

6 Metallic ore deposits

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This chapter describes the metallic ore deposits of Chile, their mineralized host rocks and the processes involved in ore formation, and provides a brief overview of the mining history of this Andean copper-rich country. The ore deposits are ordered according to their respective economic importance. Thus, after mining history and a general introduction, Chilean porphyry copper–molybdenum deposits are described first, with subsequent sections dealing with epithermal precious metals, iron oxide copper–gold and iron oxide–apatite deposits, stratabound copper–(silver) ores, precious metal veins, sedimentary-hosted gold and porphyry gold deposits, skarn rock ores and, finally, an overview of metallogenic evolution.

Mining history

About 40% of the known copper resources of the world occur in Chile, with the native populations using the red metal at least since 500 bc. Bracelets, earrings and weapons that have been found in archaeological sites in northern Chile were made of either native copper or copper-rich minerals that were melted in small quantities and subsequently hammered. Copper production during Spanish colonial times (1541–1810) amounted to some 80 000–85 000 tons, with high-grade oxidized copper minerals being exploited and melted with charcoal. Despite this mining activity, however, Spaniards regarded copper as ‘plebeian metal’ because of its relatively low value, and it was used mostly as ballast for ships returning to Spain, rather than for technological or industrial purposes. The colonial Spaniards were much more interested in gold and silver, and mining activities were consequently mostly orientated towards precious metals.

Prior to Spanish conquest the Incas dominated northern Chile and had already exploited some placer gold deposits. During the second half of the sixteenth century placer deposits were the only source of gold, which was extracted by the Spaniards with intensive use of native labour, and annual gold production was in the range of 1 to 2 metric tonnes (t), declining by the end of the sixteenth century as the shallow, highest grade placer deposits were exhausted (Cuadra & Dunkerley 1991). During the seventeenth century overall Chilean metal production was meagre, with the entire gold production amounting to only 350 kg (Millán 2001), suppressed by harsh battles between the colonizers and the natives of southern Chile, destructive earthquakes, and frequent pirate assaults. In contrast, throughout the eighteenth century gold production increased progressively and the exploitation of precious metal from hard rock deposits began. In 1749 official gold coin production started in Chile, mining laws were published in 1754, and a mining court was established to resolve innumerable title disputes. However, mining development overall remained hampered by the economic monopoly imposed by the Spanish Crown.

Chile declared independence in 1810, but conflict with the Spanish colonial powers continued until 1826. The war of independence seriously affected Chilean mining activities, and gold production fell dramatically (by 1829, for example, no

gold coins were stamped), although several bonanza-type silver-bearing vein deposits were discovered and developed (e.g. Cachinal de la Sierra in 1822, and Arqueros in 1825). Mining became revitalized once the political turmoil ended, and so in the decade from 1832 a total of 418 hard rock mines and alluvial placer exploitations were active in central Chile (Cuadra & Dunkerley 1991). This mining revival was mainly focused on copper, although precious metal exploitation continued to be important. Further bonanza-type silver deposits were mined (e.g. Chañarcillo, 1832; Tres Puntas, 1848; Caracoles, 1871), mostly hosted by Lower Cretaceous carbonate sedimentary rocks. Total nineteenth century silver production was 7544 t, which placed Chile as a major world silver producer. In contrast, gold production declined during this century, with no new gold mines having been discovered (Cuadra & Dunkerley 1991).

Thus post-colonial metallic ore mining in Chile became dominated by copper, with 100 000 t of the metal being produced from 1841 to 1850, placing Chile as the second largest copper producer in the world after Great Britain (Camus 2003). The increase in Chilean copper production during the nineteenth century coincided with a rise in demand caused by the Industrial Revolution, mainly for weapons and shipyard uses. In 1851 Chile rose to become the number one copper producing country, a position it retained until 1880. Total Chilean copper production during the nineteenth century amounted to 1 876 000 t, all of which came from the high-grade (> 6–8% Cu) oxidized and supergene enriched zones in vein deposits of the Atacama and Coquimbo regions. More specifically, it was the Carrizal Alto, Dulcinea, La Higuera, Andacollo, Brillador and Tamaya mining districts of northern Chile that were the main contributors to copper production at that time. Although the occurrence of large porphyry copper deposits such as Chuquicamata and El Teniente had been known since the second half of the nineteenth century, these were originally uneconomic due to the lack of proper technology for their exploitation, except for their richest parts (e.g. the enargite and chalcocite veins at Chuquicamata).

The use of copper for electrical applications by the end of the nineteenth century further increased the demand for the metal. However, the shallow and high-grade supergene enriched parts of vein deposits were largely exhausted, as a result of which Chilean copper production declined from 1876 to 1891. Production increased again by the end of the nineteenth century, really taking off at the beginning of the twentieth century when the introduction of novel ore drilling, blasting, loading and transport technologies allowed bulk mining of low-grade and high-tonnage porphyry copper deposits. From 1906 to 1927 large-scale copper production started at El Teniente, Chuquicamata (Fig. 6.1a, b), and Potrerillos mines, which were developed by North American interests. Since then the huge, world-class Chilean porphyry copper deposits have continuously been the main source of Chilean copper and molybdenum production, which is still on the rise. Chile regained its number one position in copper production in the late twentieth century (1982), and it remains current world leader (2005).

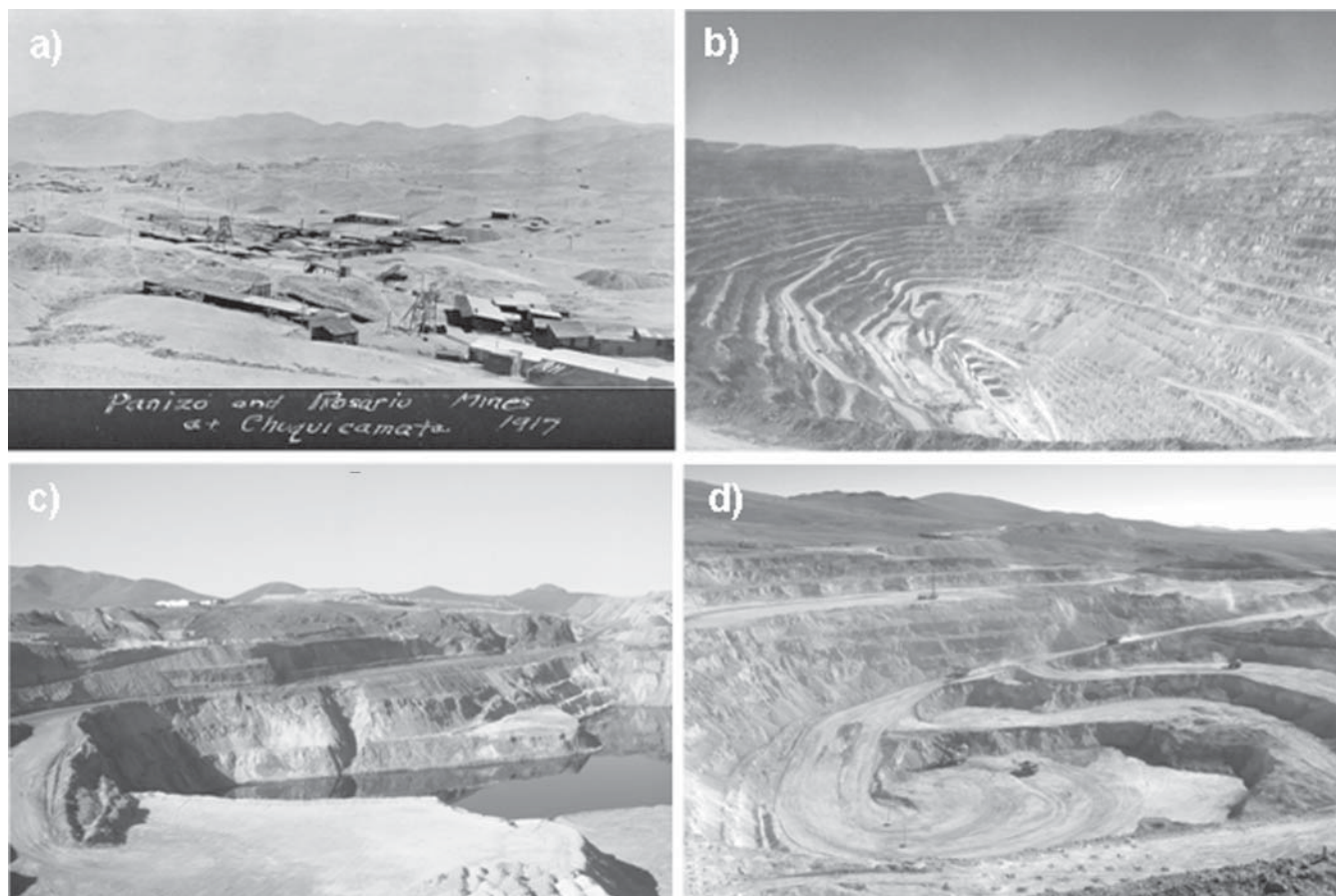


Fig. 6.1. (a) Chuquicamata porphyry copper deposit in 1917 showing shafts and camps of the Panizo and Rosario enargite–chalcocite veins (looking north). (b) The huge Chuquicamata open pit today looking north (CODELCO-Chile); dimensions are 4.0 by 2.5 km, and 850 m deep. (c) Andacollo porphyry copper (Aur Resources) showing open pit mining of the chalcocite enrichment blanket, and industrial installations and leaching pads in the background (looking south). (d) The recently developed Rosario porphyry copper deposit at 4300 m of altitude at Collahuasi (Minera Doña Ines de Collahuasi); stripping started in 2002 and mine production in 2004.

Finally, to return to precious metal mining, from the beginning of the twentieth century to 1979 gold production was, for the most part, a by-product of the exploitation of porphyry copper deposits, copper-bearing veins and disseminated deposits. Annual gold production ranged from about 2 to 10 t. Chilean gold production was subsequently boosted by the exploitation of El Indio epithermal deposit in 1980 and further discoveries and development of 20 other precious-metal-bearing deposits. Gold production peaked during the years 1996 and 2000 with 53.1 and 54.1 t per year, respectively. However, since then gold production has been dropping, and currently only four gold mines remain operating on a regular basis (La Coipa, Alhué, El Peñón and Pullalli), with 39.9 t of gold being produced in 2004.

Metallic ore deposits: an introduction

Some of the world's largest and richest porphyry Cu–Mo deposits occur in the Andes of northern Chile, and these have allowed Chile to become not only the leading copper-producing country of the world (5 418 800 t of copper were produced in 2004, 37% of annual global copper production), but also the second producer of molybdenum (41 883 tonnes in 2004), which is a by-product of the copper exploitation. A total of 14 porphyry deposits are currently exploited, and these contribute most of the Chilean copper production. Despite the fact that this overall huge richness in mineral resources is mainly due to

the large size of these copper–molybdenum deposits, there are other significant metallic ore occurrences. These latter deposits include precious metal epithermal, iron oxide copper–gold, volcanic-hosted stratabound copper–(silver), Fe oxide–apatite, mesothermal copper–gold-bearing veins, and minor skarns. Gold-rich porphyry deposits also occur, but none is currently exploited.

The Chilean Andes provide a typical example of a mountain belt developed along a convergent plate margin. This is a non-collisional orogen formed over a long-lived, currently active subduction system. The geological evolution of this orogen was characterized by the development of successive north–south trending magmatic arcs on the active continental margin. The magmatic foci of the arc migrated inland stepwise after stages of compressive deformation and associated crustal thickening. The abundance of calcalkaline igneous rocks is a distinctive characteristic of the Andes in Chile, and metallic ore deposits are mostly hydrothermal in origin, with an inherent and temporal relationship to the arc-related magmatic activity: they are believed to derive most of their metallic content from underlying subduction processes.

Most of the Chilean metallic mineralization occurs in the arid and semi-arid northern section of the country (18–34°S), an area which probably contains the richest ore deposits in the entire Andean mountain belt. In contrast, only sparse and less significant metallic deposits occur south of latitude 34°S, in the humid Andes of southern Chile. This difference arises in part from the climatic changes along the Andes, which control erosion rates and the efficiency of supergene enrichment

processes of metallic mineral deposits. However, the fundamental controlling reasons for this marked metallogenic difference along the mountain belt are still rather obscure. Most of the Chilean metallic wealth comes from exceedingly large and relatively high-grade magmatic-hydrothermal copper-molybdenum deposits that are genetically linked to Cenozoic calcalkaline porphyritic intrusions, though some metallic mineralization is also related to Mesozoic intrusions. In contrast, the Chilean Palaeozoic basement is poorly mineralized and lacks economic significance. Other regions in the world do not provide close parallels because they do not match the distinctive specifically copper-rich metallogenesis of the Andes of northern Chile, and the enormous size of its deposits.

Supergene processes have produced important enrichment of copper and silver deposits, plus some residual enrichment of gold ores in the central and northern parts of Chile. Indeed, past mining has largely been restricted to secondarily enriched parts of mineral deposits, and without the help of supergene enrichment Chile would not have been a copper-producing country until the mid-twentieth century, and would have been unimportant in terms of silver production. The uppermost several hundred metres of most major porphyry copper deposits in arid northern Chile have been affected by supergene oxidation and leaching, commonly giving rise to underlying chalcocite enrichment that made them economically viable (e.g. Sillitoe & McKee 1996). These authors reported supergene alunite K–Ar ages from about 34 to 14 Ma for deeply developed supergene profiles in porphyry copper deposits in northern Chile, while Hartley & Rice (2005) compiled geochronological data from different porphyry deposits suggesting that supergene oxidation and enrichment has been active between 17°S and 27°S from 44 to 6 Ma. Therefore supergene oxidation and enrichment processes were active from late Eocene to late Miocene times. In addition, lateral flow of copper-rich supergene solutions up to 8 km from source porphyry copper deposits in northern Chile has resulted in some 12 'exotic' copper deposits. These consist of oxidized copper minerals (mainly silicates, chlorides, oxides and carbonates) that impregnate gravels and underlying bedrock along palaeo-drainages (Münchmeyer 1996). Further south, in the Andes of central Chile, the late Miocene–early Pliocene porphyry copper deposits of Los Pelambres, Río Blanco and El Teniente (see below) also developed immature supergene chalcocite enrichment blankets. The supergene processes in these young porphyry copper deposits were still active at the initiation of mining.

Porphyry copper–molybdenum deposits

Porphyry copper deposits are the most abundant type of mineralization in the Chilean Andes, and a detailed review has been published (in Spanish) by Camus (2003). Porphyry Cu–Mo deposits occur in six longitudinal belts along the Andes of northern Chile, each representing a discrete metallogenic period. The porphyry belts are: Late Palaeozoic–Triassic (298–230 Ma), Early Cretaceous (132–97 Ma), Palaeocene–early Eocene (60–50 Ma), late Eocene–early Oligocene (43–31 Ma), and late Miocene–early Pliocene (12–4.3 Ma) (Camus 2002, 2003). The two youngest porphyry belts are unquestionably the most important from an economic viewpoint, these having the largest deposits and concentrating most current mining operations. In contrast, the majority of the older porphyry Cu–Mo deposits are currently subeconomic, except for Cerro Colorado (Eocene), Lomas Bayas (Palaeocene), Andacollo (Fig. 6.1c) and Dos Amigos (Early Cretaceous), which are currently mined (2005), and Spence (Eocene) which is under development.

Late Eocene–early Oligocene porphyry Cu–Mo belt

This is the most significant Chilean porphyry belt. It extends for more than 1400 km along the Domeyko Cordillera and can be

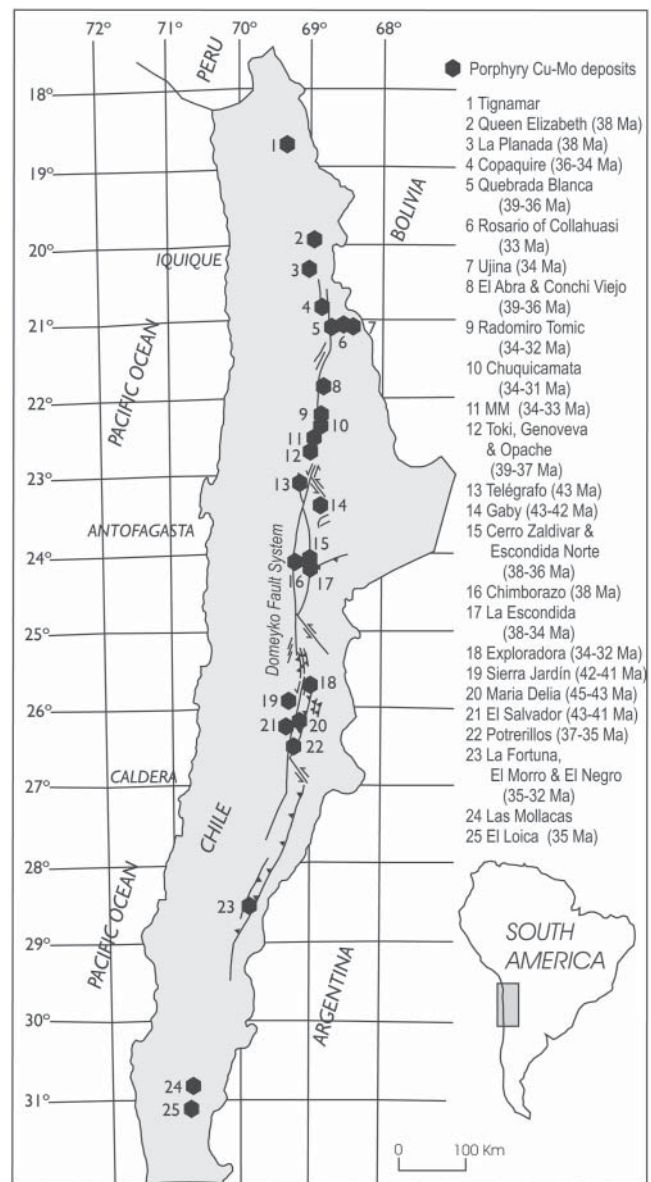


Fig. 6.2. Late Eocene–early Oligocene porphyry Cu–Mo belt of northern Chile (ages after Camus 2003).

traced from the border with Peru (18°S) to latitude 31°S (Fig. 6.2). This arc-parallel belt includes 30 porphyry Cu–Mo deposits and prospects with the highest amount of copper resources. Overall it constitutes the largest copper concentration in the world, totalling about 220 million tonnes (Mt) of copper (resources plus production; Camus 2002). The most distinctive geological characteristic of this porphyry belt is its spatial relationship with the Domeyko Fault System (Fig. 6.2) which is an arc-parallel regional fault zone that follows an uplifted crustal block (Domeyko Cordillera) cored by Upper Palaeozoic basement rocks (Maksaev & Zentilli 1988, 1999; Boric *et al.* 1990; Reutter *et al.* 1991, 1996). The copper-bearing stocks occur along-strike of master faults in this regional system (e.g. Copaquire, Quebrada Blanca, Ujina, Radomiro Tomic, Chuquicamata, La Escondida and Cerro Zaldívar), as well as along NW-trending subsidiary faults (e.g. Rosario of Collahuasi (Fig. 6.1d), El Abra, Potrerillos and La Fortuna). An apparent exception to this close link between faults and mineralization is provided by El Salvador, one of the oldest deposits of this group, which is located some 9 km west of the

major north–south trending Cerro Castillo fault. However, El Salvador and another six mineralized centres are aligned for 4 km along a NNE trend which suggests there is also a structural control, albeit more arcane than is normally the case (Gustafson *et al.* 2001). A number of authors have suggested a direct genetic relationship between the Domeyko Fault System and copper mineralization (e.g. Sillitoe 1981; Hunt *et al.* 1983; Baker & Guilbert 1987; Makshev & Zentilli 1999; Richards 2000; Richards & Villeneuve 2002), and this association has successfully been used for exploration (e.g. Lowell 1991; Ortiz 1995).

