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# Fission Track Thermochronology of the Domeyko Cordillera, Northern Chile: Implications for Andean Tectonics and Porphyry Copper Metallogenesis

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**Abstract** — The Domeyko Cordillera, the westernmost uplifted crustal block of the composite High Andes of northern Chile (20°S to 28°S) hosts a narrow N-S trending belt of Late Eocene-Oligocene giant porphyry copper deposits, which include Collahuasi, Chuquicamata, El Abra, La Escondida, and El Salvador. These deposits are spatially and genetically associated with the closing igneous activity along this range, prior to a 30 km eastward jump of the magmatic front in response to tectonic plate interaction. The porphyry copper deposits are also spatially associated with a major intra-arc strike-slip shear system, the Domeyko Fault System. Although the tectonic uplift of the 3000 m to 5000 m high range has generally been assumed to be mostly Miocene in age, field relationships suggest that the Domeyko Fault System and tectonic uplift were active as early as the Eocene, coinciding with porphyry copper emplacement between 41 Ma and 30 Ma.

Apatite fission track (FT) thermochronology provides both age data and a time-temperature history for rocks since they cooled below a temperature of ca. 125°C (equivalent to a depth of 4 km to 5 km under normal geothermal gradients) on their way to the surface during exhumation, or after a heating event. Apatite FT data from the Paleozoic crystalline basement of the Domeyko Cordillera indicate that at least 4 km to 5 km of rocks were eroded during exhumation of this tectonic block between ca. 50 Ma to 30 Ma (Middle Eocene to Early Oligocene), a time that immediately precedes and overlaps with the emplacement of giant porphyry copper deposits. The FT data constrain the age and duration of a period of crustal thickening and extensive erosion known as the Incaic compression, an event recognized in the Andes of Chile and Peru. Assuming that cooling was due to denudation alone, modelling of the FT data allow estimation of denudation rates between 200 and 100 m/My during this period. In contrast, exposures of pre-porphyry Paleocene intrusives on the western edge of the Domeyko Cordillera reveal apatite FT ages that are concordant with biotite $^{40}Ar$-$^{39}Ar$ dates, indicating shallow emplacement, fast cooling and negligible exhumation. The apatite FT ages of the Chuquicamata and El Abra porphyry copper deposits are only marginally younger than their $^{40}Ar$-$^{39}Ar$ dates, implying fast cooling/exhumation of shallow mineralizing systems (ca. 2 km to 3 km). Their FT time-temperature history is compatible with extremely low rates of exhumation (ca. 50 m/My) since about 30 Ma (Early Oligocene). This limited extent of erosion, due to exceptional aridity, has greatly contributed to the preservation of rich supergene enriched blankets above some of the porphyry copper deposits.

A model is proposed, which is consistent with field relationships, geochemical, and geochronological data. High relative velocities and intermediate-angle oblique convergence between the Nazca and South American plates in the Eocene-Early Oligocene led to intra-arc transpression partitioned into reverse faults of opposing vergence and a right-lateral shear system (the Domeyko Shear System) localized along the magmatic front. Crustal shortening and thickening led to uplift within the Domeyko Fault System and to erosion of at least 4 km to 5 km. Crustal thickening deepened the zone of magma generation, and magmas accumulated in the lower crust, allowing small volumes to rise rapidly to shallow levels along transtensional domains of the regional shear system. A similar model of crustal thickening and shortening before and during porphyry copper emplacement in zones of transtension may also apply to the Miocene-Pliocene giant porphyry copper province of central Chile. 

## Introduction

The northern Chilean province of the Central Andes (Fig. 1) is remarkably rich in copper deposits derived from igneous activity associated with sustained subduction of oceanic lithosphere under the South American margin during the last 200 My. Remarkably, most of the known giant porphyry copper type deposits preserved in the region developed in the short interval between 41 Ma and 31 Ma, a metallogenetic epoch representing merely 5% of post-Paleozoic subduction time (Clark and Zentilli, 1972; Zentilli, 1974; Clark et al., 1976; Sillitoe, 1988; Maksaev and Zentilli, 1988, Maksaev et al., 1988a, 1998b, Maksaev, 1990; Clark, 1993). Furthermore, these giant deposits, which include Collahuasi, El Abra, Chuquicamata, La Escondida, El Salvador, and Potrerillos, were all localized within a narrow longitudinal uplifted tectonic block known as the Domeyko Cordillera, the westernmost range of the composite High Andes of the Antofagasta and Atacama desert regions of northern Chile (Fig. 2). The high-elevation (3500 m to 5000 m) Domeyko Cordillera (also referred to as the Precordillera; Abele, 1988) is underlain by Paleozoic to Triassic crystalline rocks, partially covered by folded Jurassic and Early Cretaceous marine strata, in turn overlain unconformably by Upper Cretaceous (non-marine) red-beds and predominantly Paleogene volcanic rocks and various intrusives (Maksaev, 1990; Boric et al., 1990). A N-S-trending belt of Late Eocene-Oligocene porphyry copper deposits (Figs. 1 and 2) extends along the Domeyko Cordillera, presumably controlled by the longitudinal Domeyko Fault System, known in the Chuquicamata regions as the West Fault or West Fissure system (Lopez, 1939, 1942; Reutter et al., 1991; Lindsay et al., 1995; Tomlinson and Blanco, 1997a, 1997b; Lindsay, 1998).

In an attempt to understand what combination of geological conditions led to this remarkable copper metallogenetic epoch and its related tectonics, a long-term multidisciplinary study was carried out from Dalhousie University, Halifax, Canada, in collaboration with SERNAGEOMIN, Chile, leading to the doctoral thesis of the first author, and contributing to a comprehensive geological review of the Antofagasta region (Maksaev, 1990; Boric et al., 1990). This paper discusses apatite fission track (hereinafter FT) age and track-length data from the Paleozoic to Cenozoic intrusive igneous rocks. These provide information on the Cenozoic thermal evolution of the Domeyko Cordillera, demonstrate that considerable exhumation took place since the Eocene, and place constraints on tectonic and metallogenetic models for the porphyry copper mineralization at this active margin.

