ABSTRACT

Major Miocene central Andean (lat 22°–34°S) ore districts share common tectonic and magmatic features that point to a model for their formation over a shallowing subduction zone or during the initial steepening of a formerly flat subduction zone. A key ingredient for magmatism and ore formation is release of fluids linked to hydration of the mantle and lower crust above a progressively shallower and cooler subducting oceanic slab. Another is stress from South American–Nazca plate convergence that results in crustal thickening and shortening in association with magma accumulation in the crust. Fluids for mineralization are released as the crust thickens, and hydrous, lower crustal, amphibole-bearing mineral assemblages that were stable during earlier stages of crustal thickening break down to dryer, more garnet-bearing ones. Evidence for this process comes from trace-element signatures of pre- to postmineralization magmas that show a progression from equilibration with intermediate pressure amphibole-bearing residual mineral assemblages to higher pressure garnet-bearing ones. Mineralization over the shallowing subduction zone in central Chile (28°–33°S) is followed by cessation of arc volcanism or migration of the arc front away from the trench. Mineralization in the central Altiplano-Puna region (21°–24°S) formed above a formerly flat subduction zone as volcanism was reinitiating. Thus, hydration and crustal thickening associated with transitions in and out of flat-slab subduction conditions are fundamental controls on formation of these major ore deposits.

INTRODUCTION

Some of the world's richest and largest copper and gold deposits are associated with Miocene magmatism in the central Andes. This paper reviews how the formation of major ore deposits between 22° and 34°S can be linked to the late Cenozoic magmatic and tectonic response of the mantle and lower crust to the formation and subsequent steepening of shallow subduction zones (Figs. 1 and 2). Mineral districts discussed are the El

Figure 1: Central Andean map showing major Miocene mineralized areas (white boxes, yellow labels) relative to: (a) depth contours in km to Wadati-Benioff seismic zone of subducting Nazca plate (from Cahill and Isacks, 1992), (b) southern (SVZ) and central (CVZ) volcanic zones and Chilean and Peruvian flat-slab regions, (c) regions >3000 m in elevation (in red), and (d) foreland fold-thrust belts: Precordillera and Subandean–Eastern Cordillera thin-skinned belts (green), Santa Bárbara thick-skinned belt (black), and Pampean block uplifts (gray).
TECTONIC SETTING OF MIocene Laramide deposits in Western North America

...subduction zone in Peru and for Tertiary for Miocene deposits over the shallow subduction zone. The model has implications above a shallowing or recently shallow transforms into dryer, garnet-bearing crust amphibole-bearing lower crust thickens and subducting slab are released as wet mineralization that are ultimately derived from the hydrated mantle above the subducting Nazca plate. Fluids for thickness induced by the evolving geometry to changes in crustal and lithospheric districts. In the model, mineralization is linked for Miocene deposits over the shallow subduction zone in Peru and for Tertiary Laramide deposits in Western North America.

TECTONIC SETTING OF MIocene CENTRAL ANDean ORE DEPOSITS

Major Miocene central Andean ore districts are located in extinct Miocene volcanic belts underlain by thickened continental crust (50–70 km thick; Isacks, 1988) on the arc side of major fold-thrust belts. This paper explores why they occur where they do. Figure 1 shows mineral districts between 22° and 34°S relative to modern central Andean geologic provinces and contours to the Wadati-Benioff seismic zone. The most prominent province, the Puna-Altiplano Plateau with its widespread Miocene to Recent volcanic cover and average elevation of 3700 m, is second on Earth only to the Tibetan Plateau in area and height. To the north and south, the Puna-Altiplano merges with the Main Cordillera of the high Andes. Most investigators attribute uplift and crustal thickening of the Puna-Altiplano with magmatic addition playing a secondary role (e.g., Isacks, 1988; Allmendinger et al., 1990, 1997). Prominent fold-thrust belts to the east provide a temporal record of this crustal shortening. These belts include the Subandean and Eastern Cordillera and Santa Bárbara Ranges east of the plateau, and the Precordillera and block-faulted Pampean Ranges east of the Main Cordillera.