The late Eocene–early Oligocene porphyry Cu–Mo deposits are not regularly distributed along this north–south trending belt, but rather form local clusters within areas of <200 km². The clusters typically approximate alignments rather than equidimensional groupings of deposits, with the individual centres being strung out either parallel (e.g. Chuquicamata district) or transversely (e.g. Copaquaire, Quebrada Blanca, Rosario de Collahuasi, and Ujina group of deposits at 21°S; Sillitoe 2004) (Fig. 6.2). The copper-bearing stocks cluster in areas where late Eocene–early Oligocene pre-mineralization equigranular dioritic to granodioritic plutons occur ('plutonic precursors' of Sillitoe 1988). Some of these plutons have radiometric ages that are very close to, or overlap with, the age of mineralized porphyries. These plutons locally constitute the country rocks of mineralized stocks, which, in some cases, appear to be associated with porphyritic facies of otherwise more equigranular plutons. The Southern Granodiorite at El Abra (Ambrus 1977) and the Fortuna Granodioritic Complex at Chuquicamata are the most conspicuous examples of these plutonic precursors (Sillitoe 1988; Dilles *et al.* 1997).

The concurrence of the locus of unmineralized and mineralized late Eocene–early Oligocene plutons suggests that several batches of magma followed a similar ascent path through the crust, and were emplaced at distinct centres of igneous activity. This is reminiscent of the separation of modern volcanic centres along the Southern Volcanic Zone of the Andes. The separation of the igneous centres may have resulted from long-term rheological contrast of the crust along the magmatic arc, with porphyry Cu–Mo clusters/igneous centres localized at zones of diapiric uprise of magmas as envisaged by Yañez & Makshev (1994) for the Chilean porphyry belt. However, Richards (2000) and Richards & Villeneuve (2002) have suggested instead a structural magma focusing. In addition, Behn & Camus (1997) and Behn *et al.* (2001) showed that the porphyry clusters match with regional magnetic anomalies. These lie transverse to the Cordillera, and have been attributed to hypothetical batholiths at depth, which are thought to be common parental intrusive complexes for the mineralized stocks. Minor post-mineralization dykes and stocks commonly occur, which according to available geochronological data were emplaced shortly after the mineralization (e.g. latite dykes at El Salvador (Gustafson & Hunt 1975), and the rhyolite dome and rhyolite dyke at La Escondida (Padilla *et al.* 2001)).

The late Eocene–early Oligocene porphyry Cu–Mo deposits are mostly associated with granodioritic to quartz monzonite porphyritic stocks that intruded unrelated country rocks of diverse nature. Comagmatic volcanic piles, if they ever existed, are not preserved. The plutonic complexes are typically characterized by successive, small (0.5–2 km²), roughly cylindrical porphyritic intrusions, and include many pre-, syn- and post-mineralization porphyry dykes or epizonal stocks (e.g. El Salvador deposit; Gustafson & Hunt 1975). Copper and subordinate molybdenum sulphides typically occur in disseminated grains and stockwork veins throughout a large volume of intrusive rock, indicating significant brittle fracturing and volume increase of the igneous rock mass during mineralization (e.g. Burnham 1985). Breccia bodies with a tourmaline matrix occur in Quebrada Blanca and El Salvador, and breccias with a biotite matrix occur in El Abra (Hunt *et al.* 1983; Gustafson &

Hunt 1975; Ambrus 1977). However, hydrothermal breccias are only a minor part of these large porphyry systems, whose dominant copper-bearing stockwork ores are largely hosted by intrusive rocks. At Chuquicamata local mechanical brecciation related to faulting also occurs. Pervasive hydrothermal alteration zones of potassic, phyllic, argillic and propylitic types typically form annular shells around intrusions similar to the model of Lowell & Guilbert (1970), although significant distortions due to local structural conditions or multiple intrusive pulses occur (e.g. El Abra (Ambrus 1977) and Chuquicamata (Alvarez *et al.* 1980)). A late advanced argillic alteration cap occurs at El Salvador, La Escondida and Rosario de Collahuasi (Gustafson & Hunt 1975, Gustafson *et al.* 2001, Padilla *et al.* 2001; Bisso *et al.* 1998). The hypogene mineralization of chalcopyrite, bornite, pyrite, molybdenite and locally enargite is primarily associated with the potassic cores and phyllic alteration zones of the deposits. In most cases these zones were exhumed and exposed at the surface. The ore bodies are irregular, but conform to roughly domal cappings that are restricted to the exposed apical section of the host intrusive complexes and their immediate host rocks. The hypogene shape of the ore bodies has been in almost all cases modified by supergene processes, so that the richest copper ores commonly underlie partially leached, oxidized rocks forming irregular blankets containing supergene copper sulphides (chalcocite, djurleite, covellite and anilite), which have replaced hypogene sulphide minerals.

The Chuquicamata deposit is the largest porphyry copper deposit of this belt, and has produced a total of 31 Mt of Cu metal from 1915 to 1997, with an equivalent amount of contained Cu still remaining in the main orebody. Chuquicamata had an intimate and complex relationship with an active regional fault during its mineralization, which included the superposition of at least two distinct periods of copper introduction related to an early stage of potassic alteration, followed by a later structurally controlled phyllic overprint (Ossandón *et al.* 2001; Lindsay *et al.* 1995). Despite its uniqueness this enormous copper deposit may be catalogued as a classic porphyry copper deposit following the classification of McMillan & Panteleyev (1980) and the same applies to most other porphyry Cu–Mo deposits of this belt. An exception is the Fortuna cluster, where Au-rich porphyry deposits occur (La Fortuna, El Negro and El Morro prospects; Perelló *et al.* 1996). These are relatively small porphyry systems that occur along a local NW structural trend in the High Andes at 28°38'S, between major NNE-trending reverse faults that post-date porphyry mineralization. These porphyries have a potassic alteration core (dominated by an assemblage of biotite and magnetite) that grades to a narrow outer zone of propylitic alteration. Phyllic alteration is poorly developed, being only significant in the uppermost sections of El Morro and La Fortuna prospects. An epithermal advanced argillic alteration assemblage is overprinted on previous potassic and phyllic alteration zones in the uppermost section of La Fortuna prospect, extending to the SE into the Cantarito epithermal prospect. The alteration zoning is largely vertical rather than lateral and these systems seem to be vertically elongated. Stockworks of polydirectional quartz and sulphide veinlets locally occur, but most of the mineralization of these porphyry Cu–Au–Mo deposits is structurally controlled, and dominated by sheeted-vein zones formed of subparallel sets of near-vertical quartz and sulphide veinlets. Sheeted-vein orientations match the NW and NNE dominant fault trends of the district.

Abundant K–Ar, ⁴⁰Ar/³⁹Ar, Rb–Sr and U–Pb radiometric ages have been reported for the porphyry Cu–Mo deposits of northern Chile (Camus 2002, 2003). The isotopic ages demonstrate that the copper mineralization took place within a discrete interval of geological time from 41 to 31 Ma. In fact, according to the available radiometric data, most major deposits were formed within the period from 38 to 36 Ma (Copaquire,

Rosario de Collahuasi, Ujina, Quebrada Blanca, El Abra, Toki, Opache, Genoveva, Chimborazo, Cerro Zaldivar, La Escondida and Potrerillos), El Salvador and Gaby were formed at 42–41 Ma, and Fortuna and El Negro Au-rich porphyry deposits at about 35 Ma. However, the largest and richest deposits are those that are the youngest in this belt, i.e. Chuquicamata, Radomiro Tomic and MM, formed at 34–31 Ma (Camus 2003).

Late Miocene–early Pliocene porphyry Cu–Mo belt

The late Miocene–early Pliocene porphyry belt in the Andes of central Chile (32–34°S) is the second in economic importance in the country (Fig. 6.3). This belt includes world-class porphyry Cu–Mo deposits such as El Teniente, Los Bronces–Río Blanco and Los Pelambres. Overall, these deposits total 183 Mt of contained copper (resources plus production; Camus 2002). According to recent geochronological data El Teniente and Río Blanco–Los Bronces were formed at the same time, from 6.46 to 4.37 Ma (Deckart *et al.* 2003; Maksaev *et al.* 2004), whereas mineralization at Los Pelambres developed earlier, from about 13 to 10 Ma (Bertens *et al.* 2003). These Cu–Mo deposits are typically related to multiphase porphyritic stocks and associated magmatic–hydrothermal breccia complexes. The composition of the porphyritic stocks varies from quartz diorite to quartz monzonite. The porphyritic bodies were emplaced within country rocks of diverse nature, including Lower Cretaceous volcanic and sedimentary rocks at Los Pelambres (Rivano and Sepulveda 1991), a Middle Miocene granodioritic batholith at Río Blanco–Los Bronces (San Francisco Batholith) and minor Miocene volcanic rocks (Serrano *et al.* 1996), and Miocene mafic rocks at El Teniente (Cuadra 1986; Skewes *et al.* 2002).

Pervasive alteration zones of potassic, phyllic, propylitic and minor argillic types in general form annular shells around the porphyry intrusions similar to the model of Lowell & Guilbert (1970), but they are overprinted by successive intrusive phases with associated high-temperature hydrothermal events, and particularly by the development of late breccia bodies and associated phyllic alteration. Copper and molybdenum sulphides occur within a stockwork of veinlets throughout a large volume of rocks related to the potassic and phyllic alteration zones, but the occurrence of large, mineralized, biotite-bearing and tourmaline-rich breccia pipes is characteristic of these deposits (Skewes & Stern 1995; Skewes *et al.* 2002). About 50% of the copper ore occurs within the matrix of hydrothermal breccias at Río Blanco–Los Bronces (Serrano *et al.* 1996). Ore-grade material is largely restricted to the intrusive bodies at Los Pelambres and Los Bronces, but at Río Blanco it also occurs within volcanic country rocks, and at El Teniente more than 70% of the ore is contained within biotitized andesitic–basaltic–gabbroic Miocene country rocks. Diatreme breccia pipes occur at El Teniente and Río Blanco; the conspicuous funnel-shaped Braden breccia pipe occurs roughly in the centre of El Teniente orebody and is 1200 m in diameter at the surface, tapering down to 600 m at 1800 m below the surface. This diatreme is poorly mineralized (*c.* 0.3% Cu), but a rim of tourmaline-rich breccia (Marginal breccia) with high copper grades surrounds it. At Río Blanco a post-mineralization early Pliocene volcanic dacitic diatreme occurs and its tuffaceous ejecta cover part of the deposit (La Copa volcanic complex; Serrano *et al.* 1996).

Large-scale tectonic and local structural control is of utmost importance in focusing magma injection and hydrothermal fluid flow for porphyry systems. The late Miocene–early Pliocene porphyry Cu–Mo deposits of the High Andes are north–south aligned overall, but they appear to be located at NE- and NW-trending fault intersections. Recent explorations by Codelco at El Teniente and Río Blanco districts have

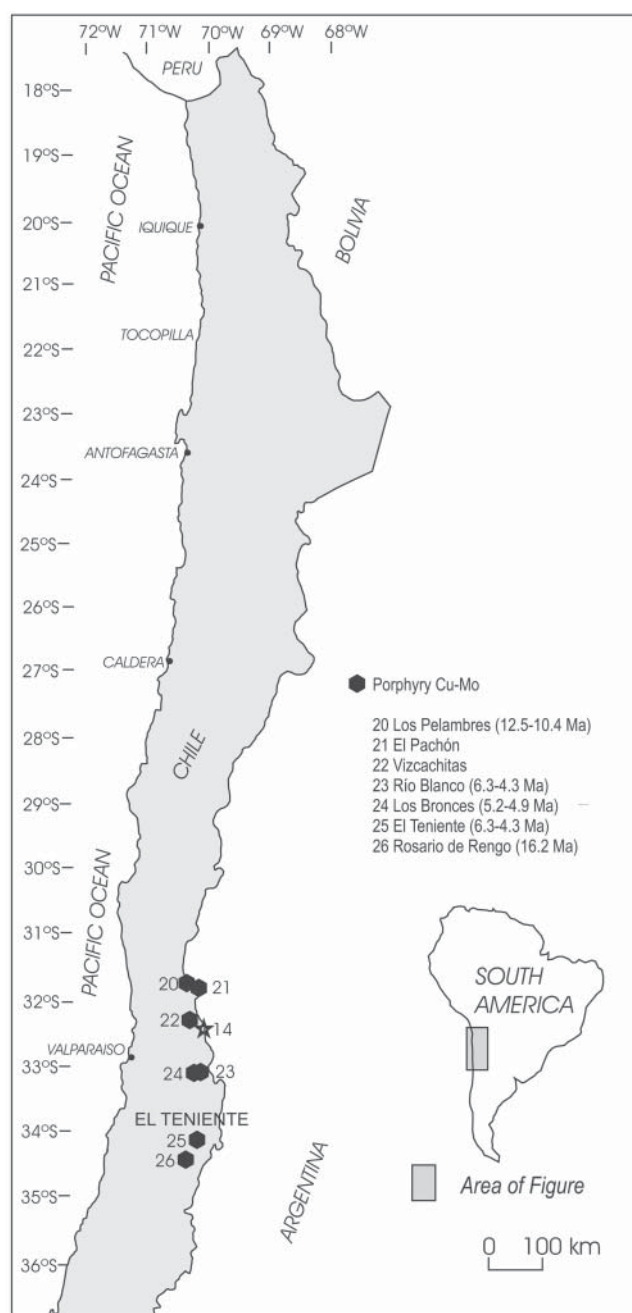


Fig. 6.3. Late Miocene–early Pliocene porphyry Cu–Mo belt of central Chile (ages after Kurtz *et al.* 1997; Bertens *et al.* 2003; Deckart *et al.* 2003; Maksaev *et al.* 2004).

identified first-order NE-trending lineaments that appear to have controlled precursor tonalitic intrusions, the location of porphyry deposits, and other prospects at these districts. NE-trending faults are also the dominant structures that controlled hypogene sulphide mineralization at Los Pelambres orebody. At all three of these porphyry systems intersecting NW-trending faults also occur, which are regarded as second order structures that controlled the dacite hypogene mineralization at El Teniente, the hydrothermal breccia development at Río Blanco–Los Bronces, and subsidiary hypogene mineralization at Los Pelambres. The occurrence of NW-trending faults appears to be related to inherited basement structures that occur as major NW-trending lineaments in the Argentinean Andes. These major lineaments coincide with the location of

modern volcanic centres of the Southern Volcanic Zone of the Andes as well as with the porphyry deposits.

Palaeocene–early Eocene porphyry Cu–Mo belt

A belt of Palaeocene–early Eocene porphyry Cu–Mo deposits extends for over 1300 km from 16°S latitude in southern Peru to 29°30'S in northern Chile. This belt comprises the porphyry Cu–Mo deposits of Mocha, Cerro Colorado, Spence, Sierra Gorda, Centinela, Polo Sur, Lomas Bayas, Fortuna del Cobre, Relincho and Las Pascualas in northern Chile (Fig. 6.4). The same belt includes the most significant Cu–Mo porphyry deposits of Peru (Cerro Verde–Santa Rosa, Cuajone, Quellaveco and Toquepala) and shows K–Ar ages ranging between 58 and 51 Ma (Sillitoe 1990). The Peruvian porphyries represent about 39 Mt of contained copper as a whole (resources plus production), whereas the numerous, but less rich, Chilean porphyries of this belt total 12.7 Mt of contained copper (resources plus production; Camus 2003). Owing to the

overall poorer mineralization of the Chilean Palaeocene–early Eocene porphyries, only those with well-developed supergene enrichment are currently economic. This is the case of Cerro Colorado that is currently exploited, and Spence that is under development.

Cerro Colorado is located 95 km ENE of Iquique, and its proven and probable reserves were 228 Mt at 1.0% Cu (0.5% cut-off; Bouzari & Clark 2002). The copper mineralization at Cerro Colorado is related to tonalitic and monzonitic porphyries ($^{40}\text{Ar}/^{39}\text{Ar}$ 52.8 ± 0.5 Ma (Bouzari & Clark 2002); molybdenite Re–Os 55.5 ± 0.3 Ma (Cotton 2003)) that were emplaced within a sequence of Cretaceous andesites, and developed apical breccias. A wide zone of propylitic alteration surrounds a core of potassic alteration, and both are overprinted by a pervasive quartz-sericitic alteration that masks the original texture and composition of the host rocks. The mineralized rocks are andesites (79%), breccias and porphyries (21%) within an area 2000 m long (east–west) by 1000 m wide (north–south). The upper 70 to 120 m of the deposit is leached and barren, followed at depth by a 60–120-m-thick oxidized ore blanket (chrysocolla, atacamite), and deeper by a supergene chalcocite-enriched blanket 50 to 70 m thick, reaching locally up to 200 m (Bouzari & Clark 2002).

The Spence porphyry copper deposit was discovered in June 1996 by Rio Algom mining company and currently a mine is being developed. The global geological resource is 330 Mt at 1.18% copper (cut-off grade of 0.6%). This enriched Cu–Mo porphyry deposit occurs under an alluvial plain of the Atacama Desert and is completely covered by gravel with an average thickness of 70 m. The porphyry copper is related to three Palaeocene dacite porphyry stocks and tourmaline-bearing hydrothermal breccias aligned NE–SW. Laser $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages for magmatic biotite in pre- and late-mineralization stocks are 57.0 ± 0.7 and 56.6 ± 0.6 Ma, whereas hypogene alunite from 46 to 113 m below the base of the enrichment blanket yielded $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 33.2 ± 1.9, 33.4 ± 1.1 and 35.6 ± 1.4 Ma (Rowland & Clark 2001). Under the gravel cover and leached rocks a 20-m-thick copper oxide zone occurs (atacamite), which is followed by a well-developed supergene enrichment blanket (chalcocite) with an average thickness of 60 m. Supergene alunite and natroalunite from leached and chalcocite enrichment zones range from 44.4 ± 0.5 to 27.7 ± 5.4 Ma (13 inverse isochron $^{40}\text{Ar}/^{39}\text{Ar}$ dates; Rowland & Clark 2001).

