissolved gas, and this is probably an important feature of iron-oxide deposits of magmatic origin. The presence of F-apatite indicates that F was present in the volatile phase. The abundance of sulphate around the El Laco volcanic centre, and the presence of anhydrite in pyroxene fluid inclusions (Sheets et al., 1997; Broman et al., 1999), suggests that SO_3 may also have been important in the system.

The relatively low Ti and P contents of the El Laco deposits relative to nelsonites, has been cited as evidence that they do not represent immiscible melts (Larson, 1994; Rhodes and Oreskes, 1996). The Ti and P contents of immiscible melts, however, will be determined by an equilibrium distribution between the melts. Experimental data for Ti and P distribution between immiscible melts (Visser and Koster van Groos, 1979; Naslund, 1983) suggest that an iron-rich liquid in equilibrium with a dacite or rhyolite containing 0.5% TiO_2 and 0.15% P_2O_5 would contain 0.75 to 2.0% TiO_2 and 0.3 to 1.8% P_2O_5. Eutectic proportions of ilmenite and apatite, as observed in nelsonites, would only be expected from magmas produced by partial melting or magmas derived from extreme crystal fractionation. In addition, the titanium content of magnetite, is controlled not by the amount of titanium in the melt from which it crystallised, but by the temperature and oxygen fugacity at which it forms, and at the high oxygen concentrations necessary to crystallise coexisting magnetite and hematite (observed locally in the stratified ore below Laco Sur), both minerals should have very low titanium contents (Buddington and Lindsley, 1964). The dissolved volatiles present in the El Laco magma will also influence the distribution of elements between immiscible liquids. If SO_3 was an important component of the volatile phase, the ores would be expected to be low in TiO_2.

An alternate to the immiscibility theory is that iron-rich magmas are formed by partial melting of iron-rich crustal rocks. Magmas formed from partial melting of sedimentary iron ores would be expected to have a distinct chemical signature (low La/Lu; anomalous Ce; low V; high Pb), and distinct isotopic signatures which are not observed. In addition, many sedimentary rocks are quite refractory, and a melt derived from partially melting a sedimentary sequence would be likely to contain xenoliths of unmelted sedimentary layers. Although rumoured to exist (Thomas, 1970; Frutos and Oyarzún, 1974), no such xenoliths have been described by any investigators.

It has been argued that iron-rich liquids, such as those advocated for El Laco, would have such low viscosity that they would form very thin, laterally extensive flows, not the thick, almost dome-like orebodies observed (Hitzman et al., 1992; Larson, 1994). While some pahoehoe features are preserved at El Laco, most of the flow features suggest a much more viscous magma. It is likely that the El Laco magmas were very fluid at depth, but upon reaching the surface they began to degas and crystallise, and as a result became very viscous. The ubiquitous presence of gas vesicles and gas escape tubes demonstrates that the magmas were very gas-rich and underwent extensive degassing upon eruption. It has also been suggested that the pyroclastic

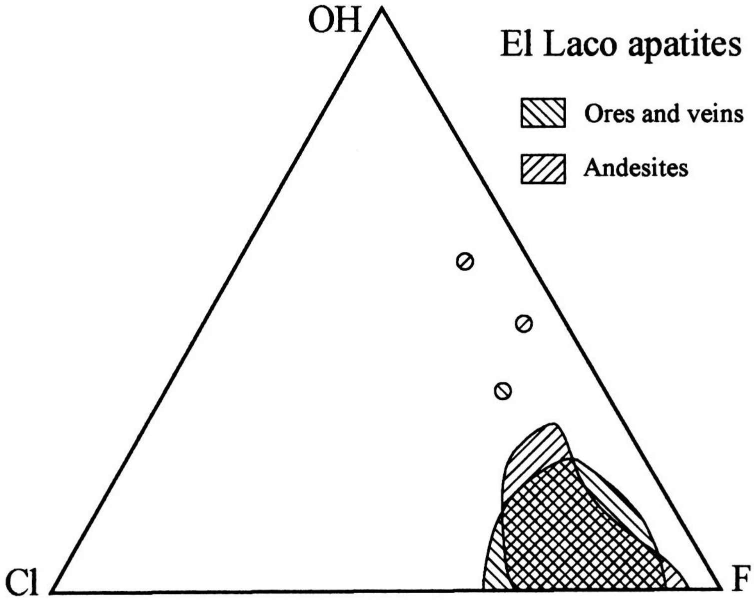

Figure 9: F-OH-Cl composition of apatites at El Laco.