A distinctive feature of the central Andes is the relatively shallow dip (<30°) of the subducting Nazca plate beneath South America compared to other circum-Pacific subduction zones. As recognized by Barazangi and Isacks (1976) and refined by Cahill and Isacks (1992), the Nazca plate can be divided at depths of ~90–135 km into nearly flat segments, above which there is no volcanism, that are flanked by relatively steeper segments associated with active volcanism (Fig. 1). The Chilean flat-slab segment between 28° and 33°S has a relatively smooth northern transition and an abrupt southern transition to the steeper segments (Fig. 1). In terms of this modern slab geometry, the El Indio belt is above the center of the Chilean flat slab, the Maricunga–Farallon Negro district above the northern transition, the El Teniente district above the southern transition, and the Potosí district above the steeper slab to the north.

Rationalizing the relationship among late Cenozoic central Andean uplift, crustal thickening, and Miocene mineralization requires understanding how the geometry of the subducting Nazca plate has changed since the breakup of the Farallon plate and the initiation of fast, nearly orthogonal convergence at ~26 Ma (Pardo Casas and Molnar, 1987). The model for evolving slab geometry used here is that proposed by Isacks (1988) on the basis of seismologic, structural and topographic constraints and modified by Kay et al. (1999) on magmatic considerations. The basic premise is that the slab beneath the modern shallow subduction zone has shallowed as the slab behind the central Puna-Altiplano has steepened. Figure 2 compares the end-member early Miocene and modern situations, and Figures 3 and 4 show reconstructed lithospheric sections depicting the temporal evolution of transects through the Chilean flat slab and the Puna-Altiplano.

CHEMICAL CLUES TO TEMPORAL CHANGES IN MAGMATIC AND TECTONIC PROCESSES ASSOCIATED WITH MINERALIZATION

Important clues to processes occurring over a shallowing subduction zone come from magmas containing chemical components from the evolving slab, the overlying...
mantle wedge, and the crust. Uniquely, diagnostic chemical fingerprints can put restrictions on evolving temperature-pressure, chemical, and fluid profiles in the mantle and crust. Chemical analyses of more than 500 pre-, syn-, and postmineralization samples from the El Indio, Maricunga–Farallon Negro, and El Teniente districts all show the relatively high K, Ba, and Th, and low Ta concentrations expected in magmas erupted over a subduction zone (Kay et al., 1999). Central to the discussion below are the rare earth elements (REE), which show a relatively small range of La/Sm ratios and a wide range of Sm/Yb ratios (Fig. 5).

Increasing Sm/Yb ratios mostly reflect pressure-dependent changes from clinopyroxene to amphibole to garnet in the mineral residue in equilibrium with evolving magmas (review in Kay and Kay, 1991). Following the proposition of Hildreth and Mooibath (1988) that arc magmas in a compressional regime evolve from mafic parent magmas in the lower crust and using Andean southern volcanic zone magmas and crustal thicknesses as depth indicators, Sm/Yb ratios can serve as guides to relative crustal thicknesses. Because breakdown pressures are influenced by factors like bulk composition and temperature, inferred depths are approximations. As a rough guide in mafic lavas, clinopyroxene is dominant at depths of <35 km, amphibole from ~30–45 km, and garnet at >45–50 km.