The belt of Palaeocene–early Eocene porphyry copper deposits extends farther south at least to 29°S latitude as attested by the occurrence of the Relincho and Las Pascualas Prospects; Relincho was explored by Outokumpu reporting resources of 130 Mt at 0.8% Cu. Tourmaline breccias are commonly associated with the Palaeocene–early Eocene porphyry copper deposits (e.g. Copucha breccia pipe at Sierra Gorda; Boric *et al.* 1990). In addition, Palaeocene granodioritic plutons in the Copiapó and Inca de Oro areas of northern Chile (about 27°S) host a great number of vertical tourmaline-bearing breccia pipes; these are mostly smaller than 100 m in diameter. Some breccia pipes are mineralized with Cu–Mo and a number also contain Au and/or W. The main examples are the San Pedro de Cachiyuyo, Cachiyuyo de Llampos, Los Azules and Cabeza de Vaca districts (Sillitoe & Sawkins 1971; Colley *et al.* 1991; Sillitoe 1985). Cu–Mo mineralization is restricted to the matrix of the breccia bodies and no disseminated or stockwork mineralization has been found to date within the host intrusives at these districts.

Early Cretaceous porphyry copper belt

Early Cretaceous porphyry copper deposits occur along a belt that extends along the Coastal Cordillera of northern Chile and comprise Galenosa, Puntillas, Antucoya–Buey Muerto, Dos Amigos at Domeyko, Pajonales, Andacollo and Colliguay (Fig. 6.5). Most of them are low-grade and only the chalcocite

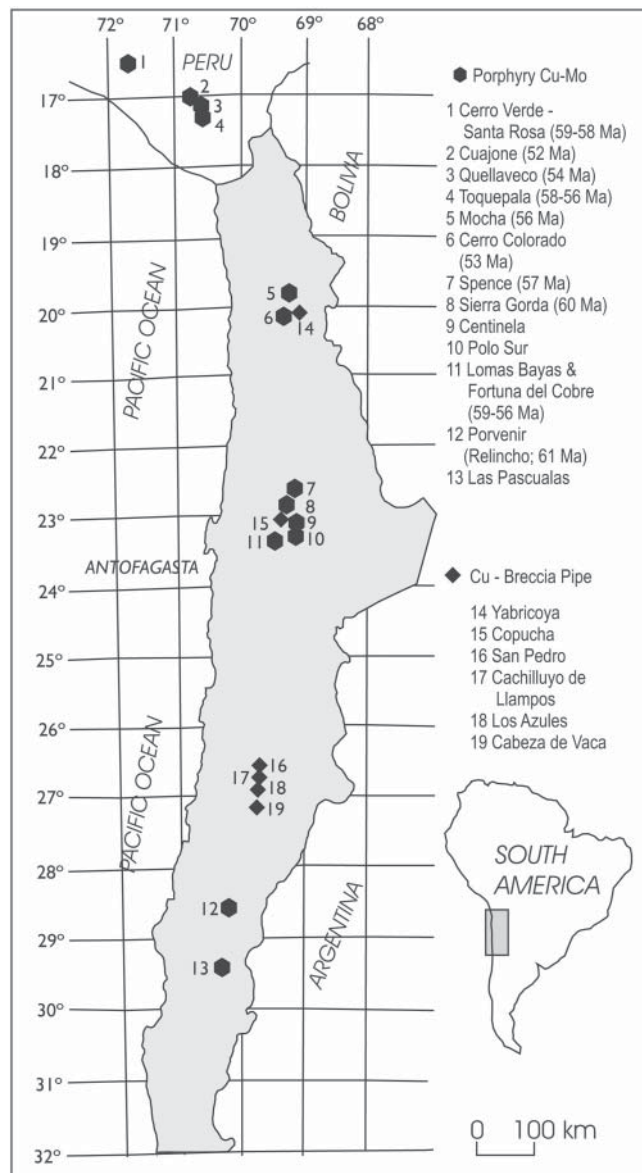


Fig. 6.4. Palaeocene–early Eocene porphyry Cu–Mo and tourmaline breccia pipes of northern Chile and southern Peru (ages after Camus 2003).

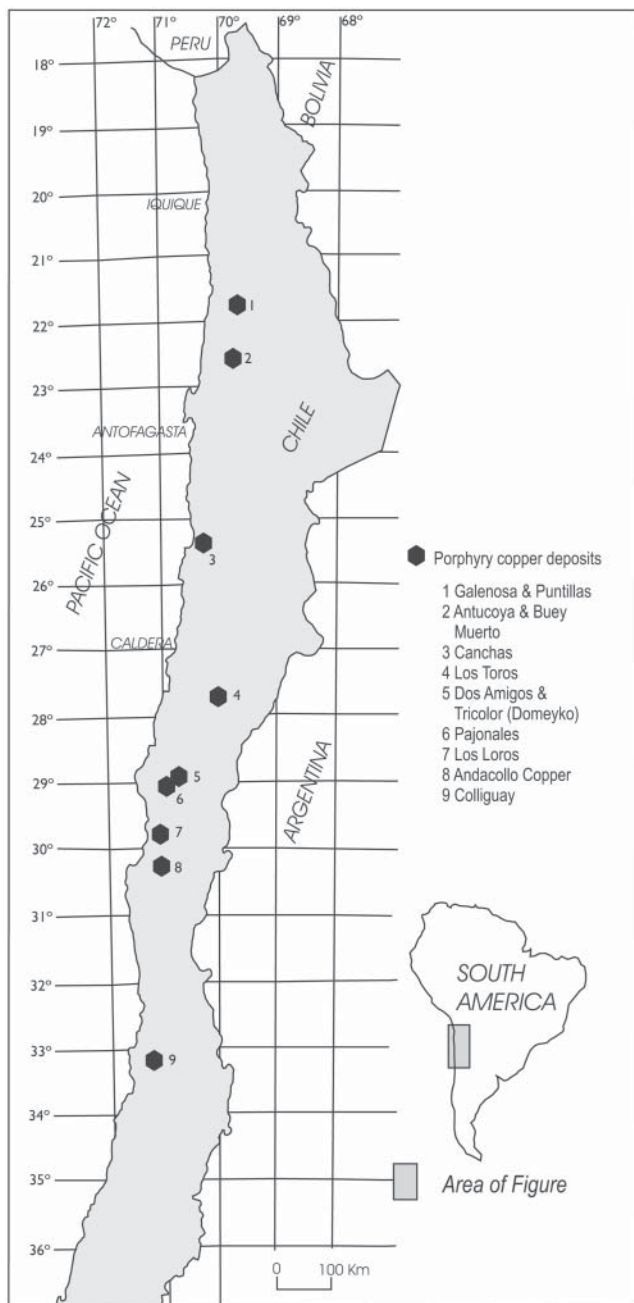


Fig. 6.5. Early Cretaceous porphyry Cu–Mo belt of northern Chile.

supergene enrichment blanket of Andacollo and Dos Amigos are currently mined. These porphyries were early recognized as forming a distinct belt that was referred as the ‘Pacific Belt of porphyry copper and hydrothermal systems’ along the Coastal Cordillera (Llaumet *et al.* 1975) and K–Ar dates ranging from 132 to 97 Ma indicated an Early Cretaceous age (Munizaga *et al.* 1985).

Andacollo is economically the most important porphyry deposit of this belt; its resources total 300 Mt at 0.70% Cu, 0.015% Mo and 0.23 g/t Au (Llaumet *et al.* 1975; Reyes 1991). Copper–gold mineralization is hosted by mostly altered Aptian–Albian andesitic and dacitic volcanic rocks, and small stocks and irregular dykes of potassium-rich tonalitic porphyry. The orebody occurs in the central section of an irregular north–south trending bleached zone of quartz-sericite alteration, 5 km long by 0.6–5 km wide, surrounded by a subtle propylitic zone. The mineralized area is roughly a north–south

elongated body (1800 × 1200 m) and mostly coincides with a core of potassic alteration (biotite and K-feldspar). The deposit shows a marked supergene zonation with a 30-m-thick capping of leached rocks (0.07% Cu) that contains mainly goethite, jarosite and hematite, and copper oxidized minerals. Under the leached rocks a supergene-enriched blanket (40 m average thickness) occupies an area of 1.5 km² with copper grade ranging from 0.6 to 1.5% Cu; chalcocite is the main ore mineral rimming subordinate amounts of pyrite and chalcopyrite. The chalcocite blanket sustains current mining operations that make use of heap leach and electrowinning for copper recovery. Hypogene mineralization (0.3–0.8% Cu) extends for 100 to 200 m under the enriched blanket and includes pyrite and chalcopyrite, with specular hematite, minor molybdenite, and trace gold. Andacollo has the highest gold grade among the currently exploited Chilean porphyry copper deposits. In addition, gold deposits occur immediately NW and west of the porphyry copper deposit within a belt 7 km long and 3 km wide (Reyes 1991). The Early Cretaceous age for copper mineralization at Andacollo is inferred from biotite K–Ar ages of 112 ± 10 and 105 ± 3 Ma and whole-rock K–Ar ages of 104 ± 3 and 98 ± 2 Ma (Munizaga *et al.* 1985; Reyes 1991).

Late Palaeozoic–Triassic porphyry Cu–Mo belt

Pre-Mesozoic mineralization is poor in the Chilean Andes, but a belt of porphyry Cu–Mo prospects is distributed along the northern extension of the Late Carboniferous to Early Triassic magmatic arc which extends for over 2500 km from northern Chile to southern Argentina (Mpodozis & Ramos 1989; Camus 2003; Fig. 6.6). This Late Palaeozoic–Triassic belt of porphyry Cu–Mo deposits was first reported by Sillitoe (1977) in the Argentinean Andes, but recent exploration and K–Ar dates ranging from 298 to 195 Ma show that this belt extends to the Andes of northern Chile (Camus 2003). The Late Palaeozoic–Triassic copper occurrences are interspersed with Cenozoic porphyry Cu–Mo deposits of northern Chile (Camus 2003). However, most of the Palaeozoic systems are poorly mineralized and despite systematic exploration none has been proven to be economically viable. The poor mineralization of these porphyry systems might be a result of significant unroofing of Upper Palaeozoic to Lower Triassic igneous rocks, leaving only the roots of the porphyry systems. However, a detailed study of the San Jorge deposit (2 Mt Cu) revealed that high levels of this porphyry Cu–Mo deposit are preserved in the Argentinean Andes (Williams *et al.* 1999) and also shallow epithermal alteration is preserved at the Lila prospect in northern Chile. This may indicate that erosion of pre-Andean rocks probably plays only a minor metallogenic role and thus suggests that conditions for significant copper concentration were not optimal during Late Palaeozoic to Triassic times.

Epithermal precious metal deposits

The Neogene volcanic belt of the Central Andes contains hundreds of hydrothermal alteration zones of epithermal affiliation, and some of which have world-class epithermal precious metal deposits (Maksaev *et al.* 1984; Camus 1990b; Sillitoe 1991; Erickson & Cunningham 1993). Neogene precious metal epithermal deposits occur in the easternmost Andes of northern Chile (Fig. 6.7), but also are present within the Andes and the Puna high plateau of Peru, Bolivia and Argentina. Rich precious metal epithermal deposits occur within the local Maricunga and El Indio belts of northern Chile (Sillitoe 1991; Bissig *et al.* 2002). In addition, a number of mesothermal to epithermal silver-bearing veins and gold-bearing veins of high-sulphidation and low-sulphidation affiliation are associated with Palaeocene–early Eocene volcanic rocks in northern Chile. Some of these precious metal deposits are peripheral to Palaeocene–early Eocene porphyry Cu–Mo deposits.

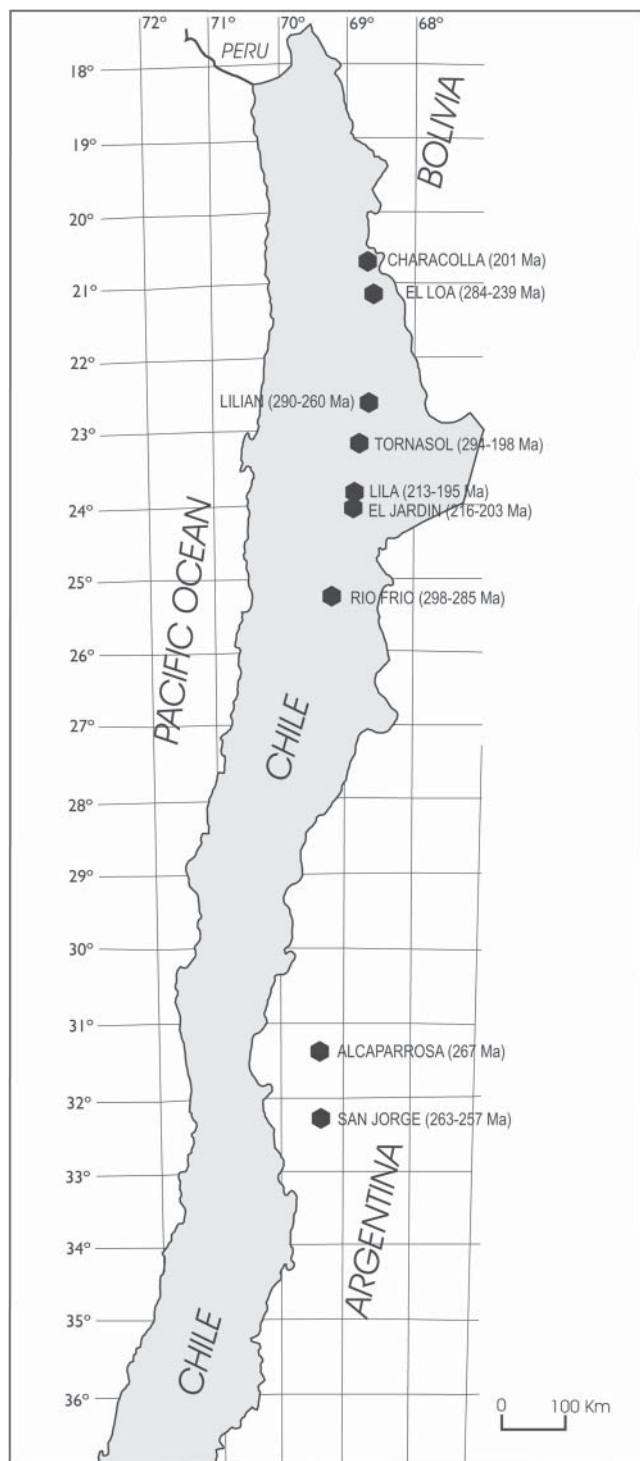


Fig. 6.6. Late Palaeozoic–Triassic porphyry Cu–Mo belt of northern Chile (ages after Camus 2003).

Miocene precious metal epithermal deposits

The most significant precious metal deposits associated with Miocene volcanic rocks occur in the High Andes of Chile and Argentina from 27°S to 30°S (Fig. 6.7), and from north to south are: La Coipa (c. 6 million ounces (Moz) of Au equivalent), Pascua (c. 12 Moz Au) and El Indio (c. 12 Moz Au). The Chilean Neogene precious metal epithermal deposits are predominantly of the high-sulphidation type according to the classification of Hedenquist (1987). The deposits basically

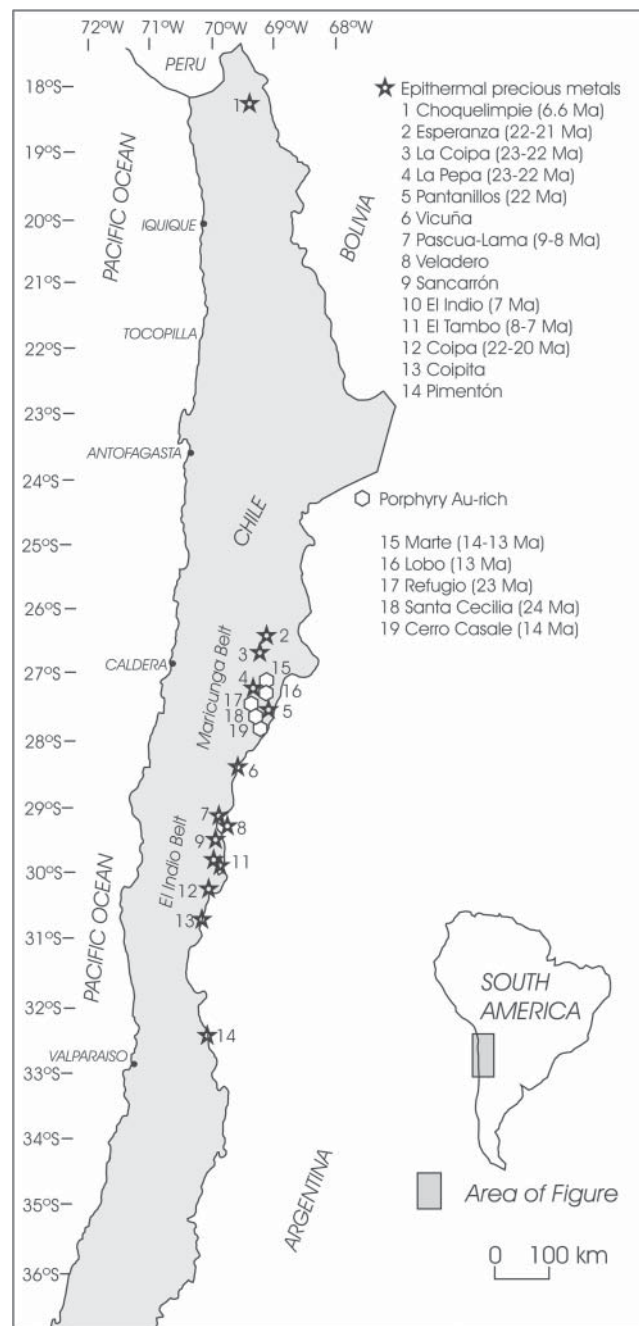


Fig. 6.7. Miocene precious metal epithermal deposits and Au-rich porphyry deposits of northern Chile (ages after Gröppner *et al.* 1991; Moscoso *et al.* 1993; McKee *et al.* 1994; Bissig *et al.* 2001, 2002; Muntean & Einaudi 2001).

correspond to complex fault-controlled vein systems (e.g. El Indio, La Pepa), but some also include breccia-hosted orebodies (Choquelimpie, Esperanza, El Tambo) and disseminated ores (La Coipa, Pascua). The rich El Indio vein system includes an early high-sulphidation copper mineralization dominated by massive enargite–pyrite–alunite veins with advanced argillic and quartz–sericite halos, and later quartz–gold bonanza-type veins with alteration halos of quartz–illite. The gold-bearing veins of El Indio show that tennantite associated with Au mineralization replaced enargite, indicating an evolution to relatively lower sulphidation states during gold mineralization (Jannas *et al.* 1990, 1999). In addition, the Rio del Medio banded quartz–rhodochrosite vein located 4.5 km

north along-strike of El Indio is akin to low-sulphidation Au–Ag–base metal systems (Jannas *et al.* 1999).