deposits around and within the main orebodies do not represent the eruption of an iron-rich magma because they contain well-developed magnetite octahedra, and not the poorly formed crystals or iron-oxide glass that would be expected by rapid cooling of an iron-oxide magma according to Rhodes and Oreskes (1995a). Unlike silicate magmas, however, iron-oxide magmas should not form glasses upon quenching. As a result of the extremely low nucleation energy for magnetite, even under laboratory conditions with extremely rapid quench rates, high-temperature, iron-oxide liquids quench to a crystal mosaic of magnetite or magnetite ± hematite and not to glass.

The high density of iron-oxide magmas might be expected to decrease the likelihood that they would reach the surface (McBirney and Naslund, 1990). This may, in part, explain why localities like El Laco are so rare. Many magmas, however, are denser than the crustal rocks through which they erupt. Even a magma with a density of 5.0 could easily erupt if it was 50% vesiculated. Many of the ore samples at El Laco have densities <3.0, and a few of the most vesicular have densities <2.0. The existence of iron-oxide lavas at El Laco indicates that the processes responsible for the formation of the ore magma occurred high enough in the crust that the magma was able to vesiculate and erupt.

A Hydrothermal Model for El Laco

Traditionally, the word "hydrothermal" has been used to describe hot fluids derived from meteoric or intraformational water with or without contributions from magmatic fluids, and restricted to temperatures below ~500ºC (Lindgren, 1933). Higher-temperature "magmatic" or "skarn" fluids are derived directly from degassing magmas, and can reach temperatures in excess of 1000 ºC. The evidence at El Laco indicates that the main, massive, orebodies were crystallised from volatile-rich, iron-oxide magmas, and that the nearby magnetite-pyroxene and magnetite-apatite veins were formed from magmatic fluids released from these crystallising magmas. There is no evidence that replacement of pre-existing rocks by iron transported in either hydrothermal or magmatic fluids is responsible for any of the massive orebodies at El Laco.

The main El Laco orebodies formed at or near the surface. The area to the west of Pico Laco (Figure 2) was glaciated, and the San Vicente Bajo deposit appears to have been formed at shallow depth, and then uncovered by glaciation. There is no evidence, however, that the San Vicente Alto, Laco Norte, or Laco Sur deposits were ever glaciated, or that any significant amount of erosion has occurred since they formed. There is no evidence that these three orebodies were ever covered. The San Vicente Alto, Laco Norte, and Laco Sur deposits appear to have formed at the surface, and to have remained there ever since. The total replacement of near-surface silicate lavas or tuffs with coarse magnetite crystals by hydrothermal fluids is difficult to imagine. In general, in near-surface hydrothermal deposits "replacement is lacking and impregnation occurs only to quite a small extent" (Ramdohr, 1980). In fact, hydrothermal magnetite "is generally an indication of high temperature and considerable depth" (Ramdohr, 1980).