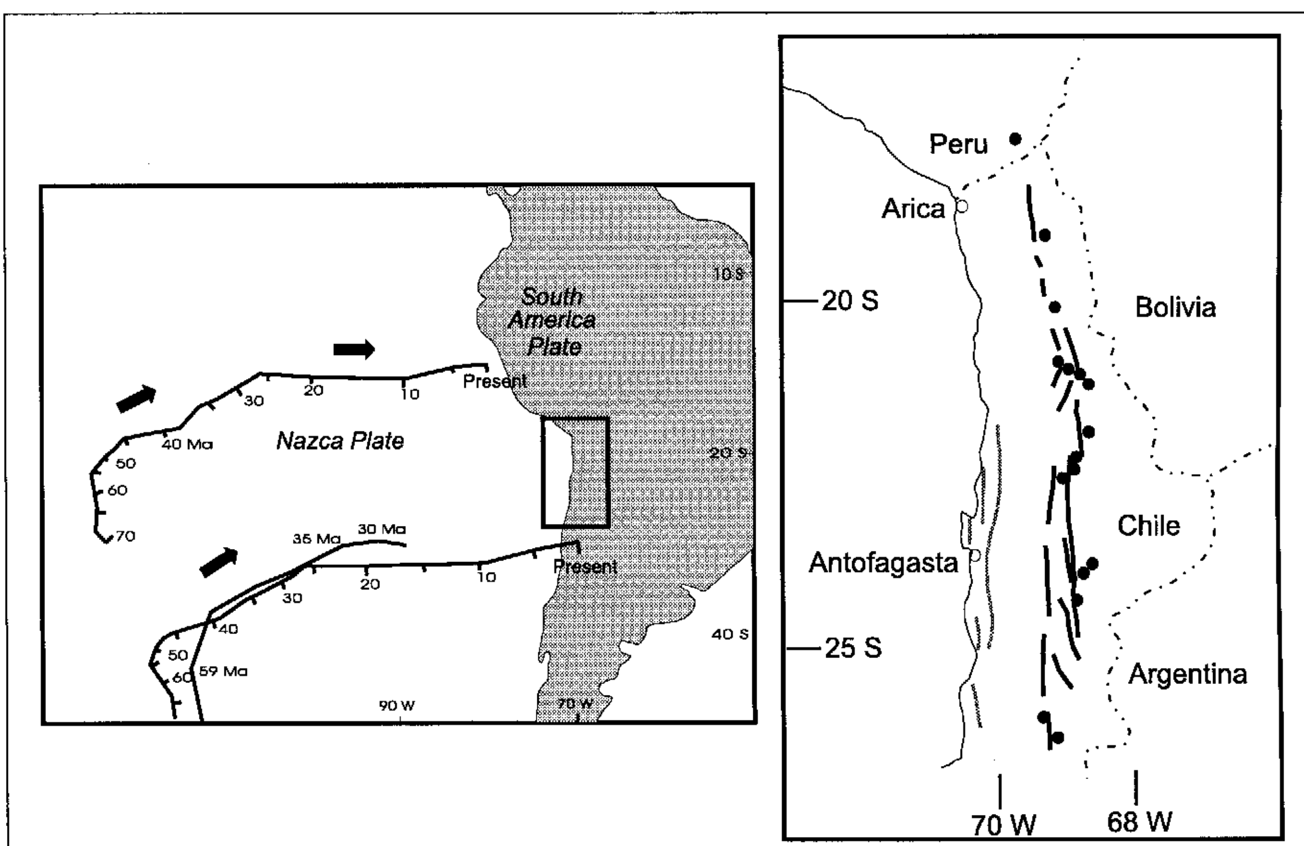


Fig. 1. Location map of the study area on the western margin of South America. Angle of incidence and rates of convergence between the Nazca and South American Plate from Pardo-Casas and Molnar (1987) are schematically indicated. The spatial association of porphyry copper deposits (black circles) and the Domeyko Fault System (bold lines) is emphasized.

## Uplift of the Domeyko Cordillera and Igneous Activity

Stratigraphic evidence indicates that the crustal block of the Domeyko Cordillera remained below sea level at least until the Early Cretaceous (Chong, 1977). It was the site of non-marine red-bed sedimentation and volcanism during the Late Cretaceous, and finally it developed high topographic relief during Tertiary times as a result of two compressive tectonic events, the Incaic and Quechua tectonic phases. The Incaic phase is constrained in this region by field and K-Ar age data from the Domeyko Cordillera

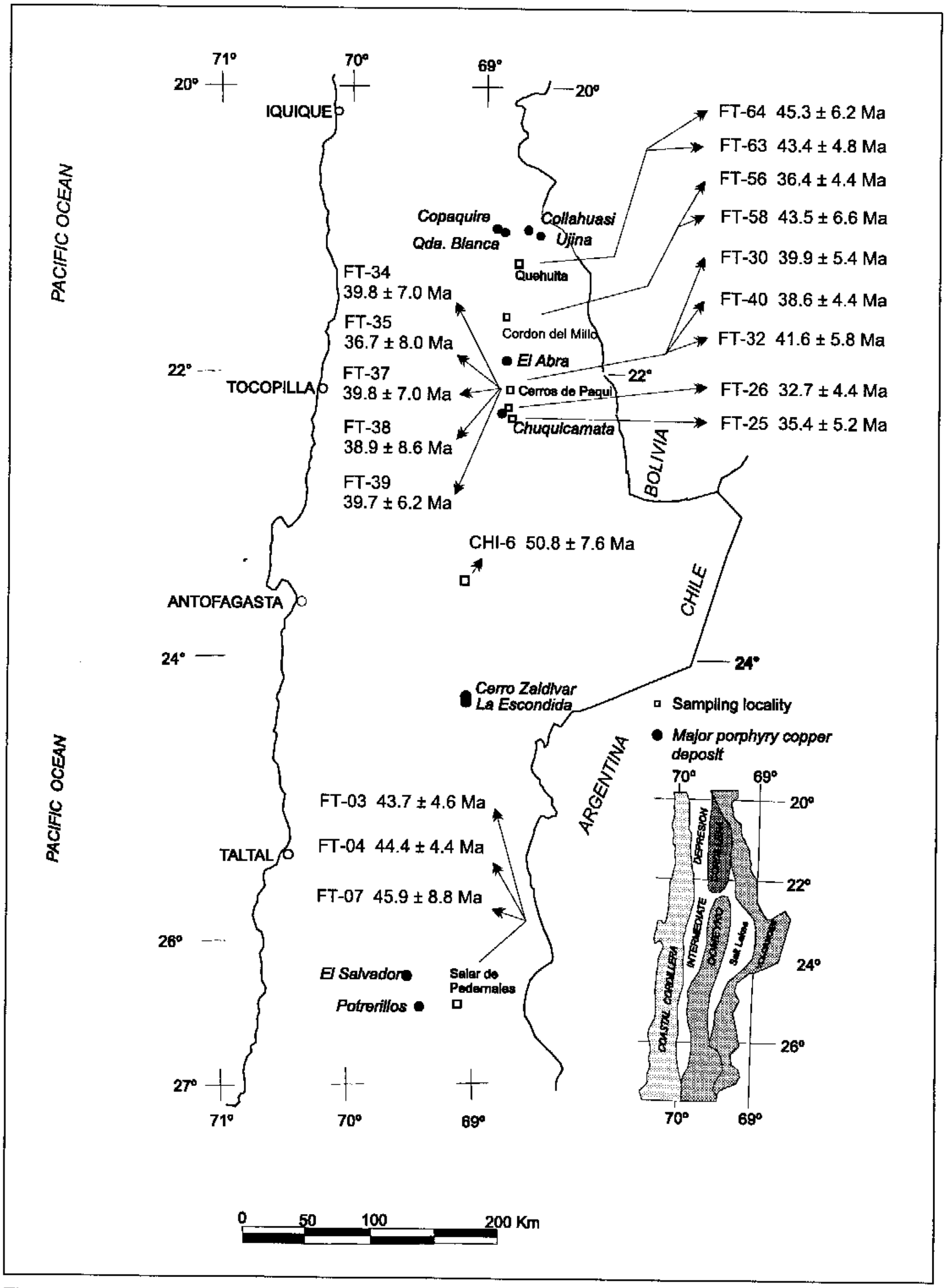


Fig. 2. Sample locations and apatite FT ages of crystalline basement rocks of the Domeyko Cordillera listed in Tables 1 and 2. The inset map shows the major geomorphic zones (e.g., Domeyko Cordillera, Intermediate Depression, Neogene volcanoes) mentioned in the text. Modified from Maksaev (1990). Sample CHI-6 from Andriessen and Reutter (1994).

within the Middle Eocene to Early Oligocene interval (Maksaev, 1979). The Incaic tectonism is usually considered to be responsible for much of the shortening recorded in the Central Andes in Peru (i.e., Mégard, 1987), and for causing a major change in the physiography of northern Chile (Mortimer and Saric, 1975).