The host rocks to these epithermal precious metal ore deposits are Miocene volcanic rocks, commonly associated with conspicuous bleached areas with advanced argillic alteration. El Indio is hosted by a Miocene rhyolitic to dacitic tuff unit (Amiga tuff, 27–21 Ma; Jannas *et al.* 1999). Choquelimpie occurs within a dacite dome complex in the core of an andesitic stratovolcano dated at 6.6 ± 0.2 Ma (Gröpper *et al.* 1991), whereas La Coipa and Esperanza are associated with 22–24 Ma dacite dome complexes (Oviedo *et al.* 1991; Vila 1991; Moscoso *et al.* 1993). In contrast, the Miocene gold dissemination and gold-bearing stockworks of Pascua are hosted by Late Palaeozoic granite and rhyolite (Bissig *et al.* 2000), and at La Coipa the early Miocene Ag–Au mineralization hosted by Miocene dacite continues in the underlying basement formed by Triassic shale beds (Oviedo *et al.* 1991). The conspicuous and extensive hydrothermal alteration zones that are associated with the epithermal deposits are largely of high-sulphidation type, and residual vuggy silica ledges flanked by quartz–alunite zones that grade outwards to kaolinite– dickite and illite–smectite to chloritized rocks are a common feature. Both structural and lithological control of hydrothermal alteration occurs, and contrasting intensity of hydrothermal alteration is common within the altered areas. Hypogene and supergene alunite are widespread in the altered rocks, and abundant within some of the ore deposits, namely Choquelimpie, La Coipa, Pascua–Lama, El Tambo and El Indio. However, quartz–illite alteration halos are related to the gold-bearing quartz veins at El Indio, which are overprinted on previous quartz–alunite alteration (Jannas *et al.* 1999). In contrast, the low-sulphidation Rio del Medio vein is hosted by dark chloritized Miocene andesites.

Since 1980, mines at El Indio and El Tambo have produced 126 t Au, 630 t Ag and 0.32 Mt Cu, with much of the Au coming from high grade ore (El Indio produced 191 000 t of direct shipping ore that averaged 209 g/t Au), but mining in the district ended in 2002. The early massive enargite–pyrite veins of El Indio are banded veins comprising an alternation of dark enargite–pyrite with light alunite crystalline bands; the late rich gold–silver-bearing quartz veins cross-cut these. Enargite also occurs at depth in Choquelimpie (Gröpper *et al.* 1991) La Coipa (Oviedo *et al.* 1991) and in at least one vein at La Pepa (Sillitoe 1991), but this mineral phase is not a significant ore mineral in these deposits.

Porphyry-type stockwork Cu–Mo mineralization occurs about 2 km south of El Indio epithermal deposit and at a lower topographic level. This porphyry occurrence appears to be part of the same overall metallogenic event (Araneda 1982; Walthier *et al.* 1985; Siddeley & Araneda 1986, 1990). Some other deposits (Choquelimpie, La Coipa and Pascua) are also thought likely to possess genetically related porphyry-type mineralization, but erosion has been insufficient to expose it (Sillitoe 1991). Although marked canyon incision is common in the High Andes, particularly in the Maricunga and El Indio areas, steam-heated alunite–quartz alteration with native sulphur occurs in the upper topographic sections of El Tambo, El Indio, La Pepa, Pascua and La Coipa, indicating that only limited unroofing of these deposits (probably in the order of 100 to 200 m) has taken place since their formation (Bissig *et al.* 2002). Limited fluid inclusion data from one vein at Choquelimpie suggest that mineralization was associated with boiling hydrothermal fluids at a minimum depth of 300 m below the palaeo-surface: this deposit is located at 4860 m altitude within the core of an eroded volcanic edifice that has a maximum elevation of 5300 m (Gröpper *et al.* 1991). Despite abundant hydrothermal alteration zones between latitudes 19°S to 26°S, no epithermal deposits are known in the Chilean Andes along the Miocene to Recent volcanic chain. This latitude is coincident with the hyperarid Atacama Desert where Miocene volcanoes and

domes preserve their original landforms, and it appears that erosion has been insufficient to expose potential epithermal mineralization associated with the volcanic chain in this part of the Chilean Andes.

The epithermal deposits of El Indio and Maricunga districts occur over the flat-slab subduction segment of the active continental margin (27–33°S) where active volcanism is absent (e.g. Thorpe *et al.* 1984). A progressive shallowing of the Benioff zone under the continental border has developed since around 20 to 17 Ma (Jordan *et al.* 1983; Isacks 1988; Allmendinger *et al.* 1983; Kay *et al.* 1987, 1988). This flattening of the subducting oceanic plate was accompanied by compressive deformation and crustal thickening, eastward migration of subduction-related magmatism, waning of main arc volcanism by c. 10 Ma, and complete termination of volcanic activity within this segment by c. 5 Ma when the modern flat-slab was completely established (Kay *et al.* 1999). According to Kay *et al.* (1999), Au and Cu mineralization at El Indio and Maricunga districts formed during the waning stages of arc volcanism, preceding either the eastward migration of the arc front or the actual end of igneous activity. In this interpretation, precious metal mineralization was ascribed to the 13–10 Ma period when hydrous, hornblende-based, residual assemblages that were in equilibrium with the erupted magmas dehydrated by amphibole breakdown to yield high-pressure, garnet-bearing assemblages in the crust. However, Bissig *et al.* (2000, 2001) reported new $^{40}\text{Ar}/^{39}\text{Ar}$ ages for ore-related minerals at El Indio (6.21 ± 0.26 and 7.62 ± 0.29 Ma), Rio del Medio (7.08 ± 0.19 Ma), El Tambo (7.08 ± 0.19 , 7.97 ± 0.37 and 8.24 ± 0.29 Ma), Pascua (8.10 ± 0.19 and 8.73 ± 0.23 Ma) and Lama (9.00 ± 0.22 , 9.40 ± 0.16 Ma). Based on these time constraints Bissig *et al.* (2000, 2001) argue that epithermal mineralization took place during the late Miocene in El Indio district during an episode of crustal thickening and uplift, in association with magmas largely derived from a deep crustal source anatectic environment in which garnet was a residual phase. This interpretation contrasts with the metallogenic model proposed by Kay *et al.* (1999) and Kay & Mpodozis (2001), who argued that no mineralization took place under these conditions.

At Choquelimpie only oxidized ore was mined (open pit and heap leaching), which in general extends to depths of 10 to 40 m, but locally reaches down to 130 m in deep oxide troughs along fracture-controlled breccia bodies. In contrast, at La Coipa silver-bearing jarosite, Ag-halides and gold dissemination resulted from oxidation within the host Miocene dacitic rocks. Thus, supergene oxidation was essential to produce economic ore in these low-grade, bulk-mining deposits. Oxidized ore also occurs at Pascua, especially in the section previously known as Nevada, but most of the Pascua orebody is hypogene. Similarly, although local oxidation occurs at El Indio (e.g. Indio Norte; Walthier *et al.* 1985; Jannas *et al.* 1990), most of this Au (–Ag, Cu) ore is hypogene.

Palaeocene–early Eocene precious metal deposits

The Palaeocene–early Eocene metallogenic belt includes a great number of mesothermal to epithermal silver-bearing veins (Challacollo, Huanchaquita, El Inca, Caracoles, Cachinal de la Sierra, Vaquillas, Sierra Juncal, Tres Puntas–Chimberos and Lomas Bayas districts; Fig. 6.8) and epithermal gold-bearing veins of both high-sulphidation affiliation (El Guanaco, Cachiyuyo de Oro districts; Fig. 6.8) and low-sulphidation affiliation (Faride, El Peñón, San Cristobal deposits and Los Morteros district; Fig. 6.8). This precious metal mineralization is associated with Palaeogene volcanic rocks and locally with volcanic calderas (i.e. Cachinal de la Sierra, El Guanaco (Puig *et al.* 1988) and the Lomas Bayas district south of Copiapó (Díaz *et al.* 1998)). The preservation of shallowly formed Palaeocene deposits was due to the unusually low denudation rates of the central Atacama Desert during the Cenozoic era (Boric *et al.* 1990; Maksiyev 1990; Mpodozis *et al.* 1999).

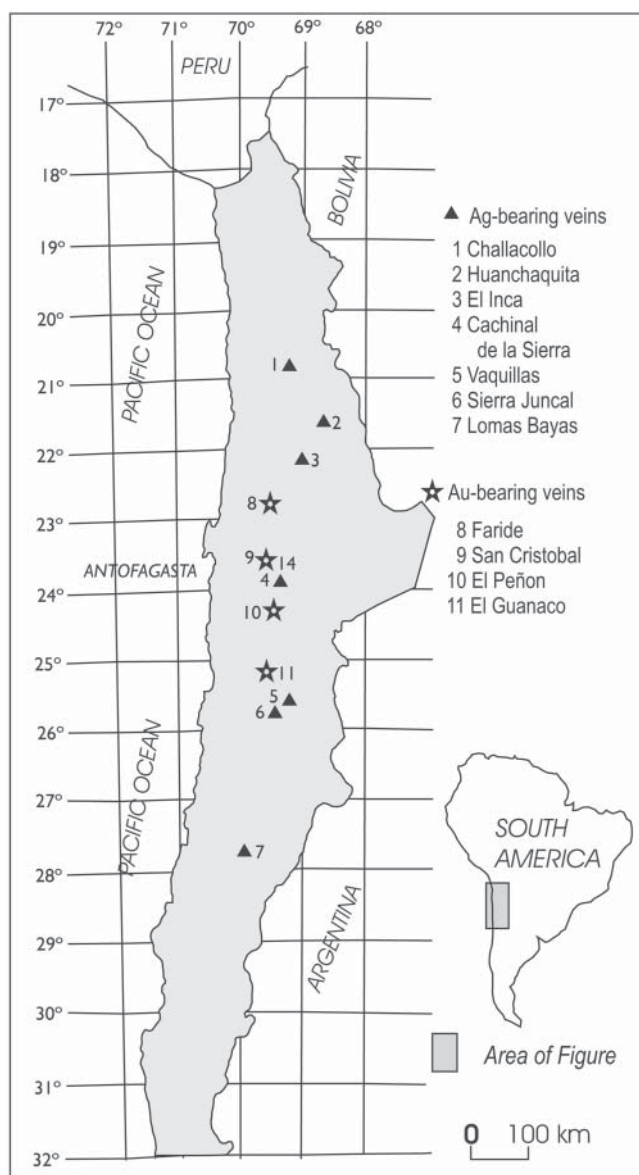


Fig. 6.8. Palaeocene–early Eocene precious metal deposits of northern Chile.

Iron oxide copper–gold deposits

Copper–gold deposits associated with iron oxides have been known since the seventeenth century in the Punta del Cobre district some 15 km south of Copiapó in northern Chile (Fig. 6.9). This mining district includes a number of veins, tabular or lens-like breccia bodies, as well as stockwork and stratabound Cu–Au deposits with Fe oxide gangue (Marschik & Fontboté 1996; Marschik *et al.* 1997; Sillitoe 2003). However, this type of deposit has only become really significant in Chile since the discovery of the Candelaria deposit in 1986 (which has resources of 479 Mt at 0.95% Cu, 0.22 g/t Au, and 3.1 g/t Ag; Marschik 2001) and Manto Verde in 1988 (which has resources of 130 Mt at 0.76% Cu at mine start-up in 1995, plus additional 180 Mt at 0.5% Cu further discovered; Zamora & Castillo 2001). A detailed review of this type of Andean deposit has been published by Sillitoe (2003).

Candelaria is the western extension of the Punta del Cobre district and lies within the contact metamorphic aureole of a granodioritic batholith with $^{40}\text{Ar}/^{39}\text{Ar}$ dates ranging from 117.2 ± 1.0 to 110 ± 1.7 Ma, which is exposed 1 km west of the

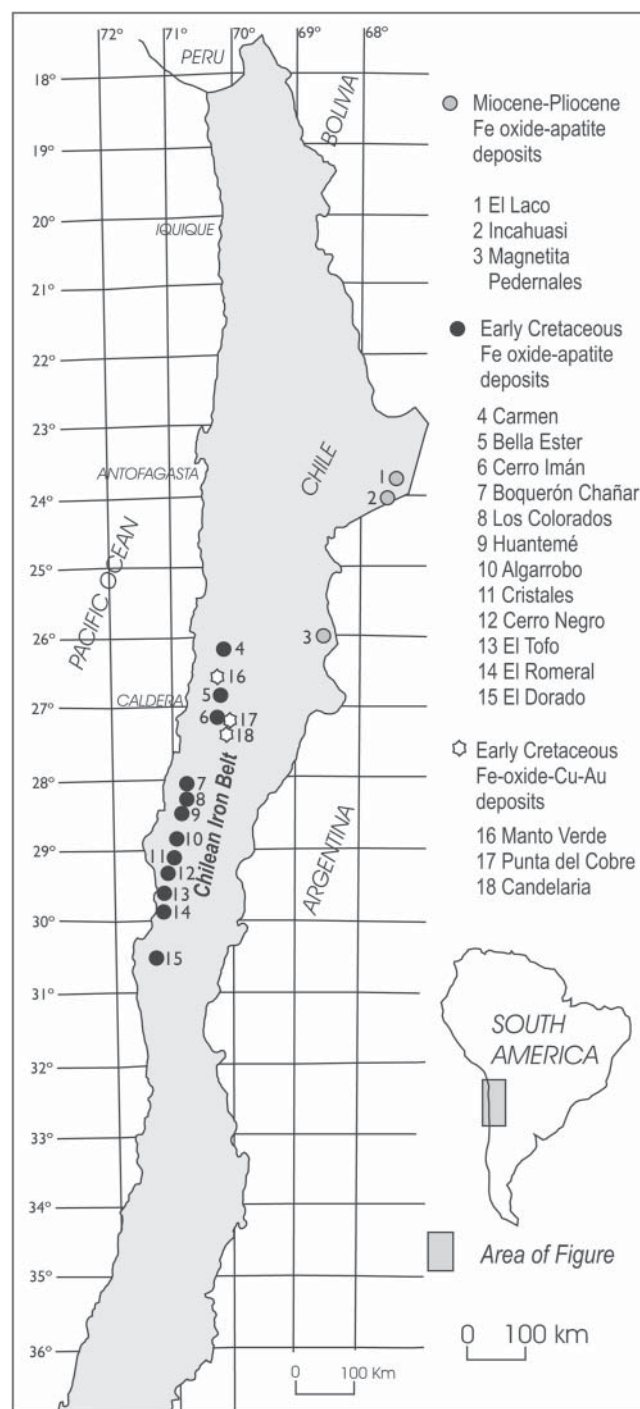


Fig 6.9. Early Cretaceous Fe oxide apatite and Fe oxide Cu–Au deposits of northern Chile. Miocene–Pliocene Fe oxide apatite deposits are also shown.

deposit (Marschik 2001; Marschik & Fontboté 2001a). The bulk sulphides that carry the Cu–Au mineralization at Candelaria are related to pervasive Na–Ca metasomatism (actinolite–albite–scapolite–quartz–K-feldspar–hornblende–hedenbergite) superimposed on a previous pervasive biotitization and magnetite metasomatism that affected Lower Cretaceous andesites and andesitic tuff-breccia strata (Punta del Cobre Formation; Arévalo 1994). The stage of Ca–Na metasomatism remobilized and introduced new magnetite that locally forms up to 20% of the host rocks. The metasomatic ore bodies at Candelaria are roughly tabular, indicating

lithological control of the mineralization, although they occur immediately adjacent to regional faults in which occur magnetite-bearing mylonites. The dominant sulphide is chalcopyrite accompanied by minor pyrite (chalcopyrite/pyrite = 5/1) and pyrrhotite, and the gold occurs as micrometre-sized inclusions within the chalcopyrite. Although the hydrothermal alteration mineral assemblage related to mineralized rocks at Candelaria indicates high temperatures ($>450^{\circ}\text{C}$), an immediate igneous source for mineralizing fluids has not been identified. Re/Os molybdenite dates of 114.2 ± 0.6 and 115.2 ± 0.6 Ma were reported by Mathur *et al.* (2002) and were interpreted as mineralization ages. These ages are coincident with the 115.14 ± 0.18 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of biotite associated with chalcopyrite–pyrite reported by Marschik & Fontboté (2001a) and the biotite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages of 114.2 ± 0.8 and 114.1 ± 0.7 Ma of Ullrich & Clark (1999). A younger amphibole $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 111.7 ± 0.8 Ma (Ullrich & Clark 1999) and similar biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 111.0 ± 1.7 and 110.7 ± 1.6 Ma (Arévalo *et al.* 2000) probably represent a later event of alteration or mineralization (Mathur *et al.* 2002). The timing of the mineralization at Candelaria indicates that the deposit was formed after the deposition of about 2000 m of carbonate marine sediments during Early Cretaceous times over the host volcanic strata (Chañarcillo Group; Segerstrom 1967; Arévalo 1994). Therefore, Candelaria was originally formed under this *c.* 2000-m-thick carbonate sedimentary cover.

At Manto Verde Cu–Au mineralized breccias with an iron oxide matrix occur within transtensional domains along the Manto Verde fault, a subsidiary branch of the regional strike-slip Atacama Fault System (Vila *et al.* 1996; Zamora & Castillo 2001). Breccias with specularite matrix occur at the Manto Verde North orebody, whereas breccias with a magnetite matrix are present at Manto Verde South pit. Only oxidized copper ores are currently mined, with oxidation extending down to 200 m and the two near-surface oxidized ore bodies being connected to a larger underlying hypogene sulphide orebody. Hypogene sulphides are dominated by chalcopyrite and pyrite (chalcopyrite/pyrite = 5/1) that are disseminated within the specularite and magnetite matrix of breccias, with a magnetite matrix dominating at depth. A pervasive potassic alteration characterized by intergrowths of K-feldspar and chlorite and minor quartz and hematite affected the volcanic and intrusive host rocks at Manto Verde. Preliminary fluid inclusion studies revealed relatively low homogenization temperatures (180 – 250°C), and high saline content (30–50% NaCl equivalent), and trapping of boiling hydrothermal fluids is suggested. Whole-rock K–Ar ages of altered andesites and dykes of 117 ± 3 and 121 ± 3 Ma were initially taken to indicate the age of the primary mineralization process (Vila *et al.* 1996), but a more precise zircon U–Pb age of 128.9 ± 0.6 and a titanite U–Pb age of 126.4 ± 0.5 Ma for a quartz monzonite to granodioritic dyke with potassic alteration, led Gelcich *et al.* (2003) to conclude that mineralization at Manto Verde is most probably related to the cooling and differentiation of the Sierra Dieciocho pluton, which forms the eastern margin of the deposit.

The Chilean iron oxide copper–gold deposits belong to a class of deposits characterized by abundant Ti-poor oxide mineralization referred to as the iron oxide (Cu–U–Au–REE) class or categorized as low-sulphur Cu–Au deposits (e.g. Hirtzman *et al.* 1992; Barton & Johnson 1996). The genetic relationships with contemporaneous plutonic rocks are controversial (see review of Sillitoe 2003). Some authors favour a model where metal-bearing fluids exsolve from crystallizing magma and deposit metals in the adjacent country rocks (e.g. Gow *et al.* 1994; Rotherham *et al.* 1998; Williams 1998; Williams *et al.* 1999). An alternative suggestion is a model of evaporite-derived, thermally driven fluids to leach and redeposit metals (Battles & Barton 1995; Barton & Johnson 1996; Ullrich &

Clark 1999). Both of these models have been proposed for Candelaria (e.g. Ullrich & Clark 1999; Marschik & Fontboté 1996). Potassic and Na–Ca high-temperature alteration assemblages related to copper sulphide deposition at Candelaria and Agustina mine at Punta del Cobre indicate that the Cu–Au mineralization processes took place at temperatures well above 450°C (Ryan *et al.* 1995; Hopf 1990). Hence, the origin of these deposits is likely to have a magmatic connection, as supported by Re–Os isotopic data, even though a hypothetical evaporite fluid source cannot be completely discarded (Mathur *et al.* 2002). A magmatic–hydrothermal origin for the copper–gold mineralization of Manto Verde has been postulated by Vila *et al.* (1996); fluid inclusion data indicate formation at *c.* 180 – 320°C with boiling of hydrothermal fluids during mineralization. Manto Verde appears to be a relatively shallow representative of this type of copper–gold mineralization.