The formation of iron-oxide-apatite ores by fluid transport requires a system that can efficiently transport iron while limiting most other cations (except P) to low abundances. Experimental studies suggest that Fe can be transported in Cl-rich or Cl-poor hydrothermal solutions as Fe, Fe-Cl, or Fe-OH complexes (Chou and Eugster, 1977; Eugster and Chou, 1979; Whitney et al., 1985; Crerar et al., 1985). Although there is little available data for higher temperatures, existing data suggest that this process is more efficient at lower temperatures. Replacement of pre-existing silicate rocks would require the complete removal of Si, Al, Ca, Mg, Na, and K. Although there is a great deal of hydrothermal alteration at El Laco, there is no evidence for such massive movement of these other components. In addition, hydrothermal fluids would be expected to leave a distinctive trace element signature, rich in easily transported elements, and poor in less-easily transported elements. The El Laco ores, however, are rich in V, Ni, and REE, which are not easily transported in hydrothermal fluids and poor in Mn, Zn, Cu, and Co, which are more easily transported. The ores appear to be the source of fluids rather than the product of fluids.

One of the most persistent arguments that has been presented as evidence of a hydrothermal origin for the El Laco deposits, is that the magnetite in the massive ores and the magnetite in small veins cutting the host andesite near the massive orebodies have similar isotopic and trace element compositions, and since the veins are hydrothermal, both must be hydrothermal (Larson, 1994; Rhodes and Oreskes, 1995a, 1995b, 1996, 1999; Rhodes, 1996; Rhodes et al., 1997, 1999). This postulated hydrothermal system at El Laco, however, covered a range of temperatures (150 to >800 ºC), a range of fluid compositions (Na-K metasomatism, Ca metasomatism, argillic alteration, siliceous alteration), and a wide range of isotopically heavy fluids ($\delta^{18}O$ in altered rocks = 7.1 to 35.8). As such, if the El Laco deposits were produced by hydrothermal processes, one would expect magnetites with a wide range of isotopic and trace element compositions that varied from one texture to the next. If, however, the deposits were formed from volatile-rich, iron-oxide magmas, and from the high-temperature, magmatic gases released from and in equilibrium with those magmas, one would expect that the magnetites would have similar, magmatic, isotopic and trace element signatures with little variation between textural types.

El Laco magnetites, like those at other iron-oxide-apatite deposits (Gleason et al., 2000), have igneous REE patterns consistent with the interpretation that they formed from iron-oxide magmas and high-temperature magmatic fluids in equilibrium with those magmas. The similarity in bulk REE content between El Laco magnetites and andesites has been cited as evidence that the magnetites inherited their REE contents from andesite protolith during replacement (Rhodes et al., 1999). The magnetites, however, are more strongly enriched in LREE and have flatter HREE patterns than either unaltered or altered El Laco andesites. Average magnetites have chondrite-normalised La/Sm = 8.5 and Sm/Lu = 1.7, while average andesites have La/Sm = 4.6 and Sm/Lu = 2.3 (Rhodes et al., 1999). These differences cannot be attributed to alteration (altered andesite samples have lower La/Sm and higher Sm/Lu than the analysed magnetites), nor can it be attributed to contamination of the magnetite samples with pyroxene or apatite (average El Laco pyroxene has La/Sm = 2.7 and Sm/Lu = 2.5; average El Laco apatite has La/Sm = 7.4 and Sm/Lu = 6.8; Rhodes et al., 1999). The fact that the El Laco magnetites and unaltered andesites have similar total REE, was also cited as evidence that the iron ores did not form by liquid immiscibility which would be expected to concentrate REE in the iron-rich magma (Rhodes et al., 1999). Average Laco Sur bulk iron ores, however, have REE contents 20 times higher than average unaltered El Laco andesites (Heriberto Fortín, Compañía Minera del Pacífico internal report, 2001). REE in the bulk samples are concentrated in late-crystallising minor phases. In addition, the bulk iron ore has undergone degassing and some of the original REE content, which would presumably be concentrated in residual liquids during early magnetite crystallisation, may have been carried off in the volatile phase.