High relative velocities and intermediate-angle oblique convergence (Fig. 1) between the Nazca Plate and the South American Plate (Pilger, 1984; Pardo-Casas and Molnar, 1987) have resulted in crustal transpression and the initiation of the Domeyko Fault System during the Late Eocene (Scheuber and Reutter, 1992; Reutter et al., 1991, 1993, 1996; Scheuber et al., 1994). The analysis of ductile deformation features, mineralized stockwork systems, and vein arrays at Chuquicamata, indicate the emplacement and mineralization at Chuquicamata developed within an active right-lateral strike slip fault zone in the Eocene-Oligocene (Lindsay et al., 1995; Lindsay, 1998). Maksaev and Zentilli (1988) and Maksaev (1990) proposed that this dextral regime persisted through the time of formation of the porphyries, but Tomlinson and Blanco (1997a, 1997b) suggest that the dextral episode was a short-lived event, and that left-lateral movement prevailed in the Domeyko Fault System since the Oligocene.

The resulting erosion from this tectonism produced extensive alluvial sedimentation during the Oligocene-Early Miocene at the eastern and western foothills of the Domeyko Cordillera. The following sedimentary formations were deposited above the Incaic unconformity: Sichal (Maksaev, 1978), Tambores (Dingman, 1963), San Pedro (Brüggen, 1942), Quebrada Justo (Lahsen, 1969), and Pampa Mulas (Chong, 1977). Until this study, the age of the onset of uplift was unconstrained, although Hammerschmidt et al. (1992) dated biotite from two ash-fall tuffs, separated by a distinct angular unconformity, obtaining identical $^{40}Ar/^{39}Ar$ ages of 38.50 ± 0.90 Ma and 38.45 ± 0.60 Ma, which they suggested would be the age of the Incaic Phase in the Domeyko Cordillera. The present study, however, suggests that the Incaic Phase was a more gradual, protracted event than envisaged by Hammerschmidt et al. (1992). Regarding the end of the deformation, May et al. (1996) dated ash (30.15 ± 0.26 Ma) in the Oligocene Calama Fm, which overlies the unconformity in the Calama Basin, east and south of the Chuquicamata district. The clastic units resulting from the Incaic Phase record mostly clastic alluvial, high-energy sedimentation, but local evaporites provide evidence for the long-lasting arid conditions prevailing in this desert region, and intercalated tuffs are evidence of minor explosive volcanism concurrent with the emplacement of the porphyries.

The subsequent Quechua compression (Mégard, 1987; Maksaev, 1979) during the Miocene was accompanied by thrust and strike-slip faulting, and further uplift of the Domeyko Cordillera to its present elevation of ca. 5000 m.a.s.l. (Maksaev, 1990). However, the extremely arid conditions since the Miocene in northern Chile (Alpers and Brimhall, 1988) and geomorphologic factors such as the barrier effect of the Coastal Cordillera for regional drainage, strongly limited further denudation. Sedimentation in the region was largely limited to the formation of piedmont alluvial plains that border the mountain ranges, evaporites, and the deposition of tuffs originating in the volcanic front further east (Maksaev, 1990). Regional geomorphologic studies have shown that tectonic processes have exerted a strong control of the relief from the late Cretaceous to the present. Yet unlike other segments of the Andes or other mountain ranges, due to extreme aridity, erosion has been largely unable to keep pace with mountain building, resulting in considerable uplift (Mortimer and Saric, 1975; Mortimer, 1980; Abele, 1988, 1989; Maksaev et al., 1988a).

A typical subduction-related plutonic-volcanic arc evolved during the Late Cretaceous, throughout the Paleogene, and most of the Eocene along the Domeyko Cordillera; this magmatic arc system also involved part of the present Intermediate Depression to the west (see inset in Fig. 2). The Late Eocene – Oligocene major porphyry copper deposits are associated with the latest recorded igneous activity of this cycle. These porphyries are not only restricted to a small age-range in a very narrow belt within the Domeyko Cordillera, but also have a more limited range of geochemical composition compared with the pre-porphyry calc-alkaline arc magmatic rocks; they are also less contaminated by the upper crust, and seem to have a deeper crustal or mantle source (Maksaev, 1990; Zentilli et al., 1994b; 1995).

After a break in volcanic activity, the magmatic front jumped more than 30 km to the east, and developed the Neogene-Quaternary volcanic complex ("volcanoes" in inset map in Fig. 2) of the Central Andes, which locally extends more than 100 km east of the continental divide, into Bolivia and Argentina (Caelles, 1979; Sasso and Clark, 1998; Kay et al., 1999).

From his detailed structural fabric studies at Chuquicamata, Lindsay (1998) deduced that since the early Oligocene (ca. 31 Ma), sinistral fault movements displaced supergene enriched and earlier formed hypogene alteration-mineralization zones, and that the displacement was concentrated along the West Fault. Finite sinistral displacements of 25 km (Reutter et al., 1996) to 35 km (Dilles et al., 1997; Tomlinson and Blanco, 1997a, 1997b) and 40 km (Ambrus, 1979) have been estimated for the West Fault in the Chuquicamata region.

## Fission Track Thermochronology

### *The Method*

Rarely applied to ore deposit studies (Naeser and Cunningham, 1984; Arne et al., 1989, 1990; Arne 1992; Ryan et al., 1992), apatite FT thermochronology is a technique more extensively used for time-temperature modelling in sedimentary basins (Naeser et al. 1989), fluid flow (Duddy et al., 1994), timing and rates of tectonic uplift (Fitzgerald et al., 1986; Gleadow and Fitzgerald, 1987; Benjamin et al., 1987; Brown et al., 1994; Laubacher and Naeser, 1994) and the timing of thrusting (Arne et al., 1998). The reader is referred to helpful reviews of the technique by Hurford (1986), Naeser et al. (1989), Ravenhurst and Donelick (1992), Wagner and Van den haute (1992), and Gallagher et al. (1998).