Iron oxide–apatite deposits

Chile is currently (2005) only a modest iron ore producer (8.0 Mt of iron oxide ore were produced in 2004), although the so-called ‘Chilean Iron Belt’ comprises more than 40 Fe oxide–apatite (Ti-poor) deposits and prospects contain resources exceeding 1000 Mt of ore at 60% Fe. This belt of iron deposits extends some 600 km along the Coastal Cordillera from $25^{\circ}30'\text{S}$ to $32^{\circ}00'\text{S}$ (Ménard 1995; Espinoza 1990), but only Los Colorados, El Algarrobo and Romeral deposits are currently mined (Fig. 6.9). These iron deposits occur within the domain of the southernmost segment of the Atacama Fault System. This is a major regional sinistral strike-slip fault system that extends more than 1000 km along the Coastal Cordillera of northern Chile, and was developed through Jurassic–Early Cretaceous times as an intra-arc fault related to oblique subduction of the Aluk plate (Boric *et al.* 1990; Scheuber & Andriessen 1990). The main iron-bearing ore bodies are irregular, formed of massive magnetite with minor specularite and local traces of pyrite and chalcopyrite. Host rocks are mostly Lower Cretaceous meta-andesites adjacent to *c.* 100 – 120 Ma (K–Ar) dioritic to granodioritic plutons or meta-andesitic roof pendants on these intrusions, although massive magnetite veins hosted by diorites also occur. The economically relevant ore bodies are 100 to 1000 m long, up to 200 m wide and up to 650 m high, although more minor magnetite veins and lenticular ore bodies are also common along the belt. Martite occurs as a product of supergene oxidation of the hypogene magnetite, and locally magnetite has been completely transformed to hematite. Although gangue minerals are generally rare, the iron ores contain apatite (chlorapatite), actinolite and chlorite. The ores are massive, black in colour, and hard, but to some extent granular as they are formed of octahedral crystals of magnetite (martite). Minor pyrite and chalcopyrite locally occur.

The andesitic country rocks of the iron oxide ores show a strong alteration that consists of a mineral assemblage of actinolite–scapolite–biotite–tourmaline–chlorite–chlorapatite–sphene–minor garnet and pyrite. This high-temperature alteration mineral assemblage commonly obliterates the original porphyritic texture of the host andesites in the vicinity of the iron ore bodies, passing outward into quartz–albite \pm K-feldspar \pm biotite alteration and silicified \pm tourmalinized \pm argillized rocks (Naslund *et al.* 2002). K–Ar ages from the exploited Romeral, Algarrobo and Los Colorados deposits range between 112 and 108 Ma (Munizaga *et al.* 1985), but the K–Ar data for the iron belt range from 128 to 100 Ma (Oyarzún *et al.* 2003). Recent U–Pb dating of apatite from a magnetite–apatite vein 500 m south of the main Carmen pit yielded 129.8 ± 3.0 Ma, coincident with the zircon U–Pb age of 130.6 ± 0.3 Ma for a quartz dioritic stock in the Carmen northern pit, and with the U–Pb ages in the Sierra Aspera pluton immediately west of Carmen (Gelcich *et al.* 2003).

Another belt of Fe oxide deposits of Cenozoic age (El Laco, Incahuasi and Magnetita Pedernales; Fig. 6.9) occurs in the High Andes and is related to late Miocene and Pliocene volcanic edifices. The El Laco deposit is located at elevations of between 4600 and 5200 m a.s.l. in a Pliocene stratovolcano, and comprises seven discrete bodies of massive magnetite within an 8×4 km area. The magnetite bodies aggregate some 500 Mt of ore at about 60% Fe. Park (1961) proposed that the El Laco magnetite is consolidated lava, occurring as a series of partly preserved flows and related feeder dykes. Several subsequent investigators have supported this concept based on textural features displayed by the magnetite (Naslund *et al.* 2002, and references therein). Others support a metasomatic replacement origin for El Laco based on detailed geological and laboratory studies (Rhodes & Oreskes 1999; Sillitoe & Burrows 2002).

A longstanding controversy persists about the origin of the Chilean iron oxide-apatite (Ti-poor) ore deposits. Some authors (e.g. Nyström & Henríquez 1994, 1995; Naslund *et al.* 2002) contend that many such deposits are derived by melt crystallization, compare them to the Kiruna deposits of Sweden, and relate the mineralization to injections of iron-rich magmas at subvolcanic and volcanic levels (Vivallo *et al.* 1994; Nyström & Henríquez 1994, 1995; Gelcich 1999; Naslund *et al.* 2002). On the other hand, others support the proposal that these iron deposits have a metasomatic replacement origin (e.g. Bookstrom 1995; Sillitoe 2003). The hypothetical iron metasomatism was accompanied by pervasive actinolitization of the volcanic host rocks in the temperature range 475–550°C and at a pressure of about 2 kbar (Ruiz *et al.* 1965, 1968; Bookstrom 1977, 1995; Ménard 1995; Hirtzman *et al.* 1992; Sillitoe & Burrows 2002). Ménard (1992, 1995) proposed that the Chilean magnetite-apatite mineralization was produced by the emplacement of deep-seated (>4 km), subduction-related, anhydrous (pyroxene-bearing) gabbros and diorites. These mafic intrusions evolved under relatively high oxygen fugacities and developed the iron oxide mineralization and associated sodic alteration at high temperature (600–450°C).

Stratabound copper-(silver) deposits

Stratabound copper deposits with subordinate silver, long known as 'Chilean Manto-type', occurs along the Coastal Cordillera of northern Chile hosted by Jurassic and Lower Cretaceous volcanic and volcanosedimentary rocks. These Mesozoic stratabound deposits historically were the second source of Chilean copper production after Cenozoic porphyry copper deposits, though have recently been displaced to third place by the exploitation of the large Lower Cretaceous iron oxide copper-gold deposits at Candelaria and Manto Verde (Marschik & Fontboté 2001a; Vila *et al.* 1996; Zamora & Castillo 2001).

Two groups of significant stratabound Cu-(Ag) occur along the Coastal Cordillera of northern Chile. The first of these is seen between 21°30'S and 26°S, and is hosted by a Jurassic volcanic sequence (La Negra Formation; Kojima *et al.* 2002). The second group occurs from 30°S to 34°S and is hosted by Lower Cretaceous volcanic and volcanosedimentary rocks (Zentilli *et al.* 1997; Wilson & Zentilli 1999) (Fig. 6.10). Together these record Late Jurassic and uppermost Early Cretaceous metallogenic episodes (see the review of Maksaev & Zentilli 2002). These unusual copper deposits do not fit neatly into any of the major ore deposit models but are characteristic of the first stage of Andean evolution, with an extensional setting for arc magmatism along the active margin of South America. The mineralization took place at the same time as structurally controlled emplacement of batholiths within the Mesozoic volcanic and sedimentary strata.

The largest stratabound copper deposit of the Jurassic belt is Mantos Blancos where more than 200 Mt of ore at 1.2% Cu has

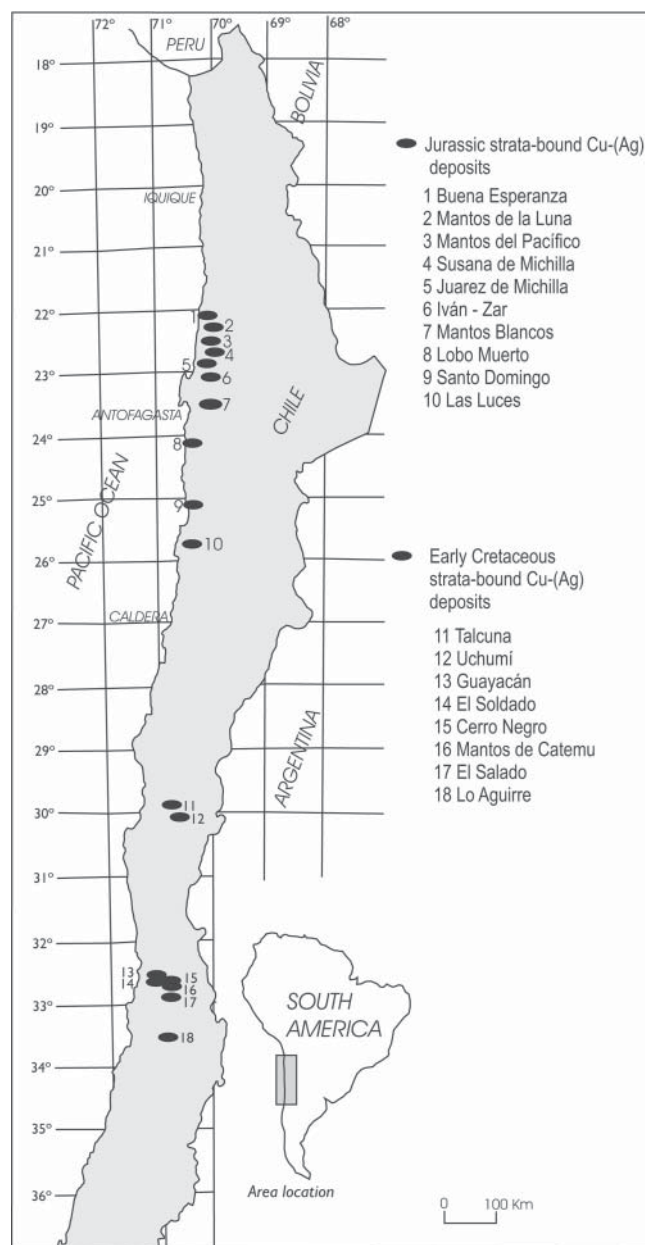


Fig. 6.10. Volcanic-hosted stratabound Cu-(Ag) deposits of northern Chile.

been extracted since 1960; the reserves are 31 Mt of oxidized ore at 0.72% Cu, 24 Mt of oxidized ore at 0.41% Cu, and 56 Mt of sulphide ore at 1.01% Cu and 12 g/t Ag (Minería Chilena 2003). The second largest in volume is Mantos de la Luna with measured resources of 40.5 Mt at 1.39% Cu (0.4% Cu cut-off grade) plus inferred resources of 6 Mt at 1.4% Cu (Minería Chilena 2001). Other deposits of this type are considerably smaller, but overall reserves and production add up to some millions of tonnes of ores at 1 to 3.8% Cu and 8 to 25 g/t Ag. The most relevant of these are at Susana and Juarez (Michilla district) and the Santo Domingo deposit, with other examples being the Buena Esperanza, and Mantos del Pacífico deposits (Fig. 6.10).

Most of the stratabound Cu-(Ag) deposits are hosted by mafic Jurassic basaltic to andesitic porphyritic lavas or breccia bodies, but the largest deposit of this group, Mantos Blancos, is hosted by a bimodal suite of rhyolitic and andesitic rocks, which according to ongoing investigations led by C. Palacios are mineralized sills, dykes and intrusive breccia pipes.

Stratabound Cu–(Ag) deposits occur near gabbroic, dioritic or andesitic subvolcanic intrusive bodies, such as dykes, sills, stocks or volcanic necks, but these intrusives in most cases are unmineralized (e.g. Buena Esperanza, Susana, Santo Domingo; Palacios 1986; Espinoza *et al.* 1996). The intrusives have been interpreted as feeder conduits of the Jurassic volcanism (Palacios & Definis 1981a, b; Espinoza 1981; Espinoza *et al.* 1996). Mantos Blancos includes at least four main lenticular disseminated and fracture-filling sulphide ore bodies (Sorpresa, Aida, Nora and Marina) forming an overall, slightly unconformable, subhorizontal tabular ore deposit (Chavez 1985). The thickness of mineralized levels at Mantos Blancos ranges from 150 to 350 m and the deposit extends irregularly over an area of 2.6 by 1.2 km. At Buena Esperanza and Susana copper sulphides occur within the matrix of a central breccia pipe and are disseminated in a number of conformable stratiform orebodies ('mantos') around the breccia pipe (Palacios 1990; Espinoza *et al.* 1996). The stratiform orebodies (2 to 25 m thick) are commonly restricted to the amygdaloidal and brecciated sections of the Jurassic lava flows and minor veins occur along local faults and fractures (Palacios & Definis 1981a, b; Dreyer & Soto 1985).

The main hypogene sulphides are chalcocite and bornite, minor chalcopyrite, and locally covellite and digenite. Gangue minerals are quartz, hematite, pyrite, chlorite and calcite. Minor magnetite occurs as dissemination within the mineralized rocks, but is mostly replaced by hematite or maghemite. At Mantos Blancos specularite occurs; this crystallized early in the paragenesis and is mostly concentrated within a barren andesite flow that overlies the orebody, unlike fine reddish hematite dissemination, which occurs within mineralized rocks. A lateral hypogene zonation has been described at Mantos Blancos and Santo Domingo. This includes copper-rich cores dominated by chalcocite–bornite–digenite surrounded by a rim of bornite–chalcopyrite or sole chalcopyrite, and an external, mostly uneconomic halo of chalcopyrite–pyrite ore (Chavez 1985; Definis 1985). The hydrothermal alteration assemblage of albite–chlorite–hematite–quartz–calcite–epidote–sphene–scapolite–anatase–minor sericite is associated with ore minerals in these stratabound copper deposits, but the primary textures of the volcanic rocks are preserved (Losert 1973; Chavez 1985; Palacios 1986, 1990). This local alteration is thought to be superimposed upon a regional alteration/metamorphism (prehnite–pumpellyite facies) of the volcanic sequence, but little or no alteration contrast occurs between mineralized and barren volcanic country rocks. The hydrothermal alteration is particularly pervasive in Mantos Blancos, where lithogeochemistry shows significant metasomatism of the host rocks with the addition of Na, Fe and Mg (Chavez 1985), although deeper parts of the orebody within intrusive breccias shows potassic alteration dominated by K-feldspar.

Most of the stratabound copper deposits have an upper oxidized zone that extends to a maximum depth of 250 m. The degree of supergene oxidation is variable: some deposits include almost exclusively copper oxides (e.g. Mantos de la Luna, Juarez), whereas others are formed exclusively of hypogene sulphides (e.g. Buena Esperanza). The boundary between the supergene and hypogene ores is gradual, so that some deposits include a zone with a mixture of copper sulphides and oxides (e.g. Susana, Santo Domingo). The oxidized ores include mainly atacamite, minor chrysocolla, malachite, copper sulphates, and rare cuprite and native Cu. Copper grades are similar in the oxidized upper section and the underlying hypogene sulphide zone, so that no significant Cu transport occurred during the supergene oxidation (*in situ* oxidation). Only the largest deposits such as Mantos Blancos and Susana have local enriched pockets with supergene chalcocite-group minerals and covellite (Chavez 1985; Wolf *et al.* 1990). The poor development of supergene enrichment within the stratabound copper deposits could be explained by the insufficiency of hypogene pyrite to release supergene acid solutions under oxidation,

and the profuse occurrence of calcite gangue that may have readily neutralized any supergene acid solutions, so precluding the leaching of metallic cations from the oxidized zone.

Sparse and imprecise K–Ar and Rb–Sr minimum radiometric dates indicate that the main stratabound deposits hosted by Jurassic volcanic rocks of northern Chile were formed at about 150–140 Ma (Boric *et al.* 1990; Venegas *et al.* 1991; Tassinari *et al.* 1993; Vivallo & Henríquez 1998). This radiometric age range is younger than the volcanic host rocks that have been dated by Rb–Sr and K–Ar methods in the 186–165 Ma range, but overlapping with $^{40}\text{Ar}/^{39}\text{Ar}$, Rb–Sr and K–Ar ages of granodioritic batholiths that intrude the volcanic sequence along the coastal area (age compilation in Vivallo & Henríquez 1998). Minor stock and dykes spatially related to the stratabound Cu–(Ag) deposits have minimum K–Ar ages ranging between 154 and 133 Ma, except for the gabbroic stock at Buena Esperanza that has yielded a K–Ar age of 168 ± 5 Ma in plagioclase (Boric *et al.* 1990).

Stratabound copper deposits hosted by Lower Cretaceous volcanic rocks occur in central Chile (30–34°S). The largest deposit is El Soldado, whose production plus resources totals over 200 Mt at 1.4% Cu (Boric *et al.* 2002). The second largest in volume is the now-exhausted Lo Aguirre deposit where 19 Mt at 1.66% total Cu (0.98% soluble Cu) were exploited (Saric *et al.* 2003). El Soldado copper deposit is located about 120 km north of Santiago, in the Coastal Cordillera of central Chile (Fig. 6.10). It is hosted by a sequence of felsic and basaltic units dipping *c.* 30° to the east of the Upper Member of the Lower Cretaceous Lo Prado Formation (Rivano *et al.* 1993). The deposit consists of numerous isolated orebodies with intervening barren zones, distributed in about a dozen individual orebody clusters or blocks. These blocks are spatially distributed within a volume about 2 km long by 1 km wide and 600 m in vertical extent (Wilson & Zentilli 1999). Within the blocks, individual subvertical orebodies are extremely variable in size, from very small to 450 m long, 150 m wide and 450 m in vertical extent (Boric 1997; Boric *et al.* 2002). Although described as stratabound on a regional scale, in detail the El Soldado orebodies are distinctly discordant, displaying a strong structural control (Ruge 1985). The orebodies are preferentially developed within a generally north–south to NNW regional fracture system, especially where north–south, east–west and NW faults intersect. Away from these zones of structural permeability, orebodies can be best described as veins (Boric 1997). Lithological control is exerted by the relatively more brittle felsic rhyolite flows (domes) and their feeders, which are more richly mineralized in comparison to basalt flows, tuffs, and underlying sedimentary rocks (Ruge 1985). Andesitic–basaltic dykes are generally barren (Boric 1997).

Hypogene ore minerals are chalcopyrite, bornite and chalcocite and occur as dissemination and veinlets, largely filling primary and secondary porosity of the volcanic host rocks. According to Boric (1997) many individual orebodies in a block show mineralogical zoning: a core of chalcocite–hematite or chalcocite–bornite–hematite is followed outwards by approximately concentric zones of bornite–chalcopyrite–chalcopyrite, and pyrite in the most external zone. The deeper roots of the orebodies contain relatively more pyrite than their upward terminations and are surrounded by a halo of pyrite dissemination. This pyrite halo and the deep pyrite bodies are early, low-temperature, diagenetic and probably biogenic in origin, related to basinal, often petroleum-rich fluids. Most of these deposits have formed in two stages: one low temperature, near the age of the host rocks, and another hydrothermal episode later. Although oxidized copper zones exist near the surface and mixed ores are exploited, supergene enrichment is not significant. Common waste or gangue minerals are pyrite, hematite, calcite, chlorite, albite, microcline, bitumen and minor amounts of sphalerite, galena and arsenopyrite. Copper grade is extremely variable (Klohn *et al.* 1990). Lateral limits of the

orebodies are characterized by abrupt changes of Cu grade from a nucleus with about 2% Cu to outside zones containing 1.2–0.5% Cu within a few metres (Ruge 1985; Klohn *et al.* 1990). The wall rock between the orebodies is generally barren (<0.15% Cu; Klohn *et al.* 1990). Hydrothermal alteration consists of abundant calcite, chlorite, albite, microcline, epidote, opaline silica, titanite and rutile–anatase, and some sericite and clay minerals, but primary rock textures are largely preserved (Holmgren 1987; Boric 1997).