MAGMATISM, DEFORMATION, AND MINERALIZATION OVER A SHALLOWING SUBDUCTION ZONE

The lithospheric cross sections in Figure 3 illustrate a working model for the post–early Miocene shallowing of the Chilean subduction zone that accounts for magmatic, structural, and basin evolution over the modern flat-slab region and its borders (e.g., Kay et al., 1991, 1999; Jordan et al., 1993; Kay and Abbuzzi, 1996). Shallowing from ~18–8 Ma can account for decreasing amounts of volcanism and cessation of andesitic volcanism by ~9 Ma in the Main Cordillera, as well as an eastward broadening of the volcanic arc, the compressional deformation front, and the foreland basin system into the Precordillera. Continued shallowing of the slab after ~7 Ma can explain eastward expansion of deformation and magmatism into the Pampean Ranges, and the end of volcanism across the transect at ~5 Ma as the asthenospheric wedge became too thin in the west and the slab too dehydrated in the east to flux mantle melting. In concert with shallowing, the mantle and crust over the slab became increasingly hydrated under the Main Cordillera and the zone of hydration broadened eastward as decreasing asthenospheric circulation increasingly limited melting and fluid removal (e.g., Kay and Gordillo, 1994). As shallowing proceeded, magnetically weakened lower crust beneath the Main Cordillera thickened (Kay et al., 1991) in conjunction with crustal shortening in the Precordillera and the Sierras Pampeanas (e.g., Allmendinger et al., 1990), as well as with shortening in the forearc. Mass balance considerations require contemporaneous thinning of the continental lithosphere (Kay and Abbuzzi, 1996).

Important Miocene mineralization took place in the Chilean flat-slab region as the subduction zone was shallowing. Evidence for the temporal and spatial association between mineralization and Miocene magmatic stages and deformational events over the shallow subduction zone beneath the El Indio, Maricunga–Farallon Negro, and El Teniente mineral districts is summarized in Table 1 and discussed below.

Figure 4: Schematic lithospheric cross sections along the central Puna-Altiplano Plateau transect lat ~22°–23°S showing temporal changes in subducting slab geometry, crustal thickness and areas of active volcanism and deformation before, during, and after mineralization. Red striped regions are crustal magma chambers below master fault detachment that feed ignimbrites (in red). Figure based on Kay et al. (1999).

### Table 1. Late Oligocene to Recent Tectonic, Magmatic, and Mineralization History of the Chilean Flat-Slab Region

<table>
<thead>
<tr>
<th>Ma</th>
<th>26°–28°S Maricunga Transect</th>
<th>29°–31°S El Indio Belt</th>
<th>32°–34°S El Teniente District</th>
</tr>
</thead>
<tbody>
<tr>
<td>7–5</td>
<td>Flat-slab magmatism ends; arc migrates east on margins of flat slab; uplift, crustal thickening.</td>
<td>Au-Farallon Negro ~7–6 Ma</td>
<td>Cu-El Teniente south ~4.9 Ma</td>
</tr>
<tr>
<td></td>
<td>Jatabeche dacite/Pircus Negras andesite (&lt;50–65 km)</td>
<td>Au-El Indio ~ends at ~6.5 Ma Vallecito Fm / Vacas Heladas Ignimbrite (&lt;50 km)</td>
<td>Cu-Los Bronces ~7–5 Ma (&gt;45 km)</td>
</tr>
<tr>
<td>11–7</td>
<td>Silicic magmatism to north; last andesitic lavas in flat-slab region; andesitic stratovolcanoes to south.</td>
<td>Copiapó Ignimbrite Complex</td>
<td></td>
</tr>
<tr>
<td>~17–9</td>
<td>Andesitic to dacitic stratovolcano complexes; Precordillera deformation and crustal thickening.</td>
<td>Au-Gold Porphyries ~12–10 Ma</td>
<td>Cu-Pachon in north ~10–9 Ma</td>
</tr>
<tr>
<td></td>
<td>Cadillal Group</td>
<td>Au-El Indio ~begins at ~10 Ma</td>
<td>El Teniente (Farrallones Fm.)</td>
</tr>
<tr>
<td></td>
<td>Ojos de Maricunga Group (&gt;50 km)</td>
<td>Tambo Formation ~13–9 Ma</td>
<td>Volcanic/Plutonic Complex (&lt;35–40 km)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cerro de Las Tórtolas Fm. ~17–14 Ma</td>
<td>(&gt;45 km)</td>
</tr>
<tr>
<td>20–18</td>
<td>Relative magmatic lull, deformation, and crustal thickening.</td>
<td>Au-La Coipa-like domes ~23–21 Ma</td>
<td>Coya Machali Formation</td>
</tr>
<tr>
<td>25–20</td>
<td>Arc and backarc magmatism in all regions; more dacitic to andesitic in north; more bimodal to south.</td>
<td>Las Máquinas backarc basalt ~23 Ma</td>
<td>El Teniente (Farrallones Fm.)</td>
</tr>
<tr>
<td></td>
<td>Refugio/La Coipa Group (andesite) (&lt;40–45 km)</td>
<td>Escal布ro Formation ~21–17 Ma</td>
<td>Explained eastward expansion of deformation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tiltil Formation ~27–21 Ma</td>
<td>and magmatism into the Pampean Ranges.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(~35–40 km)</td>
<td>(&gt;35–35 km)</td>
</tr>
</tbody>
</table>