The strongest evidence against a hydrothermal or fluid-replacement origin and in favour of a magmatic origin, are the field relationships and well-preserved volcanic textures and intrusive structures. The pahoehoe and aa flow structures, the razor-sharp contacts with chilled margins, the open vesicular textures, the abundant open gas escape tubes, and the unconsolidated, massive, as well as size-sorted and bedded ash deposits are inconsistent with a replacement origin for these deposits. Although it has been suggested that these features are "palimpsest textures" inherited from an original andesite protolith (Rhodes et al., 1999) many of these features are not common in the surrounding unaltered lavas, and no partially replaced rocks that preserve original textures have been reported by any investigator. In fact, all of the proposed replacement processes for El Laco destroy textures. In addition, all of the analytical data is consistent with a magmatic model, and much of it is difficult to reconcile with a hydrothermal origin. The high-temperature fluid inclusions and pyroxene geothermometry results, the unusual oxygen isotope patterns, and the magmatic mineral compositions are unlike those reported from any hydrothermal or replacement deposit.

General Geology of the Cretaceous Iron Belt

The Cretaceous iron belt is approximately 600 km long and 25 km wide. It consists of seven large (>100 million tonnes of high-grade ore) and about 40 medium-sized and small deposits of magnetite-apatite ore (Figure 1). Three of the large deposits (Boquerón Chañar, Cerro Negro Norte, and Cristales) are unexploited or only partially mined and a fourth (Los Colorados) has been mined only since 1999.

The iron deposits consist of primary magnetite with varying amounts of secondary hematite. The apatite content is variable within and between different orebodies, and genetically related apatite veins are abundant in the ore districts. Unlike El Laco, magnetite orebodies in the iron belt contain large bladed amphibole crystals and pyroxene has not been reported; however, in some deposits pseudomorphs morphologically resembling pyroxene at El Laco have been observed. Based on field relationships, structures and textures, the different deposits in the iron belt have been interpreted as dykes, domes, sills, and stratiform subaerial orebodies, the latter including lava flows and pyroclastic deposits (Henríquez et al., 1994; Nyström and Henríquez, 1994, 1995; Travisany et al., 1995). Iron ore breccias and stockworks are present in many deposits close to massive ore, and occur in other parts of ore districts without obvious relationships to massive orebodies.

The deposits are considered to have formed in a magmatic arc along an Early Cretaceous shallow–marine basin (Vivallo et al., 1994c, 1995b). Some of them are hosted by volcanic rocks and others by granitoids, but most of them, including the largest deposits, occur associated with altered volcanic rocks in tectonic contact with granitoids within the north-trending Atacama mega-fault zone. The iron belt coincides spatially with this zone, and individual orebodies are elongated parallel to it. Most of the granitoids are dioritic members of large plutons, and appear to postdate the lavas. The volcanic host rock is dominated by basaltic to andesitic lava flows. The lavas away from ore deposits and granitoids are affected by non-deformational regional metamorphism, usually at prehnite-pumpellyite facies. Available radiometric ages of the rocks associated with the ores fall in the range of 100 to 128 Ma (Vivallo et al., 1995b).

Pyrite and less abundant chalcopyrite occur as dissemination and veinlets in the magnetite orebodies. The grade of the alteration, as reflected by extensive development of actinolite, is higher near large orebodies. The imprint of a subsequent episode of extensive silicic-argillic alteration is evident in the outer zone, and albite, K-feldspar (or biotite), and tourmaline are also present. The hydrothermally altered host rocks display a mineralogical and geochemical zonation around the iron orebodies, with an internal halo of actinolitised rocks passing outward into quartz-feldspar (albite, K-feldspar or biotite) alteration and silicified ± tourmalinised rocks (Vivallo et al., 1995a; Vivallo and Henríquez, 1997, 1998). As a result of the hydrothermal activity, the original textures of the volcanic rocks have been obliterated and hydrothermal breccias have been formed. Breccias located close to the orebodies are typically composed of magnetite, actinolite, and sulphides (mostly pyrite). Other ore breccias, composed of specularite hematite and sulphides (mostly pyrite and chalcopyrite), occur further away from massive ore. Quartz-feldspar alteration is commonly found in these breccias. The mineralogical and geochemical zoning around the iron orebodies reflect thermal and compositional gradients in the hydrothermal fluids during and after the emplacement of the massive iron orebodies (Vivallo et al., 1995a).