Highly saline (*c.* 30–40% NaCl equivalent) three-phase fluid inclusions in quartz related to sulphide mineralization at El Soldado yielded homogenization temperatures of 200–257°C (without pressure correction) and no evidence for boiling was observed (Holmgren 1987). The same study found that fluid inclusions within late barren calcite at El Soldado are also highly saline but with lower homogenization temperatures of 82 to 104°C. On the other hand, three-phase fluid inclusions in quartz from quartz–bornite-filled amygdaloids at El Salado deposit yielded homogenization temperatures of 250 to 430°C, and some vapour-rich inclusions suggest trapping of a boiling fluid (Nisterenko *et al.* 1973). El Salado is a minor volcanic-hosted stratiform Cu–(Ag) deposit located 18 km SE from El Soldado and close to the western border of a granodioritic intrusion with a biotite K–Ar age of 118 ± 3 Ma (Rivano *et al.* 1993).

Many radiometric dates (K–Ar, $^{40}\text{Ar}/^{39}\text{Ar}$, Rb–Sr) reported from El Soldado range from 131 to 96 Ma, supporting the Early Cretaceous age for this copper deposit (Boric & Munizaga 1994; Boric 1997). The oldest radiometric dates at El Soldado (from 131 to 118 Ma) are coeval with the stratigraphic Neocomian age of the host volcanics. A group of younger ages from 113 to 96 Ma are considered to represent the alteration/mineralization event: this younger group of radiometric ages includes dates of K-feldspar and albite veinlets associated with copper sulphides that are mostly concentrated between 105 and 101 Ma (Boric 1997). Wilson *et al.* (2003a) have dated ten samples of K-feldspar (adularia) from El Soldado by the step-heating $^{40}\text{Ar}/^{39}\text{Ar}$ method. For hydrothermal K-feldspar in close association with copper sulphide precipitation, the ages range from 100.5 ± 1.5 Ma to 106 ± 1.1 Ma, with a mean of 103 ± 1.3 Ma, which these authors interpret to be the main age of copper mineralization at El Soldado. Fission track dating of apatite in the host rocks yields an age of *c.* 90 Ma, indicating post-mineralization fast cooling of the system (Wilson *et al.* 2003a). According to the radiometric data, mineralization at El Soldado coincides temporally with the K–Ar age range (118–96 Ma; Rivano *et al.* 1993) of Cretaceous batholiths that are emplaced within the Lower Cretaceous volcanosedimentary sequence and crop out some 12.5 km NE and 18 km SE of El Soldado. This is also true for the 117–94 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ age range of the Caleu Pluton that cross-cuts the Lower Cretaceous sequence some 25 km south of El Soldado (Parada & Larrondo 1999; Parada *et al.* 2001a).

At Lo Aguirre, in the western outskirts of Santiago city, copper ores within andesitic lavas and breccias of the Aptian–Albian Veta Negra Formation form a main irregular orebody 600 m long, 200 m wide and 150 m high. Lo Aguirre and two minor satellite orebodies (San Antonio and Carretón) are peripheral to a barren dioritic stock (Saric *et al.* 2003). The hypogene mineralization is a dissemination of chalcocite and bornite with successive outer zones of bornite–chalcopyrite, chalcopyrite–pyrite and pyrite. The copper sulphides concentrate within the most porous levels of the host volcanic rocks and were partly oxidized by supergene processes. Hydrothermal alteration minerals are calcite, quartz, hematite, chlorite, epidote and clay minerals. A whole-rock K–Ar age of 110 ± 4 Ma obtained for an andesite sample, and an Rb–Sr isochron of 113 ± 3 Ma were interpreted to represent the probable age of hydrothermal alteration related to mineralization at Lo Aguirre (Munizaga *et al.* 1988b). However, further $^{40}\text{Ar}/^{39}\text{Ar}$

dating of albite from mineralized rocks yielded 102 ± 5 Ma, which places the mineralization at Lo Aguirre temporally concurrent with El Soldado (Saric *et al.* 2003).

Another group of relatively small stratabound Cu–(Ag) deposits are hosted by Lower Cretaceous sedimentary or volcanoclastic rocks in central Chile (Cerro Negro, Uchumí, Talcuna; Camus 1990a). These are roughly tabular conformable orebodies ('mantos') restricted to a specific stratum and interconnected by poorly mineralized sections. Copper sulphides concentrate in the upper few metres of a specific sedimentary or pyroclastic level that normally underlies either a massive volcanic or mudstone impervious stratum. Typical examples occur at Talcuna (Boric 1985) and Cerro Negro (Elgueta *et al.* 1990). Hypogene minerals are bornite, chalcopyrite, chalcocite, pyrite, and minor sphalerite and galena. Gangue minerals are calcite, chlorite, hematite, epidote, zeolites and local magnetite. At Talcuna stratiform Cu–(Ag) mineralization occurs mostly along a subhorizontal bituminous lapilli tuff level 10 to 15 m thick, but highest-grade ore occurs along the intersection between the lapilli tuff horizon with subvertical NNW-trending veins. These veins have open-space filling textures, evidence of hydraulic fracturing, and have been interpreted as feeders for hydrothermal fluids that mineralized the bituminous lapilli tuff level under an impervious manganese-rich level of tuffaceous sandstones and mudstones (Boric 1985). Present exploitation in the Talcuna district concentrates on NW-elongate ore shoots that developed along the intersection of the above-mentioned lapilli tuff level with veins: rich pockets coincide with the occurrence of abundant bitumen. Coexisting liquid-rich and vapour-rich fluid inclusions in calcite (suggesting trapping of a boiling hydrothermal fluid) associated with copper sulphides at Talcuna orebodies yielded homogenization temperatures of 70 to 170°C and salinities from 5 to 27 wt% NaCl equivalent (Oyarzún *et al.* 1998). These authors interpreted the wide range of salinity variation as the result of complex interaction between boiling of the hydrothermal fluid and mixing with non-saline waters during mineralization at Talcuna. In addition, fluid inclusions in calcite from cavity fillings in andesites in the area of the deposits yielded homogenization temperatures in the range 120–205°C and salinities from 11 to 19 wt% NaCl equivalent. These were taken to represent an earlier stage of mineralization at Talcuna (Oyarzún *et al.* 1998).

The stratabound deposits of this group have been formed within the Lower Cretaceous sequences deposited in shallow marine to intermontane lacustrine sedimentary basins with coeval volcanic deposits (Camus 1990a). These intravolcanic depressions probably were pull-apart basins related to active sinistral strike-slip faults during Early Cretaceous times. In addition, Zentilli *et al.* (1997) and Wilson & Zentilli (1999) documented that bitumen commonly occurs within these stratabound Cu–(Ag) deposits hosted by Lower Cretaceous volcanosedimentary strata (Uchumí, Talcuna, Cerro Negro), suggesting that copper mineralization in these deposits was deposited within degraded petroleum reservoirs.

The origin of the copper stratabound deposits of northern Chile has long been a matter of controversy. Stratiform orebodies were first regarded as syngenetic volcanic exhalative (Ruiz *et al.* 1965, 1971; Stoll 1965). However, their epigenetic origin has now been widely demonstrated, due to the subsequent discovery of unconformable orebodies, the spatial relationship of copper mineralization around Upper Jurassic intrusive stocks and sills, and significant hydrothermal alteration (albite, chlorite, quartz, sericite, calcite, sphene, scapolite and anatase) associated with copper-rich sulphide dissemination (chalcocite, bornite) within the volcanic host rocks (Palacios & Definis 1981a, b; Dreyer & Soto 1985; Espinoza *et al.* 1996). The most widely accepted hypothesis is the hydrothermal derivation of these volcanic-hosted copper deposits, related to subvolcanic intrusive bodies (Espinoza 1981, 1982;

Chavez 1985; Palacios 1990; Espinoza *et al.* 1996), although some authors have also suggested a diagenetic–metamorphic origin (Sato 1984; Sillitoe 1990) or a genetic connection with underlying batholiths (Losert 1974; Vivallo & Henríquez 1998; Maksaev & Zentilli 2002).

There has been some debate over the origin of the stratabound Cu–(Ag) deposits that occur peripheral to Upper Jurassic batholiths emplaced within transtensional sinistral regional faults within the Mesozoic volcanic pile. The prevalent view that these Cu–(Ag) deposits have an inherent genetic relationship with hydrothermal fluid derivation from subvolcanic stocks and dykes was contested by Maksaev & Zentilli (2002), as these minor intrusions are largely unmineralized and this hypothesis does not fit well with Sr, Os and Pb isotopic data that call for contribution of these elements from the country rocks (e.g. Maksaev & Zentilli 2002; Wilson *et al.* 2003b). The Cu–(Ag) stratabound mineralization appears to be the product of hydrothermal fluids of mixed origin that were mobilized during the emplacement and cooling of Upper Jurassic batholiths within the Mesozoic sequence and deposited copper where cooled and mixed with meteoric fluids away from their igneous heat sources.

Copper-, silver- and gold-bearing veins

Numerous copper-bearing veins with iron oxide gangue occur along the Coastal Cordillera of central and northern Chile, especially between latitudes 21°S and 26°S. Upper Jurassic dioritic to granodioritic batholiths host these veins, the most significant of which occur in the Tocopilla, Gatico, Naguayán–Desesperado, Julia and Montecristo districts (Fig. 6.11). These vein districts of the Coastal Cordillera were of primary economic importance in the second half of the nineteenth century and early twentieth century when most of the Chilean copper production came from the exploitation of copper-bearing veins, but they are now long abandoned. The largest veins are Minita–Despreciada (Tocopilla District), Toldo–Velarde (Gatico District), and Julia–Reventón (Julia District). These are mostly NE-trending, steeply dipping veins (though some veins are striking WNW, east–west and north–south), from 750 to 2000 m long, 1 to 12 m wide and with about 370 to 670 m of known vertical extent. Copper ores concentrate in rich pockets along these structures, separated by low-grade or barren sections. The hypogene paragenetic sequence of the largest veins is tourmaline–actinolite–quartz–magnetite–hematite–pyrite–chalcopyrite–bornite–calcite (Boric *et al.* 1990). Strong silicification, argillite alteration and chloritization occur within these copper-bearing veins, and extend some metres into their intrusive wall rocks. The hypogene minerals fill fractures and openings, either as irregular and discontinuous veinlets or massive pockets with banded textures, or as fine dissemination. The structure of the Cu-bearing veins is regular and continuous within intrusive bodies, but quite irregular and discontinuous when veins extend within the intruded volcanic rocks, as in the Naguayán–El Desesperado district (Boric *et al.* 1990). Cu-bearing veins from the Tocopilla and Guanillos districts contain hypersaline fluid inclusions (48–68 wt% NaCl equivalent) with homogenization temperatures of 320 to 540°C, but mostly between 380 and 420°C. The characteristics of this type of magnetite- and copper-bearing vein indicate that they are high- to moderate-temperature mesothermal deposits (Ruiz *et al.* 1965; Boric *et al.* 1990; Vivallo & Henríquez 1998). They are genetically related to the emplacement and cooling of Upper Jurassic batholiths (K–Ar and ⁴⁰Ar/³⁹Ar ages ranging from 167 to 140 Ma; Maksaev 1990).

The apparent time and space relationships between the Jurassic volcanic-hosted Cu–(Ag) stratabound deposits and the mesothermal Cu-bearing veins hosted by plutonic rocks, plus geochemical and isotopic comparison, led Vivallo & Henríquez

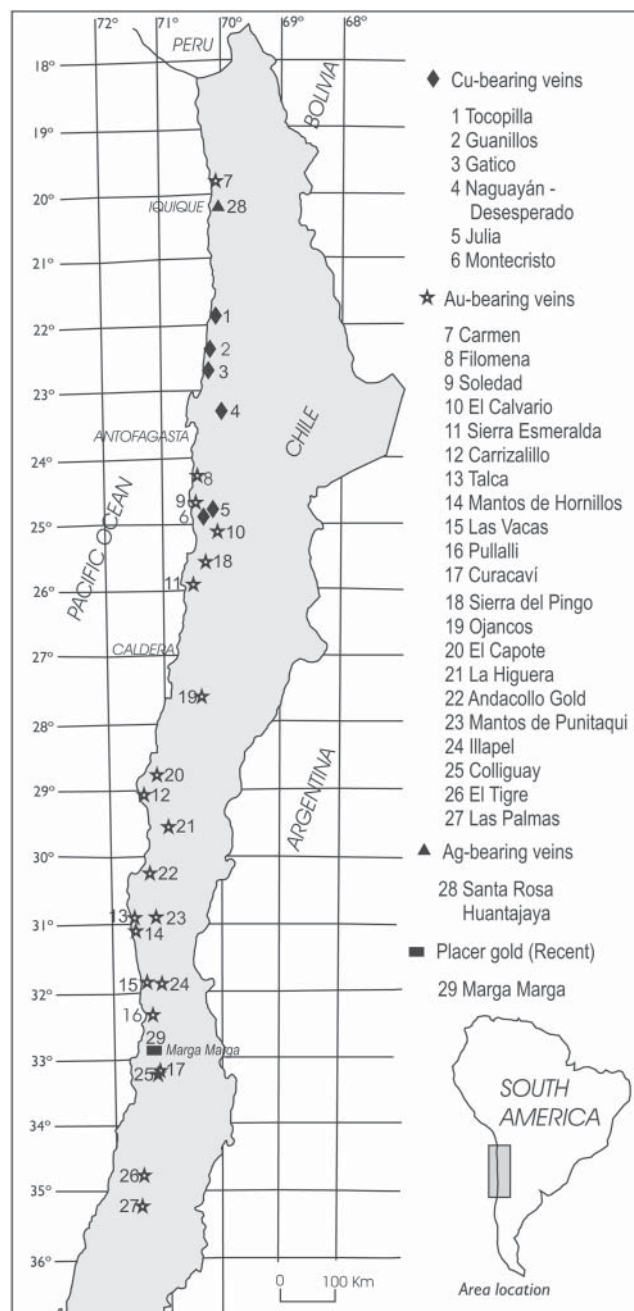


Fig. 6.11. Districts with mesothermal vein deposits in northern Chile.

(1998) to postulate that the batholith-hosted, iron-rich, Cu-bearing veins were the deep feeder structures for the hydrothermal fluids that produced the stratabound deposits within the Jurassic volcanic strata. Although this hypothesis cannot be discarded, these two types of deposits were formed under rather different pressure and temperature conditions, their timing of formation is poorly constrained, and there are no direct field relationships between them.

Mesothermal gold-bearing veins occur in faults and fractures within Jurassic batholiths and their country rocks near the contact with the intrusive masses along the Coastal Cordillera in northern and central Chile (from 20°S to 34°S; e.g. Carmen, Filomena, Carrizalillo, Talca, Hornillos, Las Vacas, Pullalli and Curacaví districts). Individual veins are generally less than 350 m long, 0.5 to 1 m wide, and most have been exploited only within the oxidized zone down to depths of less than 100 m. The

upper exploited section of these deposits includes fine dissemination and stringers of native gold and minor chrysocolla, malachite and atacamite with gangue of quartz, hematite, limonite, calcite and rare magnetite, sericite and tourmaline. The hypogene ore minerals at depth are mostly specularite, auriferous pyrite and variable amounts of chalcopryrite. Although these are minor, largely abandoned gold deposits, they appear to be the primary source of alluvial gold that has been exploited in placer deposits within the Coastal Cordillera of central Chile. A good example is the Marga-Marga placer deposits 120 km NW from Santiago city (Fig. 6.11), where about 2 Moz of alluvial gold have been extracted since colonial times.

Silver-bearing veins occur in the old district of Santa Rosa Huantajaya in the Coastal Cordillera of Iquique (latitude 20°15'S; Fig. 6.11). This district has been exploited intermittently since colonial times, becoming a significant silver producer by the end of the nineteenth century and early years of the twentieth century, but has been abandoned since 1942. The veins are hosted by Middle to Late Jurassic limestone strata that are intruded by dioritic and andesitic dykes and sills with local skarn development. Veins are orientated N60°E/60°S, N60°W/60S and east-west/subvertical, extend up to 700 m long, are 0.5 to 2 m wide, and have rich pockets at vein intersections, which were exploited down to 500 m. The exploited ores came mainly from the oxidation zone, which extends down to 100 m and includes an assemblage of chlorargyrite, malachite, azurite, and limonite, but also from a well-developed supergene enrichment sulphide zone, from 100 to 400 m deep, with native silver, acanthite, chalcocite and covellite. The hypogene ores at depth include arsenopyrite, pyrite, chalcopryrite, enargite, argentiferous tetrahedrite, galena and sphalerite.

Various mesothermal gold deposits are associated with Lower Cretaceous volcanic and intrusive rocks (e.g. Henriquez *et al.* 1991; Fribla 1991; Camus 1990b). These deposits are widespread along the Coastal Cordillera between latitudes 27°S and 36°S (Fig. 6.11) and some districts were once significant gold and copper producers, although all are now inactive.

About 100 NW-trending gold-bearing veins of various sizes, along with stratabound deposits associated with K-feldspar-hematite or chloritic alteration haloes occur at Andacollo Gold (Oyarzún *et al.* 1996). Gold is contained in gold-bearing pyrite, which is disseminated along pink-coloured, pervasively altered strata, and accompanied by sphalerite, galena and chalcopryrite within veins. Cinnabar and tennantite also occur within the most distal veins (3 km away from the Andacollo porphyry copper deposit). A whole-rock K–Ar age of 91 ± 6 Ma obtained from a gold-bearing sample is regarded as a minimum age for the mineralization (Munizaga *et al.* 1985; Reyes 1991).

In the Mantos de Punitaqui district a number of gold-bearing quartz-sulphide veins and carbonate-sulphide breccia pockets occur (Los Mantos, Delirio, Milagro, Azogues, La Culebra and Farellón mines). These deposits are distributed along the Punitaqui NNE-trending sinistral shear zone and en echelon NNW-trending subsidiary faults. The Punitaqui fault has juxtaposed a sequence of Lower Cretaceous andesites and skarnified limestone intercalations with a granodioritic pluton that has yielded a biotite K–Ar age of 114 ± 3 Ma. Hypogene sulphides are pyrite, chalcopryrite, tetrahedrite, bornite, cinnabar, schwartzite and realgar; specular hematite and magnetite are also common. Highest gold and copper grades occur within breccia bodies with a calcite matrix and abundant chalcopryrite and hematite. At Colliguay gold-bearing quartz-pyrite veins (Tirillenta) and gold-bearing silicified breccia bodies (Vizcaino) occur peripheral to Lower Cretaceous granodioritic porphyry pervasively altered to quartz-sericite-chlorite and containing a stockwork of pyrite veinlets (Hernandez *et al.* 1999; Townley *et al.* 2000). A sericite K–Ar age of 129 ± 3 Ma was obtained for gold-bearing breccias in this district.