Note: See Kay et al. (1999) for further references to these transects.
area. Each successive volcanic stage has a distinctive distribution and composition, shows a decrease in overall erupted volume, and is generally bounded by compressional deformation (Kay et al., 1991, 1999).

The initial Doña Ana Group is characterized by voluminous 27–21 Ma andesitic to rhyodacitic tuffs that unconformably underlie 21–17 Ma mafic anidesite flows and are cut by small shallow intrusives (Martin et al., 1997). Small ~23 Ma (Las Máquinas) backarc alkali basalt flows are related to faults. Low-pressure pyroxene-bearing mineral residues are anecdotally associated with these magmas (Fig. 5; Kay et al., 1991) and consistent with ascent from lower levels of a normal thickness crust over a relatively steep subduction zone. This first volcanic stage terminated with high-angle reverse faulting in the Main Cordillera and initiation of thrust faulting in the Precordillera (Jordan et al., 1993). The second volcanic stage, emplaced near the end of the 26–21 Ma tectonic episode, is characterized by backarc amphibole-bearing andesitic to dacitic units (Martin et al., 1997). Their REE patterns are consistent with equilibration in a thickening crust in which the mafic mineral residue changed from amphibole to garnet-bearing in the final stages (Fig. 5; Kay et al., 1991). This second stage also included small backarc amphibole-bearing andesitic to dacitic centers that extended into the Precordillera. The end of the second volcanic stage overlaps the peak of Precordillera thrust faulting at ~11–9 Ma (Jordan et al., 1993) and initiation of the major mineralization episode in the El Indio belt at ~10 Ma which lasted until 6.5 Ma (Clavero et al., 1997; Bissig et al., 2000). The terminal volcanic stage consists of minor ~7–6 Ma hornblende-bearing dacitic centers with amphibole-bearing residual mineralogy in the Main Cordillera (Vallejo et al., 1993) and Precordillera, and mafic andesitic to dacitic centers in the Maricunga transect east of the arc (Gardeweg et al., 1997). Their REE patterns are consistent with a more garnet-rich residual mineral assemblage in transition to amphibole-bearing mineral residues. Mild extension and normal faulting at this time are consistent with models of stress relaxation during porphyry mineralization (Tosdal and Richards, 2001).

The Maricunga Transect. Deformational and magmatic peaks in the Maricunga transect to the north near the boundary with the steeper slab are virtually analogous to those in the El Indio belt, but initial mineralization is older (Kay et al., 1994; Mpodozis et al., 1995). This mineralization at 25–21 Ma is linked to dacitic domes equilibrating with garnet-bearing minerals (Vila and Sillitoe, 1991; Mpodozis et al., 1995). Unlike first-stage El Indio magmas, REE patterns of these magmas point to amphibole-bearing lower crustal residues (Fig. 5) consistent with a thicker crust over a shallower subduction zone (see Fig. 2). This episode is followed by deformation and diminished volcanism from 20 to 18 Ma. Andesitic units that erupted at the beginning of the next volcanic stage from 17–12 Ma have steeper REE patterns consistent with a more garnet-rich residual mineralogy at deep levels of a thicker crust. In accord with this observation, regional uplift in the middle Miocene is signaled by thick sequences of alluvial sediments (Atacama gravels) which show syntectonic deformation east of the arc (Gardeweg et al., 1997).