Apatite FT dating is a method based on the spontaneous fission of $^{238}U$ present in trace amounts in the crystal lattice of the ubiquitous accessory mineral apatite. Fission produces the expulsion of two positively charged nuclear fragments that repel each other in opposite directions, leaving a trail of radiation damage about 16 μm in length (latent fission tracks) in the host apatite crystals. These latent tracks can be made visible by etching the polished surface of a grain mount

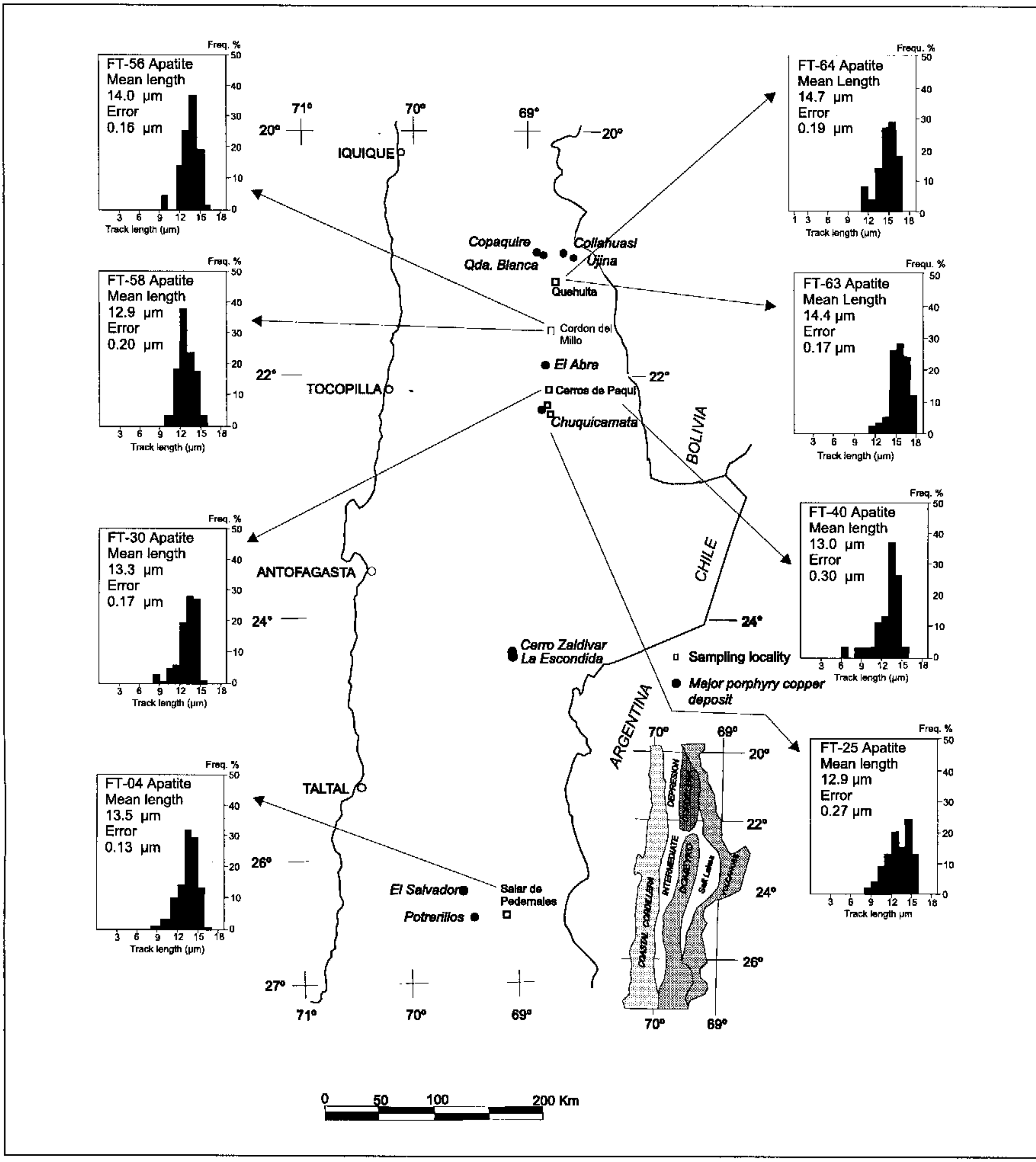


Fig. 3. Distribution of apatite track-length data from basement intrusive samples of the Domeyko Cordillera. See text for discussion. Modified from Maksaev (1990).

Table 1. Sample location and lithology

| Sample Identification | Lat.-Long. (S)-(W) | Elevation (m) | Rock Type | Locality |
|---|---|---|---|---|
| **1. Granitoids of the Domeyko Cordillera** | | | | |
| 1A. Northern plutons | | | | |
| FT-64 | 21°04'10" - 68°42'10" | 4560 | Hb granite | Quehuita |
| FT-63 | 21°09'45" - 68°45'30" | 4250 | Hb granite | Quehuita |
| FT-58 | 21°32'30" - 68°48'30" | 4150 | Bt-Hb granite | Cordon del Millo |
| FT-56 | 21°30'30" - 68°48'50" | 4060 | Bt-Hb granodiorite | Cordon del Millo |
| FT-32 | 22°04'00" - 68°47'55" | 3930 | Diorite (68 Ma K/Ar) | Cerros de Paqui |
| FT-34 | 22°03'06" - 68°44'15" | 3600 | Monzogranite | Cerros de Paqui |
| FT-30 | 22°04'50" - 68°46'15" | 3500 | Diorite (68 Ma K/Ar) | Cerros de Paqui |
| FT-37 | 22°07'00" - 68°50'00" | 3390 | Bt-Hb granite | Cerros de Paqui |
| FT-38 | 22°07'10" - 68°49'40" | 3385 | Diorite | Cerros de Paqui |
| FT-39 | 22°07'10" - 68°49'40" | 3385 | Granite | Cerros de Paqui |
| FT-35 | 22°05'30" - 68°44'20" | 3340 | Monzogranite | Cerros de Paqui |
| FT-40 | 22°07'55" - 68°49'40" | 3275 | Bt granite | Cerros de Paqui |
| FT-26 | 22°16'10" - 68°52'40" | 3150 | Elena granodiorite | East of Chuquicamata |
| FT-25 | 22°19'52" - 68°51'15" | 2600 | Hb diorite | East of Chuquicamata |
| 1B. Southern plutons | | | | |
| FT-07 | 26°21'30" - 69°17'05" | 3650 | Granodiorite | Salar de Pedernales |
| FT-04 | 26°22'00" - 69°16'20" | 3425 | Bt Hb granodiorite | Salar de Pedernales |
| FT-03 | 26°20'30" - 69°15'15" | 3370 | Granite | Salar de Pedernales |
| **2. Cerros de Montecristo Pluton** | | | | |
| FT-14 | 22°20'00" - 69°05'50" | 2880 | Monzodiorite | Montecristo |
| FT-13 | 22°19'30" - 69°04'50" | 2690 | Monzonite | Montecristo |
| FT-12 | 22°20'55" - 69°08'55" | 2440 | Diorite | Montecristo |
| FT-11 | 22°20'10" - 69°08'40" | 2435 | Diorite | Montecristo |
| FT-9 | 22°18'55" - 69°11'30" | 2100 | Granodiorite | Montecristo |
| FT-8 | 22°17'15" - 69°11'30" | 1990 | Granodiorite | Montecristo |
| **3. Fortuna Granodiorite Complex** | | | | |
| FT-16 | 22°17'15" - 68°57'50" | 3270 | Hb Bt Gd. porphyry | San Lorenzo |
| FT-43 | 22°14'10" - 68°55'30" | 3190 | Hb Bt granodiorite | Fortuna |
| FT-41 | 22°13'50" - 68°58'50" | 3110 | Hb diorite | Los Picos |
| FT-17 | 22°18'00" - 68°58'30" | 3080 | Hb Bt granodiorite | Fortuna |
| FT-15 | 22°18'50" - 68°58'35" | 2970 | Hb Bt granodiorite | Fortuna |
| FT-24 | 22°20'47" - 68°58'00" | 2750 | Hb Bt granodiorite | Fortuna |
| **4. Chuquicamata and El Abra Porphyry Copper Deposits** | | | | |
| FT-19 | 22°16'45" - 68°53'45" | 2580 | East porphyry | Chuquicamata |
| FT-44 | 21°55'06" - 68°50'00" | 3920 | Dacite porphyry | El Abra |
| FT-45 | 21°55'45" - 68°51'05" | 3700 | Hb Bt granodiorite | El Abra |
| FT-48 | 21°55'10" - 68°48'50" | 4050 | Granodiorite | El Abra |