Mineralisation Associated with Magmatic Iron-oxide Deposits in Chile

No Cu-Au mineralisation has been reported from El Laco, but there are anomalous values of Au and U in areas affected by hydrothermal alteration, and pyrite has been found in a drill core at Pasos Blancos south of Pico Laco. A gold prospect, El Perro, is spatially related to the magmatic iron ore at Magnetita Pedernales (Grez et al., 1991). In the Coastal Cordillera, iron-oxide-Cu mineralisation can be divided into a northern group (18º to 26˚ S) and southern group (26º to 32˚ S) according to the concentrations of Au and Ag, the age of the host rocks, and the genetic relationship with tectonic and magmatic events (Henríquez and Vivallo, 2000).

Deposits Located Between 18º and 26˚ S

The Cu-mineralisation associated with iron deposits in this segment are related to Jurassic and Cretaceous magmatic arcs and the Atacama fault system (Espinoza, et al., 1996). The Jurassic deposits are veins in gabbroic to granitic intrusive rocks, and stratiform bodies (“mantos”) and breccia pipes in basalts and andesites of the La Negra Formation. Vein deposits hosted by intrusive rocks consist of Fe-Cu-Au mineralisation with magnetite, pyrite, chalcopyrite, actinolite and scarce apatite as primary minerals. Examples of these deposits are Toldo-Velarde, Montecristo and Mina Julia (Figure 1). The veins range from tens of metres to 2 km in length, from 1 to 12 m in width, and they are up to 700 m deep. They are fault-controlled and the mineralisation shows a vertical supergene zonation with an upper Cu-oxide zone, followed by a secondary enrichment level, and hypogene mineralisation at depth. The primary mineralisation is also vertically zoned: magnetite and actinolite increase with depth, whereas quartz and specularite are more abundant in the upper levels. Alteration halos with albite, quartz, chlorite, actinolite and sericite surround the veins.

The stratiform bodies consist of Cu ± Ag mineralisation with a lower Fe content than the veins. The host rocks are mainly andesitic to basaltic lava flows; dacites-rhyolites and andesites host the mineralisation at one deposit, Mantos Blancos. The mineralisation shows a vertical zonation similar to that of the vein deposits: copper-oxide minerals in the upper part are followed by a mixed zone with copper-oxides and sulphides, and a primary sulphide zone at depth. Hypogene mineralisation consists of bornite, chalcocite and chalcopyrite with specularite, pyrite, quartz and calcite as gangue minerals. The host rocks are altered and contain albite, chlorite, epidote and quartz.

Mineral-chemical and isotopic (Pb, S, C) studies, radiometric dating, and spatial relationships between the two types of deposits suggest a common hydrothermal origin related to Jurassic granitoids (Vivallo and Henríquez, 1998). The vein mineralisation was deposited in a high-temperature, plutonic environment, from neutral to slightly alkaline, magmatic solutions in equilibrium with magnetite, as shown by the association of magnetite, sulphides, and albite or potassic feldspar. These hydrothermal fluids moved up to a near-surface volcanic environment characterised by lower temperature and higher oxygen fugacity, with hematite as the stable phase, to form stratiform deposits (Vivallo and Henríquez, 1998).

Deposits Located Between 26º to 32˚ S

As in the previous segment, the distribution of the Fe-Cu-Au deposits is related to the evolution of Jurassic-Cretaceous magmatic arcs and especially to the Atacama fault system. In addition, there is a relationship with the iron orebodies of the Cretaceous iron belt. Manganese deposits are also spatially related to the iron and sulphide ores (Henríquez and Vivallo, 2000).