The second mineralization episode is linked to the emplacement of 13–10 Ma fault-controlled “gold porphyries” (e.g., Maricunga Belt; Sillitoe et al., 1991) near the end of the second volcanic stage. REE patterns of these magmas again indicate an amphibole-bearing mineral residue. Mild extension and normal faulting at this time are consistent with models of stress relaxation during porphyry mineralization (Tosdal and Richards, 2001). The final stages of Maricunga belt volcanism are dominated by the ~11–7 Ma dacitic units from the Copiapó center, and 7–5 Ma mafic anidesite flows.

Important central Andean Miocene mineralization also occurred in the northern Puna and southern Altiplano near 21°–23°S in an area where the present slab dip is steep.

**Figure 5:** Plots of La/Sr (light REE) vs. Sm/Yb (heavy REE) ratios for more than 500 magmatic rocks from the Maricunga–Farallon Negro, El Indio, and El Teniente districts. Premineralization or between mineralization (Maricunga belt) units are in blue, synmineralization units in red, and postmineralization units in yellow. Hatched fields are plutonic units. Thick lines enclose fields for magmas erupted near times of gold (Au) and copper (Cu) mineralization. Presence of plutonic units and Cu rather than Au in the El Teniente district reflects greater erosion in this region. Samples in Au and Cu field have Sm/Yb ratios that are in equilibrium with amphibole-bearing residual mineral assemblages in transition to garnet-bearing. Data sources in Kay et al. (1999) and Kay and Mpodozis (1999).
SUBDUCTION ZONE AND A THICKENING CRUST MODEL FOR MINERALIZATION OVER A SHALLOWING
al., 1993).

Brazilian shield (Allmendinger et al., 1997). No major mineralization
were triggered by horizontal compressional collapse of the melt-
migrated eastward. Plateau uplift accompanied Subandean Belt
terminated under the plateau region and upper crustal deformation
huge late Miocene plateau ignimbrite sheets as brittle deformation
uplift. Mineralization at ~14–12 Ma appears to be associated with the
mantle-derived magmas in a compressional regime provides a
overlying hydrated mantle and lower crust. Heating of the crust by
Sacks, 1999). Steepening of the slab in the middle Miocene increased
mantle hydration above a shallowly subducting slab (James and
is interpreted to have inhibited arc magmatism while enhancing
mineral residues in normal thickness continental crust over relatively steep
Miocene mineralization occurred as the mafic mineral residue
samples in Figure 5 indicate that major periods of central Andean
Miocene mineralization occurred as the mafic mineral residue
changed from hornblende to garnet. Garnet-dominated signatures are
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Maricunga, El Teniente, and El Indio belts. Major ore deposits are not
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Maricunga-bearing Miocene lavas in the El Indio and El Teniente belts.

Another essential factor linking fluid sources associated with ore
deposits to tectonic stresses is that magmas intruding a thickened
crust under compression have difficulty ascending and evolve at
dept depth leading to high intrusive/extrusive ratio magmatic systems.
Storage at depth promotes repeated crustal melting, enhances crustal
ductility, and makes the crust susceptible to horizontal compressional
failure leading to crustal shortening and thickening. Such conditions
promote pressure-induced amphibole breakdown in lower crustal
melts zones. Multiple melt and freeze cycles in these melt zones could
enhance metal enrichment. The erupted magmas are a combination of
crustal and mantle components that last equilibrated with lower
crustal mineral residual assemblages. These fluids