Gold-bearing veins hosted by Lower Cretaceous plutons occur at El Capote, Ojancos and La Higuera districts. The

hypogene section of the veins consists mainly of quartz, pyrite, chalcopryrite and arsenopyrite, with significant specularite occurring locally. Limited sericitic hydrothermal alteration haloes extend some metres in the host rocks. Similar gold-bearing quartz-sulphide veins, but hosted by Lower Cretaceous volcanic rocks, occur at Las Palmas and El Tigre districts, whereas gold-bearing breccia bodies with quartz-sulphide matrix hosted by a Lower Cretaceous granodiorite occur at El Chivato (Camus 1990b).

Numerous mesothermal to epithermal Au–Cu-bearing vein systems are related to Late Cretaceous igneous rocks. Along the Coastal Cordillera of north-central Chile the districts of Inca de Oro, El Espino, Farellón Sanchez, El Bronce de Petorca, Alhué and Lo Chancón (Fig. 6.11) are all representative of this type of mineralization. Moreover, silver-bearing mesothermal veins that are hosted by Jurassic and Lower Cretaceous limestone strata at Caracoles, Chimberos-Tres Puntas, Chañarcillo and Arqueros districts probably were also formed in the Late Cretaceous. Although these districts have long mining histories, most of the mines became uneconomic and are now abandoned. Mechanized mining activities are at present limited to Alhué and Farellón Sanchez districts in central Chile.

Complex structurally controlled vein systems characterize the Late Cretaceous metallic ore deposits. Individual veins are formed of quartz, sulphides and sulphosalts (pyrite, arsenopyrite, with variable amounts of chalcopryrite, sphalerite, galena and tetrahedrite), and common but minor barite and carbonates. In addition, some veins include significant amounts of magnetite and hematite (Alhué). The veins are hosted by Cretaceous volcanic rocks, but are commonly close to Upper Cretaceous stocks (Inca de Oro, El Espino, Farellón Sanchez, El Bronce de Petorca, Alhué and Lo Chancón districts). Most veins were primarily exploited down to depths of 20–60 m in their richest upper oxidized sections formed by quartz, limonite and hematite with grades of 8 to 30 g/t Au, but quartz-sulphide ores have also been mined from the largest veins and richest ore shoots down to 100 to 200 m depth, and exceptionally to 400 m (El Bronce de Petorca) with grades of 3 to 6 g/t Au; copper has commonly been recovered from these sulphide ores.

K–Ar dates at El Bronce de Petorca bracket the precious metal mineralization between 86 and 79 ± 3 Ma (Camus *et al.* 1991). K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dates from pre- and post-mineralization intrusive bodies at Alhué district encompass the ore deposition roughly between 95 and 83 ± 4 Ma, and adularia $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Maqui vein indicates that mineralization occurred between 86 and 82 Ma (Cotton *et al.* 1999). At Inca de Oro intrusive bodies with K–Ar ages between 80 and 77 Ma have been taken to indicate that precious metal mineralization occurred in Late Cretaceous times (Palacios *et al.* 1993). Although no radiometric data are available for other districts, the Upper Cretaceous stratigraphic age of the host volcanic rocks provides a maximum age for the deposits.

Silver-bearing veins hosted by Jurassic and Lower Cretaceous limestone strata occur in the old Caracoles, Chimberos-Tres Puntas, Chañarcillo and Arqueros deposits. These silver deposits were intensively exploited during the nineteenth century, but are long abandoned. The Chañarcillo district, renowned for its rich bonanzas, is located 50 km south of Copiapó city and includes a large number of subvertical veins with NE, NW and north-south strikes (Whitehead 1919; Segerstrom 1962; Díaz *et al.* 1998).

Sedimentary rock-hosted gold deposits

A Jurassic limestone sequence cropping out east of the old Potrerillos copper mine hosts disseminated gold deposits such as El Hueso, Agua de la Falda and Jerónimo. El Hueso is a sedimentary rock-hosted epithermal gold deposit located 3 km east of the Potrerillos porphyry copper. The original mineral

reserves at El Hueso were 12 Mt at 1.5 g/t Au (0.4 g/t Au cut-off) that were mined out by Homestake Company from 1988 to 1994. El Hueso occurs within a 150-m-thick section of Jurassic limestone strata intruded by stocks and dykes of dacitic porphyry. The sedimentary rocks at the mine pit are pervasively silicified, and show siliceous–argillic alteration. Intermediate argillic alteration encompasses the mineralized rocks. Disseminated gold occurs along some stratigraphic levels forming an overall stratabound orebody (0.4 to 2.0 g/t Au) striking north–south and dipping 30°W. In addition, the stratabound gold dissemination is intersected by a number of mineralized east–west to N70°W trending faults that dip 70°N. Irregular, massive siliceous ledges occur along these faults (Esperanza, Hueso, Tunnel 3 and Central veins). The highest gold grades occur within these siliceous veins assaying locally up to 30 g/t Au, but averaging 4 g/t Au. Ore minerals consist of native gold, electrum, rare nagyagite, scorodite, cervantite, stibnite, enargite, pyrite, cinnabar, realgar, orpiment, arsenopyrite, galena, chenevixite, chalcantite, malachite and chrysocolla; gangue minerals are quartz, alunite, goethite, jarosite and minor barite. Gold occurs as 3 to 25 µm sized grains as native metal and electrum. The exploited oxidation zone extends down to 150 m depth, with the underlying hypogene ores being dominated by pyrite with low gold content (Illanes 1991).

El Hueso was customarily regarded as a high-sulphidation epithermal gold deposit with ‘Carlin-like’ affinities and a genetic link to the Potrerillos porphyry copper deposit (Colley *et al.* 1989; Sillitoe 1991; Davidson & Mpodozis 1991). However the $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological data and field observations of Marsh *et al.* (1997) indicate that El Hueso was formed from 40.2 to 40.8 Ma as a low-sulphidation epithermal gold deposit pre-dating by some 5 Ma the nearby Potrerillos porphyry copper deposit. An advanced argillic alteration assemblage present at Cerro Silica at El Hueso, consisting of alunite, pyrophyllite, diaspore, zunyite, dickite, rutile and pyrite, was dated at 36.2 Ma (alunite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau). According to Marsh *et al.* (1997) this is a late event lacking significant gold mineralization and also pre-dating the Potrerillos porphyry copper.

The Agua de la Falda mine located 2 km east of El Hueso is another oxidized sedimentary-hosted gold deposit that was exploited by Homestake mining company. Silicified and Fe–Mn-stained Jurassic limestone beds host gold dissemination. The oxidized mineral resources at Agua de la Falda amount to 1.3 Mt at 7.4 g/t Au (2.0 g/t Au cut-off). Down-dip in the same stratigraphic level the Jerónimo sulphide orebody occurs, though both deposits are separated by a post-mineralization normal fault. Homestake has delineated a resource of 15.05 Mt at 5.79 g/t Au at Jerónimo (Homestake, unpublished data). Jerónimo is a stratabound disseminated gold deposit within a section of Jurassic limestone strata of El Asiento Formation located 5 km east of Potrerillos porphyry copper deposit. The mineralized level (N20°E/12–15°W) is about 12 m thick and extends at least 200 m down-dip. Mineralization is also structurally controlled as ore minerals are focused in subvertical fractures and joints.

The ore mineral assemblage at Jerónimo consists of pyrite, arsenopyrite, sphalerite, lead sulphosalts, orpiment and realgar, with minor coloradoite, altaite, cinnabar and cassiterite. Although minor visible gold is present, it generally occurs as 5 to 30 µm grains that are encapsulated in pyrite, arsenopyrite, quartz and realgar, and gold also occurs within vugs in the silicified matrix of host rocks. Alteration phases include: (1) strong, pervasive replacement-style silicification; (2) vug-filling by manganese carbonate and calcium carbonate (rhodocrosite, kutnohorite); and (3) argillization consisting of widespread disseminated and veinlet illite, and vug-filling kaolinite in the centre of the deposit. Other common alteration minerals include apatite, rutile, monazite and barite. Pb and O isotopic data indicate ore fluid derivation mostly from a magmatic

Table 6.1. Tonnage and Au grade of the main Au porphyry deposits of the Maricunga area, northern Chile (data from Muntean & Einaudi 2000)

Deposit	Ore (Mt)	Au grade (g/t)	Cut-off grade (g/t Au)	Au (Moz)	Cu grade (%)
Marte	46	1.44	0.5	2.1	0.05
Lobo	53	1.73	1.0	2.9	0.12
Refugio (Pancho)	216	0.88	0.5	6.1	0.05–0.2
Verde (Refugio)	81	0.85	0.5	2.2	0.03
Cerro Casale	1183	0.71	0.5	26.9	0.26

source with some input from wall-rock sources (Gale 2000). Some affinity to Carlin-type sedimentary rock-hosted gold deposits has been suggested for Jerónimo (e.g. Sillitoe 1991). However, according to Gale (2000) Jerónimo is a gold-rich carbonate replacement deposit that shows distinct differences from Carlin-type gold deposits. These differences include enrichment in base metals, manganese carbonate alteration, the presence of gold as visible grains, occurrence in a district containing porphyry and related styles of mineralization, and evidence for a magmatic contribution to metals and hydrothermal fluids. Although no geochronological data are available from Agua de la Falda and Jerónimo gold-bearing orebodies, a number of upper Eocene stocks intrude the Jurassic limestone sequence in the region (e.g. Marsh *et al.* 1997; Mpodozis *et al.* 1994) suggesting that the hypogene mineralization at these deposits was formed in late Eocene times, probably close in age to the nearby El Hueso epithermal gold deposit.

Gold-rich porphyry deposits

A number of gold-rich porphyry deposits occur in the High Andes between latitudes 27°S and 28°S (Maricunga belt; Muntean & Einaudi 2001) and include Refugio, Cerro Casale, Marte, Lobo, Santa Cecilia, and other prospects (Fig. 6.7). This group of deposits probably forms the largest gold concentration in the Andes as their resources exceed 40 Moz of gold, but they are contained within low-grade deposits (Table 6.1). Only Refugio was recently under exploitation, but closed mineral extraction in May 2001 due to a low gold price, although it is expected to start up again in May 2005, with expanded crushing facilities. Marte was also exploited from 1989 to 1991, but the mine closed due to deficient gold recovery from heap leaching. The rest of these Au porphyry deposits are essentially low-grade, large-tonnage prospects (Table 6.1), except for Cerro Casale, which contains a core of hydrothermal breccia with high Au and Cu grades (reverse circulation hole 176 intersected 42 m of breccia at 17.7 g/t Au and 1.5% Cu; Bema Gold Corporation, unpublished data, 1999).

Miocene volcanic edifices of the southernmost Central Volcanic Zone of the Andes (Mpodozis *et al.* 1995) host the Au porphyry deposits of the Maricunga area. An example is the Marte deposit which occurs in the core of Pastillitos volcano and is a parasitic cone on the eastern foothills of the large compound Copiapó stratovolcano. Geochronological data indicate that a group of deposits was formed from 24 to 22 Ma (Refugio, Santa Cecilia, Cavanacha at La Pepa) and another group was formed from 14 to 13 Ma (Cerro Casale, Marte, Lobo) (Sillitoe *et al.* 1991; McKee *et al.* 1994; Mpodozis *et al.* 1995). The Miocene gold-rich porphyries occur within a section of the Chilean Andes with a NNE structural trend (deflection) between latitudes 27°S and 29°S and dominated by major reverse faults. Despite this structural disposition, the porphyry stocks, intrusive breccia bodies, gold-bearing sheeted veins,

local faults, dykes and silica ledges related to these deposits are preferably aligned along NW trends. Thus, on a local scale the Au porphyry mineralization appears to be controlled by subsidiary NW-trending faults that are especially common in the Maricunga area. These NW faults accommodate structural shortening and have sinistral strike-slip displacements at 27°S latitude (Tomlinson *et al.* 1994; Mpodozis *et al.* 1994). A similar structural setting is shared by the older Lower Oligocene gold-rich porphyry copper deposits at the La Fortuna cluster located some 50 km SSW from the Maricunga Au porphyries. Whether this specific structural setting was a controlling factor for Au-rich porphyry mineralization or just the effect of large-scale tectonic processes unrelated to mineralization cannot be resolved at this time, although gold porphyries are notably absent outside of the mentioned NNE-trending deflection of the Chilean Andes.

The Au porphyry systems of the Maricunga area are related to dioritic-quartz dioritic multiphase stocks intruded into Miocene andesitic to dacitic volcanic strata. The systems tend to be vertically elongated with a core of potassic alteration characterized by widespread biotite as dissemination, veinlets and fine aggregates forming amphibole pseudomorphs. K-feldspar occurs in veinlets and veinlet haloes, and both biotite and K-feldspar are accompanied by profuse magnetite and hematite that forms up to 5 to 10% of the mineralized zones. A marginal propylitic zone occurs, mostly developed within the host volcanic rocks. The upper sections of the Au porphyries include irregular zones with assemblages of quartz-sericite-chlorite-clays or pyrite-albite-clays that may be interspersed with patches of propylitic or potassic alteration. The upper section of some deposits (Marte, La Pepa and Santa Cecilia) includes advanced argillic alteration overprinting. This juxtaposition of early deep porphyry-type hydrothermal and late shallow epithermal alteration appears to be the result of active erosion and collapse of the volcanoes that host the porphyries during active hydrothermal circulation (Sillitoe 1994). Gold-rich porphyry orebodies are characterized by the occurrence of multiphase quartz stockworks, which are either multidirectional or frequently occur as closely spaced subparallel sets or sheeted veins with a preferred NW strike. Sulphides are dominated by pyrite, with minor chalcocopyrite, and traces of bornite and molybdenite. Copper content is variable, from 0.03% Cu in the Verde orebody at Refugio, to 0.26% Cu at Cerro Casale (Table 6.1). Volcanic reconstruction and fluid inclusion data indicated to Vila & Sillitoe (1991) that the gold-bearing stockwork and sheeted-vein ores were generated some 600 to 1000 m beneath palaeo-surfaces. Further fluid inclusion studies of Muntean (1998) revealed that early quartz-magnetite-biotite-chalcocopyrite veinlets were formed at temperatures >500°C and at pressures estimated as between 200 and 400 bars, suggesting depths of 800 to 1600 m under lithostatic loads. In contrast, late banded quartz veinlets indicate temperatures <350°C and estimated pressures <200 bars, suggesting depths of 200 to 1500 m under hydrostatic conditions. Gold crystal composition at Cerro Casale indicates that gold was introduced both during the early potassic alteration and a late white-mica-dominated alteration (Palacios *et al.* 2001). Hydrothermal breccias are common within the deposits, but the only significant richly mineralized breccia occurs underground at Cerro Casale. It is a funnel-shaped fragment-supported breccia body with a vertical extent of 300 m and an elliptical plan section up to 100 × 150 m in size. The breccia comprises subrounded intrusive fragments 5–10 mm in diameter within a matrix of anhydrite, barite, tourmaline, rhodochrosite, dolomite, chalcocopyrite, pyrite, galena, sphalerite, quartz and sericite.

Depths of supergene oxidation in the Maricunga Au-rich porphyries range from a few metres to a hundred metres with Cu leaching in the zone of oxidation, but constant Au. Locally, supergene enrichment blankets occur, e.g. Bema Gold reported a 5–35-m-thick chalcocite blanket at Cerro Casale averaging

1–1.3% Cu (*Northern Miner*, 22 April 1996), whereas at Cavancha porphyry a supergene chalcocite blanket ranging from 15 to 70 m in thickness and containing 0.1 to 0.5% Cu occurs (Muntean 1998). However, supergene copper-enriched zones are lacking in the other gold-rich porphyries.

Transitions between gold-bearing porphyry-type stockworks and shallower zones of advanced argillic alteration containing precious metal mineralization occur in Maricunga (Vila & Sillitoe 1991). Silica ledges that are replacement veins of quartz-alunite alteration with local core zones of vuggy silica host high-sulphidation epithermal mineralization. The most obvious example occurs at La Pepa where gold-bearing bonanza-type veins that strike N10°–40°W overlie and are peripheral to the low-grade Cavancha porphyry deposit (Muntean 1998). The advanced argillic alteration partly overprints the Cavancha porphyry, and geochronologic data of Muntean (1998) indicate that there is a 140 000 to 900 000 year gap ($^{40}\text{Ar}/^{39}\text{Ar}$ at 95% confidence level) between the Au-rich Cavancha porphyry alteration and later advanced argillic overprint and precious-metal-bearing vein formation at La Pepa. Another example is Cerro Casale where a N50°–80°W striking vein system containing Pb, Zn, Ag, Sb and minor Au occurs immediately west of the porphyry (Vila & Sillitoe 1991). $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological data from alunite in silica ledges at the Vein Zone, as well as biotite from the Cerro Casale porphyry, indicate that both were formed synchronously at 13.9 Ma (Muntean 1998).

Skarn deposits

At least 13 skarn deposits have been documented in Chile, but only a few are currently being exploited (e.g. Cabildo and El Toqui districts). Cu skarns are by far the dominant type, but local examples of Zn–Pb and Fe skarns also occur (Fig. 6.12). The most prominent are Cu skarn deposits that occur along the Coastal Cordillera of northern Chile hosted by skarnified units of Lower Cretaceous carbonate rocks intercalated within a volcanosedimentary sequence. A few examples also occur in the High Andes and are again hosted by skarnified Lower Cretaceous carbonate rocks, which were deposited in a shallow backarc marine basin. In the Patagonia area of southern Chile the El Toqui Zn–Pb skarn is the only Chilean producer of Zn, and is also hosted by a metamorphosed Lower Cretaceous marine sedimentary sequence. One unusual skarn with lazurite-rich pockets, within an outer marble facies, occurs in the High Andes of northern Chile. It provides lapis lazuli widely used as a gemstone in the country. This gemstone was extracted by native Andean inhabitants long before Spanish colonization (Rivano & Sepúlveda 1991).