Abbreviations: Gd = Granodiorite; Hb = hornblende; Bt = biotite

with a mild acid, and viewing them under a high power (>1000x) optical microscope. For every FT damage zone that intersects the polished surface of apatite grains, the acid produces etch pits, and these are counted (*density* per unit area). A small proportion of the tracks that lie parallel to the polished surface are only accessed by the acid through other tracks or fractures, and these confined tracks thus reveal their total length, which can be measured under the microscope. The fission process of uranium is continuous and invariable, irrespective of temperature or pressure, therefore the number of spontaneously produced tracks is a function of the original uranium concentration of the apatite (commonly between 1 and 100 ppm U) and the time elapsed since the crystal formed. Thus fission tracks can be viewed as the "daughter products" of $^{238}U$ fission decay, analogous to other parent/daughter relationships used in dating techniques, such as Rb/Sr or K/Ar. The uranium content of the apatite is measured precisely by induction of tracks from $^{235}U$ through irradiation with thermal neutrons in a nuclear reactor. The FT age of individual grains is then computed from the measured track-density data, using an internal calibration factor (*zeta*) and apatite standards of known age (Hurford and Green, 1983; Hurford and Carter 1991). These apparent ages are commonly displayed in radial plots, which reveal whether the individual apparent ages represent one coherent population of apatite, as well as their quality or analytical uncertainty (Galbraith, 1988, 1990).

However, latent tracks are unstable when temperatures rise above 20°C over geologic time periods, their length being progressively shortened. When the temperature reaches about 125°C the tracks totally disappear (are said to *anneal*). Consequently, the calculated age decreases when the thermal effect increases and reduces to zero at about 125°C. The FT apparent age becomes much younger than it would if a temperature of less than ca. 25°C had been maintained. Under normal geothermal gradients, a temperature sufficient to totally anneal the latent tracks is achieved at a depth of ca. 4 km to 5 km, and accordingly, at depth in a well or drill hole the apatite FT apparent age decreases (partial annealing zone) and eventually becomes zero (total annealing zone). Therefore, during erosion/exhumation of a mountain range, the apatite from rock outcrops at high elevations may display

Table 2.