Highly hydrated amphibole-bearing lower crustal units. These fluids
can release a significant amount of fluid during melting resulting in
hydrous magmas. Fluid can come from amphibole in underplated
magnas and their cumulate and melt residues, as well as from
metamorphosed amphibole-bearing lower crustal units. These fluids
can be liberated as magma are emplaced and cooled in shallow level
magma chambers. Oxidizing conditions, which prevent the early
removal of sulfide minerals and allow metals to be concentrated in the
residual fluids of crystalizing magmas are consistent with trace-
element signatures (e.g., small Eu anomalies with low Sr contents; see

The observation that major central Andean Miocene ore deposits
generally form near the end of a deformational peak in a setting
where compression leads to crustal shortening and thickening
highlights a role for thickened crust in the breakdown of hydrous
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and solidification of long-lived magmatic systems to metal-rich fluid
release over a shallowing subduction zone is addressed for the El
Teniente district by Skewes and Stern (1994). Temporal coincidence
of ductile crustal thickening beneath the arc and upper crustal
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Kurtz et al., 1997; Godoy et al., 1999) and Maricunga (Kay et al.,
1994) transects.

Crustal thickening due to shortening must also be compensated in the
lithospheric mantle. Thickening of the lithosphere reduces space for the
asthenospheric wedge above the shallowing slab and forces
of mineralization are in equilibrium with an amphibole-bearing mafic
mineral residue that is changing to garnet. These features are
incorporated in the general model in Figure 6 which builds on long-
standing ideas of associating these deposits with hydrous magnas
over subduction zones (e.g., Barnes, 1997).

Linking mineralization with a shallowing (or formerly shallow) subduction zone over a thickening crust is important as decreasing
mantle flow in the cooling wedge above the dehydrating, shallowing
slab increasingly limits melting and fluid removal from the wedge.
Melts entering the thickening lower crust from this wedge become
increasingly hydrous as fluids are progressively concentrated in the
cooling mantle. As shown in stage 1 of Figure 6, arc magnas erupted
through a normal thickness crust before the slab shallows have
anhydrous residual mineral assemblages and are not linked to large
ore deposits. In contrast, arc magnas formed above the cooling,
hydrating mantle in stage 2 of Figure 6 contain fluids that cause
amphibole to crystallize in them as they are underplated and intruded
into a thickening crust. This process can occur for as much as 6–8
m.y. before mineralization as shown by eruption of amphibole-
bearing Miocene lavas in the El Indio and El Teniente belts.

An implication for mineralization is that breakdown of amphibole
can release a significant amount of fluid during melting resulting in
hydrous magnas. Fluid can come from amphibole in underplated
magnas and their cumulate and melt residues, as well as from
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TIMING OF TECTONIC AND MINERALIZATION EVENTS

Mineralization events generally correspond to crustal deformational peaks which can be argued to be approximately contemporaneous along the Andean front from Peru to Chile. Sibbier and Soler (1991) suggest that peaks near ~17 Ma, ~10 Ma, ~7 Ma, and ~2 Ma occurred at times of little or no westward retreat of the subducting Nazca plate relative to South America. Such a regime could be related to changes in plate directions and spreading rates. Whether mineralization occurs in a given place depends on local crustal thickness, asthenospheric wedge volume, and the geometry of the subducting plate. This unifying model links mineralization to the changing dip of shallow subduction zones beneath continents.

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REFERENCES CITED


Barazangi, M., and Isacks, B.L., 1976, Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America: Geology, v. 4, p. 682-692.


Hildreth, W., and Mountber, S., 1988, Crustal contributions to arc magmatism in the Andes of central Chile: Contributions to Mineralogy and Petrology, v. 98, p. 455-489.


Martin, M., Claverio, J., and Hpodozis, C., 1995, Eocene to Late Miocene magmatic development of the El Indio belt, ~30°S, north-central Chile: Antofagasta, Chile, VIII Congresso Geológico Chileno Actas, p. 149-153.


Pardo-Casas, F., and Molnar, P., 1987, Relative motion of the Nazca (Farallon) and South American plates since Late Cretaceous time: Tectonics, v. 6, p. 233-248.


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