The Fe-Cu-Au deposits consist of veins, hydrothermal and tectonic breccias, and stratabound bodies hosted by Jurassic and Lower Cretaceous volcanic rocks; examples are Candelaria (Ryan et al., 1995), Manto Verde (Vila et al., 1997; Gelcich and Ugalde, 1997), Punta del Cobre (Marschick et al., 2000), and Panulcillo (Hopper and Correa, 2000) (Figure 1). Magnetite and/or hematite, variable amounts of pyrite and chalcopyrite, and traces of molybdenite and arsenopyrite characterise the primary ore. Common gangue minerals are actinolite, quartz, albite, K-feldspar and calcite. Most of the deposits show a vertical, primary zonation defined by an increasing Fe/Cu ratio with depth. Mineralogically, this zonation is represented by magnetite at depth, passing successively to magnetite-pyrite, pyrite-magnetite-chalcopyrite, and specularite-pyrite-chalcopyrite at higher levels. Supergene oxidation processes were important in some deposits (eg., Manto Verde), resulting in oxidation minerals of higher copper grade. Ore breccias with abundant specularite in the matrix are found in most of these deposits. The host rocks of the Fe-Cu-Au mineralisation and the Cretaceous iron deposits show the same alteration characterised by a zonal arrangement around the iron orebodies (Figure 10): an inner zone of actinolitised rocks passes outward into albitised (or K-feldspar-rich) rocks; outside these zones the rocks are silicified and locally contain tourmaline; this silicification generally overprints the actinolitised and albitised rocks. Some districts have an additional zone of quartz-sericitic alteration. Garnet and pyrrhotite are present in some districts (Panulcillo).

Based on field, textural and geochemical relationships, two main hydrothermal events are suggested (Vivallo et al., 1995a, b). The first event is characterised by neutral to slightly alkaline, high-temperature solutions, rich in Mg, Fe, Ca, CO_2, F, P, Cl, which produced the actinolitisation and albitisation in the rocks hosting the iron ores. This hydrothermal activity is related to the emplacement of the iron orebodies, which released volatiles rich in Cl, capable of remobilising Cu and Au originally contained in magnetite. A second hydrothermal stage, overprinting the first, consists of neutral to slightly acid hydrothermal solutions of lower temperature, rich in silica, F and B, resulting in an outer zone of silicified ± tourmalinised rocks.

Manganese deposits (eg., Venus, Coquimbana) composed of Mn-oxides, subordinate magnetite, and chert occur in many iron districts situated in limestones immediately overlying the volcanic rocks containing Fe ores and Fe-Cu-Au mineralisation (Henríquez et al., 2000). They appear to represent a last, volcanic-exhalative stage in the evolution of the mineralisation system with magmatic iron ore and hydrothermal Fe-Cu-Au schematically depicted in Figure 10.

Conclusions and Implications

The field relationships and the well-preserved volcanic structures and textures at El Laco indicate that the main, massive orebodies were crystallised from volatile-rich, iron-oxide magmas, and that the nearby magnetite-pyroxene and magnetite-apatite veins were formed at high temperature from magmatic fluids released from the crystallising magmas. All of the available analytical data are consistent with a magmatic origin for these deposits. There is no evidence to suggest that any of the iron orebodies at El Laco were produced by hydrothermal fluids. The clear evidence for iron-oxide magmas at El Laco suggests that a magmatic origin should be considered for other similar, less-well-preserved, deposits. The nature of such volatile-rich magmas is that they alter their host rock during emplacement and crystallisation. As such, even orebodies with alteration halos and gradational contacts may have formed from iron-oxide magmas. In other localities, hydrothermal breccias and vein deposits may be related to the degassing of an iron-oxide magma at depth.