| Sample, Location | Grain | $N_s$ | $N_i$ | $Rho_s$ | $Rho_i$ | $p>X^2$ | $N_d$ | $Rho_d$ | Age (Ma) | Lengths (µm) | Mean (µm) | Error |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | Age Data | | | | | | Length Data | | |
| **1. Granotoids of the Domeyko Cordillera** | | | | | | | | | | | | |
| 1A. Northern plutons | | | | | | | | | | | | |
| FT-64, Quehita, 4560 m | 26 | 530 | 824 | 3.67 | 5.70 | .9735 | 870 | 0.123 | **45.3 ± 6.2** | 51 | 14.7 | 0.19 |
| FT-63, Quehita, 4250 m | 41 | 1159 | 1881 | 4.43 | 7.19 | .9996 | 870 | 0.123 | **43.4 ± 4.8** | 58 | 14.4 | 0.17 |
| FT-58, Cordon del Millo, 4150 m | 37 | 389 | 629 | 2.43 | 3.93 | .9836 | 870 | 0.123 | **43.5 ± 6.6** | 30 | 12.9 | 0.20 |
| Ft-56, Cordon del Millo, 4060 m | 31 | 714 | 1389 | 4.35 | 8.47 | .1824 | 870 | 0.123 | **36.4 ± 4.4** | 56 | 14.0 | 0.16 |
| FT-32, Cerros de Paqui, 3930 m | 51 | 369 | 989 | 1.43 | 3.84 | .9082 | 1225 | 0.194 | **41.6 ± 5.8 (#)** | - | - | - |
| FT-34, Cerros de Paqui, 3600 m | 51 | 258 | 457 | 0.81 | 1.43 | .6428 | 870 | 0.123 | **39.8 ± 7.0** | - | - | - |
| FT-30, Cerros de Paqui, 3500 m | 41 | 516 | 906 | 2.87 | 5.04 | 1.0 | 870 | 0.123 | **39.9 ± 5.4(#)** | 93 | 13.3 | 0.17 |
| FT-37, Cerros de Paqui, 3390 m | 34 | 256 | 459 | 1.68 | 3.01 | .8313 | 870 | 0.124 | **39.8 ± 7.0** | - | - | - |
| FT-38, Cerros de Paqui, 3385 m | 44 | 148 | 270 | 0.61 | 1.11 | 1.0 | 870 | 0.124 | **38.9 ± 8.6** | - | - | - |
| FT-39, Cerros de Paqui, 3385 m | 51 | 351 | 628 | 1.25 | 2.23 | .9936 | 870 | 0.124 | **39.7 ± 6.2** | - | - | - |
| FT-35, Cerros de Paqui, 3340 m | 51 | 146 | 284 | 0.53 | 1.04 | .9998 | 870 | 0.124 | **36.7 ± 8.0** | - | - | - |
| FT-40, Cerros de Paqui, 3275 m | 51 | 660 | 1957 | 2.14 | 6.36 | .6645 | 1200 | 0.199 | **38.6 ± 4.4** | 38 | 13.0 | 0.30 |
| FT-26, Elena Granodiorite, 3150 m | 40 | 519 | 1107 | 1.94 | 4.13 | .1790 | 870 | 0.121 | **32.7 ± 4.4(*)** | - | - | - |
| FT-25, Mesa Granite, 2600 m | 34 | 394 | 772 | 2.59 | 5.07 | 1.0 | 870 | 0.121 | **35.4 ± 5.2** | 46 | 12.9 | 0.27 |
| 1B. Southern plutons | | | | | | | | | | | | |
| FT-07, Salar de Pedernales, 3650 m | 39 | 181 | 421 | 0.96 | 2.24 | 1.0 | 1472 | 0.186 | **45.9 ± 8.8** | - | - | - |
| FT-04, Salar de Pedernales, 3425 m | 31 | 2086 | 3218 | 17.2 | 26.6 | .2521 | 870 | 0.119 | **44.4 ± 4.4** | 101 | 13.5 | 0.13 |
| FT-03, Salar de Pedernales, 3370 m | 27 | 859 | 2170 | 8.14 | 20.6 | .3572 | 1472 | 0.192 | **43.7 ± 4.6** | - | - | - |
| **2. Cerros de Montecristo Pluton** | | | | | | | | | | | | |
| FT-14, Montecristo, 2880 m | 16 | 697 | 1267 | 6.15 | 11.2 | .9252 | 1200 | 0.198 | **62.4 ± 7.4** | 100 | 14.0 | 0.12 |
| FT-13, Montecristo, 2690 m | 22 | 1056 | 1997 | 10.8 | 20.4 | .6968 | 1225 | 0.194 | **58.9 ± 6.0** | - | - | - |
| FT-12, Montecristo, 2440 m | 41 | 1344 | 2574 | 7.17 | 13.7 | .5330 | 1225 | 0.194 | **58.2 ± 5.6** | 101 | 14.1 | 0.13 |
| FT-11, Montecristo, 2435 m | 51 | 1666 | 3607 | 6.66 | 14.4 | .0010 | 1225 | 0.194 | **51.7 ± 5.0** | - | - | - |
| FT-9, Montecristo, 2100 m | 38 | 891 | 1784 | 4.75 | 9.51 | .5494 | 1225 | 0.194 | **55.6 ± 6.0** | - | - | - |
| FT-8, Montecristo, 1990 m | 40 | 667 | 1570 | 4.27 | 10.0 | .2781 | 1225 | 0.194 | **47.4 ± 5.4** | 102 | 13.4 | 0.12 |
| **3. Fortuna Granodiorite complex** | | | | | | | | | | | | |
| FT-16, Fortuna, 3270 m | 31 | 1200 | 3690 | 9.91 | 30.5 | .0781 | 1200 | 0.199 | **37.1 ± 3.6** | 100 | 14.6 | 0.09 |
| FT-43, Fortuna, 3190 m | 25 | 786 | 2853 | 5.16 | 18.7 | .1615 | 1225 | 0.190 | **30.1 ± 3.2** | - | - | - |
| FT-41, Fortuna, 3110 m | 21 | 328 | 1059 | 2.40 | 7.75 | .9455 | 1225 | 0.193 | **34.4 ± 5.0** | 36 | 14.5 | 0.15 |
| FT-17, Fortuna, 3080 m | 28 | 910 | 3272 | 8.32 | 29.9 | .0010 | 1225 | 0.194 | **30.8 ± 3.6** | 100 | 14.8 | 0.09 |
| FT-15, Fortuna, 2970 m | 21 | 547 | 1906 | 3.29 | 11.5 | .2368 | 1225 | 0.194 | **32.0 ± 3.8** | 46 | 14.3 | 0.16 |
| FT-24, Fortuna, 2750 m | 31 | 527 | 1941 | 4.35 | 16.0 | .5459 | 1225 | 0.194 | **30.3 ± 3.6** | - | - | - |
| **4. Chuquicamata and El Abra porphyry copper deposits** | | | | | | | | | | | | |
| FT-19, Chuquicamata, 2580 m | 53 | 302 | 1130 | 1.46 | 5.46 | .9873 | 1200 | 0.197 | **30.2 ± 4.4** | - | - | - |
| FT-44, El Abra Dacite, 3860 m | 22 | 1145 | 3680 | 13.3 | 42.8 | .0001 | 1200 | 0.199 | **36.3 ± 4.4** | 32 | 14.1 | 0.13 |
| FT-45, S. Granodiorite, 3835 m | 31 | 1191 | 3998 | 9.65 | 32.4 | .0003 | 1225 | 0.189 | **33.6 ± 3.8** | - | - | - |
| FT-48, S. Granodiorite, 4050 m | 31 | 1386 | 4460 | 9.59 | 30.9 | .0000 | 1200 | 0.200 | **35.9 ± 4.4** | - | - | - |

Samples with a $p>X^2$ greater than .05 pass the $X^2$ test at the 95% confidence level (i.e., they appear to be composed of one age population). Ages reported for these samples are calculated using pooled statistics. Ages reported for samples which fail the $X^2$ test are the central age (Galbraith and Laslett, 1993). Abbreviations are as follows: $N_s$, $N_i$, $N_d$ = the number of spontaneous (fossil), induced, and flux dosimeter (SRM-614) tracks counted; $Rho_s$, $Rho_i$, $Rho_d$ = density ($x10^5/cm^2$) of spontaneous (fossil), induced, and flux dosimeter (SRM-614) tracks. Error estimates are at the 95% ($2\sigma$) confidence level. All analyses by V. Maksaev using a zeta of 11 533.3 ± 235.6. (#) Pluton with (biotite) K/Ar age of 68 ± 2.5 Ma (Vega and Bordones, 1981).

(*) Sample obtained from 500 m east of the Chuquicamata porphyry copper deposit. It is inferred to be reset by the intrusion/alteration because the FT age is identical to the radiometric ages of the porphyry intrusion.

older apatite FT ages than those in the valleys, because the former passed through the critical isotherm for track retention earlier than the latter, and thus under certain conditions (e.g., rocks having resided within the total annealing zone), a rough estimate of denudation rates can be made (Gallagher et al., 1998). We refer to exhumation rates rather than uplift, because "uplift" of a block may not result in uplift of the surface (net tectonic uplift) if, for instance, erosion keeps pace with the rising of the block (England and Molnar, 1990).

In the 1980s, researchers established the kinetics of track annealing (Green, 1980; Harrison, 1985; Gleadow et al., 1986; Lasslett et al. 1987; Galbraith and Lasslett, 1988, Green et al., 1989) and a relationship between track-length reduction and the delay of the FT clock; hence, results from time-temperature annealing experiments can be extrapolated with some confidence to geologic time scales. Track lengths are measured, plotted in histograms of track-length distributions (Fig. 3), and these are compared with track-length distributions predicted theoretically from track annealing models for apatite with different thermal histories. Several empirical apatite FT annealing models have been developed based on laboratory annealing data extrapolated to geologic time scales (Duddy et al., 1988; Lutz and Omar, 1991; Corrigan, 1991; Willett, 1992, 1997; Willett et al., 1997; Crowley, 1993; Gallagher, 1995).