The Cabildo copper district (Fig. 6.12) is one of the largest Cu skarn mineralization systems in the Coastal Cordillera, and the only one currently exploited by underground mining. The copper deposits are hosted by sedimentary intercalations within the mostly volcanic Lower Cretaceous Lo Prado Formation, immediately west of a batholith composed of diorite, tonalite and granodiorite, with K–Ar dates ranging from 106 ± 3 to 96 ± 3 Ma (Rivano *et al.* 1993). The limestone, sandstone, marl, tuff and conglomerate beds form a 400–500-m-thick intercalation within Lower Cretaceous volcanic rocks. The calcareous units are altered to skarn in the central and northern section of the district and comprise garnet (grossularite–andradite), clinopyroxene (diopside–hedenbergite), amphibole (tremolite–actinolite), epidote, calcite, scapolite, quartz, chlorite and titanite. Clastic and pyroclastic units are normally transformed to siliceous hornfels. North–south trending normal faults dipping 60–70° west displace the skarn units (with a sinistral lateral component) and some fault zones are mineralized. A number of stratiform, irregular and vein-type copper-bearing orebodies are hosted by the skarnified units. Stratiform orebodies are 2

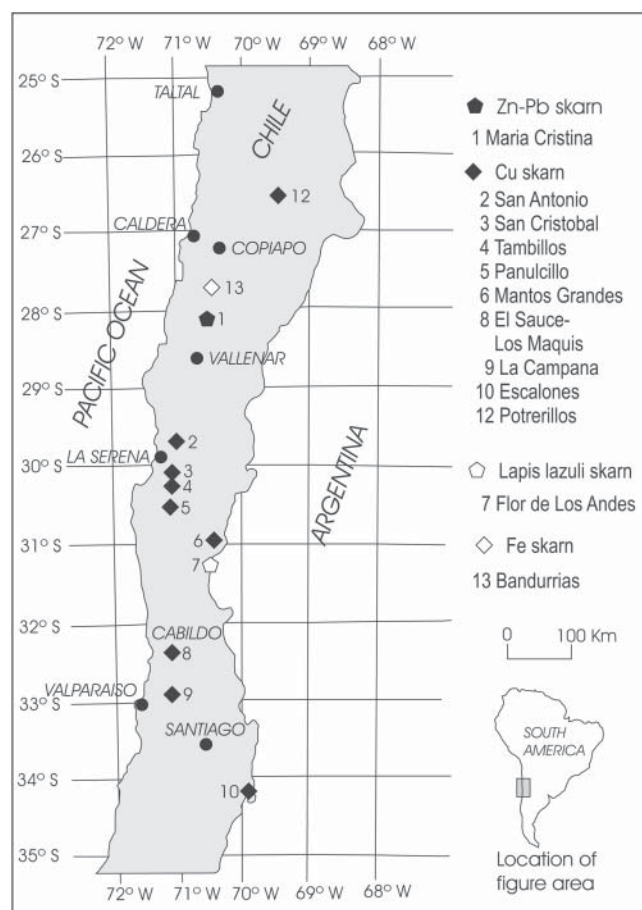


Fig. 6.12. Skarn deposits of northern Chile.

to 15 m thick, but sections up to 70 m thick have been exploited in places where mineralized horizons occur close together, the richest pockets being found adjacent to mineralized fault zones. The dominant hypogene ore mineral of the skarn deposits is chalcopyrite with minor pyrrhotite and pyrite, and some supergene chalcocite. However, copper mineralization extends outside of the skarnified areas, where chalcopyrite, bornite and chalcocite constitute the ore minerals.

The San Antonio district is located 24 km NE of La Serena (Fig. 6.12), and has a historic production of 200 000 t at 1.5–2.2% Cu and 1–2 g/t Au. Carbonate units intercalated in the Lower Cretaceous, mostly volcanic Arqueros Formation have been skarnified within a 300–600-m-wide metamorphic contact aureole of the Santa Gracia granite (K–Ar date, 98 and 89 ± 0.6 Ma; Moscoso *et al.* 1982a). A number of stratabound and stratiform orebodies occur in the district: these are 4 to 6 m thick, extend for 75 to 120 m along-strike, and 20 to 90 m along-dip. The host rocks are altered to a skarn assemblage of garnet, clinopyroxene, scapolite, epidote, amphiboles, calcite, biotite, phlogopite, K-feldspar, quartz, chlorite, illite, prehnite and laumontite. The ore minerals are magnetite, pyrite, hematite, chalcopyrite, minor galena, sphalerite, arsenopyrite, pyrrhotite and bornite (Ardila 1993; Ardila *et al.* 1994).

The Panulcillo mine is located 50 km south of La Serena ($30^{\circ}27.3'S$, $71^{\circ}13.1'W$; Fig. 6.12) and is famous for producing high-grade chalcopyrite ores: the mine has produced 500 000 t at grades of 2.75–3.50% Cu and 0.5–1.0 g/t Au (Sugaki *et al.* 2000). The skarn occurs within Valanginian carbonate units striking north–south, and dipping 60° to 70° east, and comprises gaudite garnet, phlogopite, subordinate clinopyroxene, scapolite, pargasite, calcite, phlogopite, plagioclase, epidote, clinozoisite, chlorite, illite, K-feldspar and axinite. Irregular

orebodies occur within the garnet zone; these are 20 to 50 m thick, extending up to 160 m along-strike, and 60 m along-dip. Chalcopyrite is the main ore mineral, but pyrrhotite, pyrite, sphalerite, arsenopyrite, galena, magnetite, hematite and minor molybdenite occur in the interstices of skarn minerals. Ore deposition appears to have taken place at 115 ± 3 Ma (K–Ar in phlogopite; Ardila 1993), although Sugaki *et al.* (2000) reported a K–Ar date of 132 ± 7 Ma for a dioritic stock located at the east of the Panulcillo mine. Faults and fractures filled with hypogene minerals in the Panulcillo deposit are compatible with a north–south sinistral shear zone active at the time of mineralization and focusing fluid flow and ore deposition (Ardila 1993).

The La Campana district is located in the Coastal Cordillera 70 km north of Santiago (Fig. 6.12) in the southern and southwestern slopes of the Campana hills. The Opositora, Veta Grande, Felicidad, Pronosticada, Guanaco and Mina de Hierro deposits occur within the metamorphic contact aureole of the Caleu pluton. Stratiform Cu and Fe deposits are hosted by skarnified limestone intercalations of the Lower Cretaceous Lo Prado Formation near the contact of a gabbro that is a western facies of the Caleu Pluton, located within La Campana National Park. The Opositora mine indicated resources of 68 500 t at 1.5% total Cu, 25 g/t Ag and 8 g/t Au, which increase to 197 000 t if inferred resources are included. The Caleu pluton shows variations in lithology, mineralogy, texture and chemical composition that allow identification of three north–south trending plutonic units: Gabbro/Diorite, Tonalite and Granodiorite zones. Its age according to $^{40}\text{Ar}/^{39}\text{Ar}$ dating ranges from 117 to 94 Ma (Parada *et al.* 2002), but the La Campana gabbro stock has a plagioclase $^{40}\text{Ar}/^{39}\text{Ar}$ date of 130 ± 1.5 Ma (Parada & Larrondo 1999). The skarn units are composed of garnet (grossularite–andradite), pyroxene (diopside–salite), calcite and quartz. The orebodies are up to 20 m thick, extending up to 200 m along the NNW strike of the host rocks, and 100 m along the dip. They comprise disseminated chalcopyrite, pyrite, pyrrhotite, bornite, molybdenite, sphalerite, magnetite and minor galena arsenopyrite, marcasite and hematite. Gold and silver are associated with the sulphides.

The Maria Cristina Zn–Pb skarn occurs in Quebrada Galena, about 80 km south of Copiapó, and is hosted by carbonate rocks of the Lower Cretaceous Chañarcillo Group. The alteration of the carbonate host rocks at Maria Cristina involved recrystallization of the limestone to marble and the formation of coarse crystalline skarn lenses. The prograde assemblage comprises garnet (andradite $\text{Ad}_{30}\text{Gr}_{70}$), diopside pyroxene, magnetite and epidote. The occurrence of former pyrrhotite is evidenced by its pseudomorphic replacement by marcasite, pyrite and calcite. The mineralization consists of coarse-grained pyrite-rich massive sulphides within the carbonate rocks, and massive and semi-massive sulphides in retrograde skarn and within the matrix of coarse-grained volcanoclastic rocks. It occurs at the contact of potassically altered diorite porphyry of mid-Cretaceous age (93.6 ± 0.4 Ma, amphibole $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date; Lieben *et al.* 2000). The ore minerals comprise sphalerite, pyrite and galena, minor marcasite and magnetite, and trace amounts of chalcopyrite and tetrahedrite. The deposit occurs at the intersection of two NE and NNE trending faults, which controlled the mineralization within folded carbonate rocks (Lieben *et al.* 2000).

A stratiform iron oxide deposit occurs at Cerro Bandurrias, 55 km south of Copiapó (Fig. 6.12), hosted by the Lower Cretaceous Chañarcillo Group. The ore horizon is 7 m thick, extending 1.7 km across the western slope of Bandurrias hill. Dioritic intrusives crop out in the area, and a skarn assemblage (garnet, scapolite) associated with these intrusions occurs at the same stratigraphic level that hosts the iron oxide horizon. The ore is formed of idio- to hypidiomorphic magnetite crystals, partially martitized, with local garnet bands. Although the setting of this iron deposit is strongly suggestive of skarn affiliation (e.g.

Oyarzún 2000), it was formerly interpreted to be of primary volcanogenic exhalative origin with superposed contact metamorphism (Cisternas 1990; Espinoza 1990). Magnetite is again the dominant ore mineral at the San Cristobal skarn deposit in the Coastal Cordillera south of La Serena, although the skarn has actually been exploited for copper. Iron oxide is also abundant in the copper skarns of the San Antonio district 24 km NE of La Serena.

The Zn–Pb (Au) deposits of El Toqui are located in Chilean Patagonia about 1700 km south of Santiago, and provide the only source of zinc production in Chile. The district covers about 25 km² and contains stratiform (Concordia, San Antonio, Mallín–Mónica, Doña Rosa–Mallín Sur, Aserradero and Estatuas orebodies) and discordant vein deposits (Antolín and Zúñiga orebodies). It is hosted by skarnified Lower Cretaceous marine sedimentary rocks with volcanic intercalations, which are intruded by a mid-Cretaceous sill of quartz feldspar porphyry altered to an assemblage of quartz, albite, K-feldspar and sericite. Whole-rock K–Ar and ⁴⁰Ar/³⁹Ar dates of the porphyry range from 100 ± 2 to 108 ± 4 Ma (Townley & Godwin 2001). North of El Toqui River, a tonalite stock also occurs. The Cretaceous succession strikes north–south to NNW–SSE, dips ENE from 5° to 25°, and shows propylitic alteration. A number of carbonate units, varying from 0.5 to 10 m in thickness, are skarnified to garnet, pyroxene, amphibole, epidote, chlorite and ilvaite. Ore minerals include marmatite (iron-rich sphalerite), galena, pyrrhotite, pyrite, arsenopyrite, tetrahedrite, chalcopryite, electrum, native gold, magnetite, silver sulphosalts, native bismuth, hessite and maldonite. Veins up to 6 m thick and 60 m in length are exposed in the district and comprise a stockwork of veinlets of sphalerite, galena, pyrite, chalcopryite, quartz and calcite. The mineralization was formerly interpreted as exhalative volcanogenic (Wellmer *et al.* 1983; Wellmer & Reeve 1990), but subsequent studies have shown its skarn affiliation (Palacios *et al.* 1994), and the skarn model has successfully been applied in exploration in the district (Kakariekka 2002). Fluid inclusion studies have shown that the deposits were formed by two stages of hot fluid circulation: an early stage of saline fluids (11% NaCl equivalent) under temperatures of 350 to 300°C, and a later stage of more dilute fluids (4–8% NaCl equivalent) at 200 to 160°C (Palacios *et al.* 1994).

The only example of skarn development related to Chilean porphyry copper deposits occurs at Potrerillos (26°30'S, 69°22.9'W; Camus 1985) where the mineralized Cobre porphyry stock dated at 35.87 ± 0.21 Ma (⁴⁰Ar/³⁹Ar in hornblende; Marsh *et al.* 1997) intrudes a Jurassic–Lower Cretaceous carbonate sedimentary sequence. Surrounding the porphyry stock is a 150-m-wide halo, in which metamorphic and metasomatic assemblages and minor skarn are developed. Within this aureole occurs an assemblage of garnet, wollastonite, enstatite, diopside, epidote, zoisite, clinozoisite and tremolite, with pyrite and specular hematite (Marsh 1935).

The Mina Lar, a small (c. 200 000 t) oxidized Cu–Au skarn deposit, occurs 20 km south of Copiapó in the Coastal Cordillera of Atacama region, northern Chile. Phelps Dodge Mining Company explored this deposit, which is hosted by skarnified Lower Cretaceous limestone strata of the Chañarcillo Group, and discovered in 1986 the underlying large, volcanic-hosted Candelaria iron oxide copper–gold deposit (Ryan *et al.* 1995; Marschik 2001).

At the Mantos Grandes Cu–Au mine (30°51.4'S, 70°33.2'W) a sequence of Lower Cretaceous marine carbonate rocks crops out. These are folded, repeated by a reverse fault (Tulahuencito fault), and intruded by diorite to gabbro stocks ascribed to the Tertiary (Mpodozis & Cornejo 1988). In a north–south elongated zone of 500 × 250 m the carbonate rocks are altered to a skarn of andradite–hematite, wollastonite-bearing marble, and silicified rocks. A number of en echelon subhorizontal orebodies (N20°E/30°W) occur within the skarn, controlled by tensional subsidiary fractures related to the Tulahuencito

reverse fault. Individual orebodies are from 1.8 to 3 m thick, extend 100 m along-strike and 20 m down-dip. The mineralized rock has abundant andradite crystals 2 to 4 mm in diameter that form aggregates within calcite, and specular hematite forming discoidal crystals up to 5 cm in diameter. The main ore mineral is partly oxidized coarse-grained chalcopryite.

The Escalones deposit is located in the High Andes of central Chile, east of Santiago (34°07.2'S, 69°57.9'W). The copper deposit is hosted by a Lower Cretaceous sequence of marine carbonate rocks and gypsum of the Lo Valdes Formation (striking NW and dipping 60° NE) that is intruded by Neogene andesite porphyry and north–south andesitic, vertical dykes 2–3 m thick. Skarn alteration occurs within an area of 800 × 100 m, with garnet, actinolite, tourmaline, quartz, calcite, zoisite, quartz, chlorite and minor sericite. Ore minerals are chalcopryite, bornite, magnetite, hematite, pyrite, pyrrhotite and galena, and hypogene grade is 1.71% Cu (Flores 1943). Drill and channel sampling by General Minerals Corporation reported high-grade copper–gold skarn mineralization at Escalones Alto, including an 81-m channel sampled in a road cut in the 'core area' that yielded 1.54% Cu, 0.74 g/t Au and 9.0 g/t Ag (GNM 2000), but no further work was subsequently undertaken on the deposit.

Metallogenic evolution

Hydrothermal processes related to suprasubduction igneous activity (mostly felsic to intermediate plutonism) are the source of the metallic wealth of Chile. These hydrothermal ore deposits are genetically associated with a succession of north–south trending magmatic arcs that developed since Jurassic times over a poorly mineralized basement that corresponds to an accretionary prism of Late Paleozoic to Triassic age (Mpodozis & Ramos 1989). The arc-parallel ore deposit belts young progressively from Jurassic in the west to late Miocene–Pliocene in the east.

Jurassic to Early Cretaceous Andean evolution took place under an overall extensional tectonic setting characterized by a subsiding volcanic arc with structurally controlled emplacement of shallow batholiths within the nearly coeval volcanic succession. Two other important components of the tectonic setting were the development of the intra-arc, sinistral, strike-slip Atacama Fault System along the present Coastal Cordillera, and the formation of backarc marine carbonate sedimentary basins. This early period of the evolution of the Chilean Andes was strongly dominated by copper mineralization characterized by a number of Late Jurassic and Early Cretaceous volcanic-hosted Cu(Ag) stratabound deposits and many Cu-bearing vein deposits, as well as Early Cretaceous iron oxide copper–gold deposits. In addition, Fe-oxide apatite deposits were formed during the Early Cretaceous within the domain of the Atacama Fault System, related to contact zones of Neocomian volcanic rocks with Early Cretaceous intrusions. Porphyry copper deposits were unimportant during this initial stage, except for some low hypogene grade porphyry Cu–Au deposits that were formed in the latest Early Cretaceous and related to shoshonitic porphyry intrusions (e.g. Andacollo; Reyes 1991).

A compressive tectonic setting, with basin inversion and episodic uplift, has characterized the evolution of the Chilean Andes since Late Cretaceous times. The magmatic arc migrated stepwise east (inland), and the new geological setting along the active continental margin led to a strong dominance of calcalkaline porphyry Cu–Mo deposits which continued during Tertiary times. A belt of Palaeocene–early Eocene porphyry Cu–Mo deposits extends from southern Peru down to 29°30'S latitude in northern Chile, totalling 12.7 Mt of contained copper in Chile (resources plus production; Camus 2003). However, only those porphyries with a well-developed supergene

enrichment blanket are currently economic, such as the Cerro Colorado and Spence deposits. In addition, a number of precious metal epithermal deposits, also related to volcanic rocks, occur along the Palaeocene–early Eocene outcrop in northernmost Chile, which have been preserved due to the limited extent of denudation in the Atacama Desert.

A major metallogenic episode took place during late Eocene–early Oligocene times when enormous porphyry Cu–Mo deposits were formed along the Domeyko Cordillera of northern Chile. These constitute the largest copper concentration in the world, totalling about 220.5 Mt of copper (resources plus production; Camus 2002). The mineralizing period extended from 43 to 31 Ma, though about 60% of the copper resources of the belt were accumulated from 34 to 31 Ma (Camus 2002), and hydrothermal ore precipitation was genetically associated with the closing igneous activity along this range, prior to a 30-km eastward arc migration in response to plate tectonic interaction. The late Eocene–early Oligocene porphyry Cu–Mo deposits are also spatially associated with a major intra-arc strike-slip shear system, the Domeyko Fault System, that developed during a period of NE-directed oblique convergence, transpression, crustal thickening and denudation of the Domeyko Cordillera (Maksaev & Zentilli 1999).

The main precious metal epithermal deposits of Chile are related to later Miocene volcanism that developed in the easternmost section of the Chilean Andes (extending to the east into Argentinean and Bolivian territories). Most deposits are of the high-sulphidation type with structural control of the orebodies, and occur within extensive zones of altered, bleached volcanic rocks. Miocene porphyry Au mineralization occurs in close spatial and temporal association with high-sulphidation

epithermal deposits in the Maricunga area (27–28°S), particularly in the core of eroded volcanic edifices (Vila & Sillitoe 1991).

The last major metallogenic episode of the Chilean Andes was the formation of enormous porphyry Cu–Mo deposits in the Andes of Central Chile (i.e. Los Pelambres, Río Blanco–Los Bronces and El Teniente) which taken together have an estimated yield of 183 Mt of copper (resources plus production; Camus 2002). This youngest porphyry Cu–Mo mineralization followed stages of compressive deformation, crustal thickening and denudation during Miocene and early Pliocene times. Hydrothermal ore precipitation was once again genetically associated with the cessation of igneous activity along the western slope of the main Andes, prior to a 50-km eastward arc migration.

The economic viability of many Cu and Ag deposits in arid northern Chile (north of *c.* 30°S), especially porphyry copper deposits, is dependent on the size and quality of their supergene enrichment blanket. Supergene enrichment processes have been active there, episodically or continuously, from Eocene to Miocene times (Hartley & Rice 2005), a significant supergene metallogenic epoch in Chile.

The strong dominance and recurrence of copper deposits in the Chilean Andes probably results from a relatively homogeneous source of the metal, and the existence of a long-lived subduction-related system of magma generation. However, the distinct and relatively short duration of the major mineralization episodes within this subduction setting implies that a number of geological conditions have to combine to create conditions suitable for the generation of such anomalous metal concentrations along the Andean Cordillera.