Economic and subeconomic concentrations of Cu, Au, U, and REE are often associated with iron-oxide-apatite ores, either within the iron-oxide orebodies, or occurring nearby as breccias, veins, or stockworks, or in areas without known magnetite-apatite deposits. From an exploration standpoint it is important to determine if a Cu-(Au-U-REE) deposit is associated with the intrusion or eruption of iron-oxide magma, or unrelated to such magma and formed from hydrothermal processes alone. Iron-oxide magmas that have vented to the surface as ignimbrites, airfall ash, or lavas are less likely to have associated Cu or Au mineralisation because these elements are likely to be partially lost during degassing. Exploration targets for these "vented" systems will be at depth below the iron ores. U and REE are less likely to escape in the vented gases and may be present within the iron orebodies. Iron-oxide magmas that have intruded as sills or dykes are more likely to have associated Cu, U, Au, or REE mineralisation within, adjacent to, or stratigraphically above the iron ores, although additional mineralisation at depth can not be ruled out. Systems, formed by hydrothermal solutions, may have Cu, U, Au, or REE mineralisation as well. In these deposits, Cu, U, Au, and REE distribution will be most likely controlled by fluid circulation pathways, temperature gradients, host-rock compositions and permeabilities, and fluid mixing within the hydrothermal system, but may be derived from an iron-oxide magma at depth.

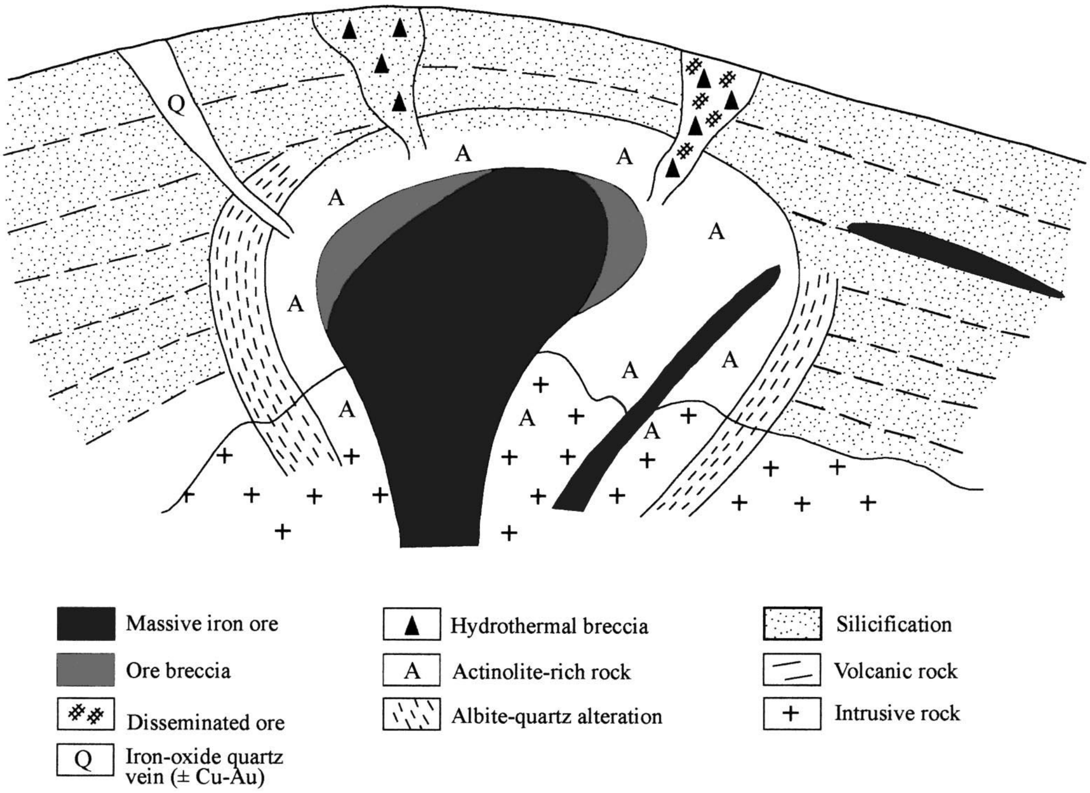

Figure 10: Model of the alteration zone distribution and Fe-Cu-Au mineralisation around magnetite-apatite orebodies based on the Cretaceous iron belt in Chile.

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