*Procedures and Results*

Apatite concentrates from samples listed in Tables 1 and 2 were obtained using conventional heavy-mineral

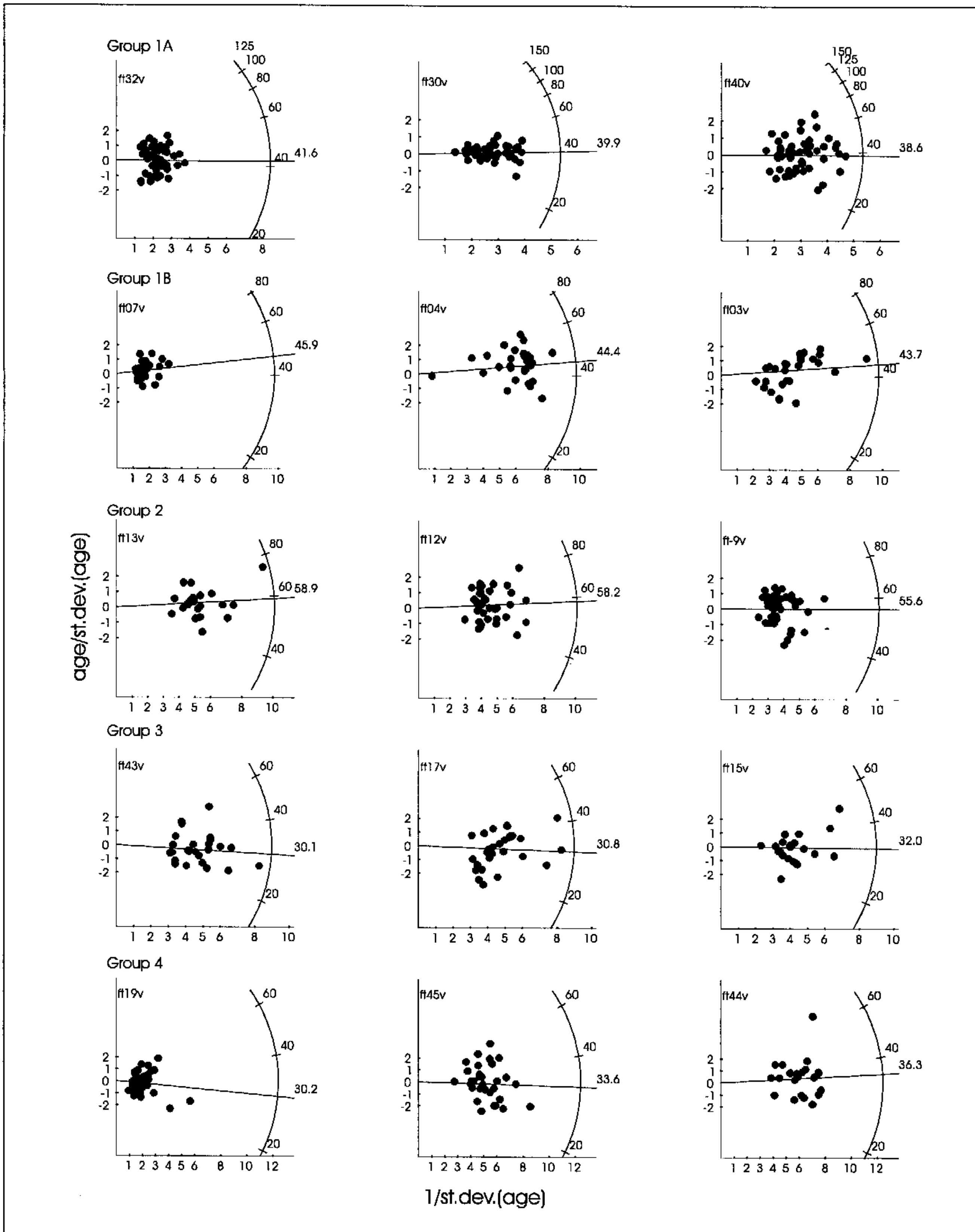


Fig. 4. Radial plots (Galbraith, 1988, 1990) of apatite FT apparent ages for representative samples in this study. Each dot represents one dated grain in the sample. Each estimate has unit standard error on the y scale. Points with larger x have a higher precision, reflecting higher U contents. The line represents the central FT age, which takes into consideration both the dispersion and the relative quality of each individual data point (Galbraith, 1990).

separation techniques (Grist and Ravenhurst, 1992) and were analyzed by the first author (VM) at the Dalhousie Fission Track Research Laboratory (FTRL) in Halifax, Nova Scotia, using the external detector method (Hurford and Carter, 1991). Thermal neutron irradiations for apatite were performed in the McMaster Nuclear Reactor, Hamilton, Ontario. Fission track ages were determined using the *zeta* calibration method as recommended by Hurford and Green (1983) and Hurford (1990). The "central FT age" was determined for all samples (Galbraith and Laslett 1993). An indication of the spread in FT age from individual apatite grains in each sample is given by the age dispersion (Table 2), with samples showing a statistically significant spread in individual grain age having age dispersions greater than ~10%. Between 16 and 57 clean and suitably oriented grains were counted for each individual sample. Dosimeter glass SRM614, together with Fish Canyon and Durango apatite age standards, were used to determine the *zeta* calibration factor for VM given in Table 2. Horizontal confined track lengths were measured using the procedure described in Gleadow et al. (1986) and also described for the Dalhousie FT lab in Ravenhurst and Donelick (1992). The standard error for track-length determinations indicated is the standard error of the mean. Although no microprobe data on apatite composition (F/Cl) have been acquired, on the basis of their etching characteristics (Ravenhurst and Donelick, 1992), we assume that the apatite samples are all fluorine-rich similar to the Durango standard used. Before rigorous modelling routines were commonly used, it was normal to report "length corrected" apatite fission track data (Gleadow et al., 1986; Green, 1988; Maksaev et al., 1988a), using a weighted mean track length of 14.9 μm for the Fish Canyon and Durango standard apatite used in (*zeta*) calibration. Modern modelling makes that correction unnecessary, thus only central ages are reported in this paper.

To model time-temperature histories, we used a FORTRAN algorithm developed by Sean Willett (Willett, 1992, 1997) and modified by Dale Issler (cf. Issler, 1996; Willett et al., 1997), which employs the annealing data of Laslett et al. (1987), and Crowley et al. (1991). The algorithm uses a constrained random search technique to generate a number of populations of fission tracks at discrete time intervals as a thermal history is evolved. Each population is annealed (shortened) in accordance with all the temperatures that are subsequently encountered, resulting in a single distribution of track lengths, that is compared to the measured distribution of lengths using a Kolmogorov-Smirnov (K-S) objective function goodness-of-fit statistic. The algorithm produces 250 such length distributions, which are maintained as the solution set. The 251st and all following solutions are compared to the main solution set and if a better fit to the data is obtained (a lower K-S statistic), the worst-fitting of the solutions in the main set is discarded. The algorithm continues in this way, producing sometimes several thousand thermal histories until all 250 solutions fit the measured data to a desired level of confidence (usually 2σ).

Sample locations and lithology are given in Table 1 and fission track results in Table 2, organized into four coherent groups to facilitate discussion: 1) Granitoids of the Domeyko Cordillera, 2) Cerros de Montecristo pluton, 3) Fortuna Granodiorite complex, and 4) Chuquicamata and El Abra porphyry copper deposits.

## Group 1: Granitoids of the Domeyko Cordillera

*Age Data*

Apatite samples from basement granitoid rocks of the Domeyko Cordillera are organized into a northern sub-group (1a) and a southern sub-group (1b) (Tables 1 and 2; Figs. 2 and 3). The northern group samples are in general considered to be Paleozoic in age, and exhibit biotite K-Ar ages ranging from 324 Ma to 202 Ma (Huete et al., 1977; Vega and Bordones, 1981; Boric et al., 1990; Maksaev, 1990). They gave FT ages from 45.2 Ma to 35.4 Ma (Quehuita, Cordon del Millo, and Cerros de Paqui in Tables 1 and 2). Two samples (FT-30 and FT-32) represent a diorite pluton with a K/Ar biotite age of 68 Ma ± 2.5 Ma (Vega and Bordones, 1981) that intrudes the Paleozoic basement in the Cerros de Paqui (Fig. 2), yet they behave like the others in terms of apatite fission tracks. One sample of pre-Tertiary granodiorite (FT-26) probably lies within the area of thermal influence of the Chuquicamata porphyry copper (Reynolds et al., 1998), and for that reason is not further considered in the following discussion.

In addition, comparable FT dates from 45.9 Ma to 43.6 Ma (Table 2) were obtained on apatite from a southern sub-group (1b) of Upper Paleozoic granitoids in the Salar de Pedernales area, east of the El Salvador and Potrerillos porphyry copper deposits (Figs. 2 and 3).

The radial plots (Fig. 4) for three representative samples of each group or sub-group show that, in general, samples portray single populations of apatite with variable contents of uranium, which translate in variable precision (standard deviation) in the age of each grain. Each age estimate has unit standard error on the y scale; points with larger x have a higher precision. The line in each diagram represents the "central age," which takes into consideration both the dispersion and the relative quality of each individual data point (see Galbraith, 1990).

Figure 5A illustrates that the FT ages of the samples from the northern (1a) sub-group (sample FT-26 excluded) reveal a regular variation with altitude, whereby ages generally decrease with decreasing altitude. Because of track-length patterns measured, we interpret this profile to represent the zone below the partial annealing zone (Fitzgerald et al., 1995). If the ages are interpreted to represent denudation cooling, this relationship roughly represents the time when the rocks now exposed at the surface intersected the ca. 100°C isotherm on their way to the sur-

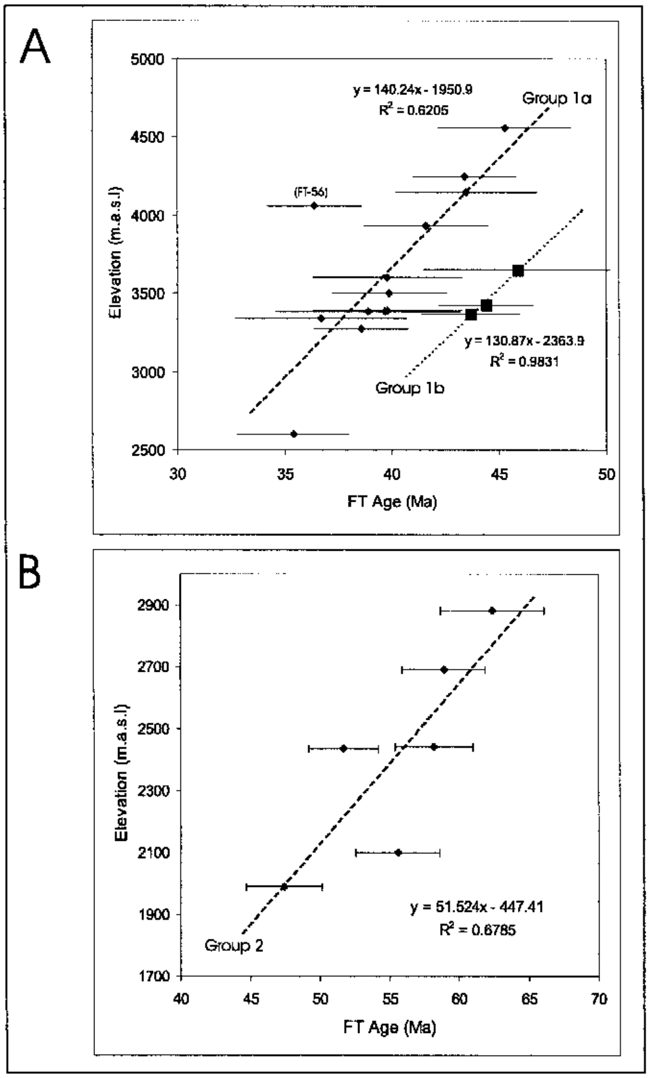


Fig. 5. Elevation above sea level of sample locations in the study area versus apatite FT ages (from Table 2). Basement granitoid rocks in the Domeyko Cordillera (5A) and Cerros de Montecristo pluton (5B).

face. A rate of exhumation cooling of ca. 140 m/My would be suggested by the data. Sample FT-56 is younger than expected for its elevation. This could be because FT-56 was collected west of the Domeyko Fault, whereas most samples are from east of that structure. However, sample FT-58 is also from west of the fault yet fits the general trend, hence this discrepancy remains unexplained. The FT ages of the southern sub-group from the Paleozoic crystalline basement in the Salar de Pedernales region also show a good correlation between elevation and age (1b, Fig. 5). Their exhumation would have taken place about 5 My earlier than in the northern region and the rate of exhumation cooling computed would be ca. 130 m/My. These rates are in the low end of denudation rates estimated by the same method in the European Alps (Wagner et al., 1977). However, Gallagher et al. (1998) argue convincingly that track-length distributions must be considered before these estimates can be considered reliable.

*Thermal Modelling*

If rocks remain at very low temperatures after their formation, fission tracks in apatite maintain their original track length of ca. 16 μm. This is the case for apatite from the Miocene (magnetite-apatite-bearing) lava flows of the El Laco volcano (5000 m), in which Maksaev (1990) measured tracks of 15 μm to 16 μm in length. In contrast, the measured mean etchable lengths of horizontal confined tracks in apatite from the basement rocks of Group 1 (Fig. 3) range from 12.9 μm to 14.7 μm, suggesting that the rocks have had protracted residence at higher temperatures at depth before cooling to surface temperatures. Only the basement apatite samples obtained at highest elevations (4250 m to 4560 m) have mean track-lengths longer than 14 μm, yet standard deviations are all more than 1 μm. Therefore, all granitoid basement samples fall under the range of confidence suggested by Green (1988), indicating that these apatites have significantly reduced track length and track density, and their apatite FT age must be younger than their actual geological age, which is the case. The track-length distributions in these samples are consistently unimodal, and negatively skewed similar to "undisturbed basement" rocks of Gleadow et al. (1986).

*Group 1a*

Results from time-temperature modelling of the data for the two samples with most abundant track length data (Table 2) are shown in Figure 6.

Sample FT-64 represents the Quehuita pluton (hornblende granite), and was collected at 4560 m above sea level (Table 1; Fig. 2). Sample FT-30 is from a diorite pre-Tertiary diorite body (68 ± 2.5 Ma K/Ar in biotite; Vega and Bordones, 1981) that intrudes a Paleozoic granitoid pluton in Cerros de Paqui (Fig. 2), collected at 3500 m (Table 1).

The modelling parameters are the same for both samples, which have been allowed "cooling only" since 100 Ma. Figure 6 shows both the model curve that best fits the measured track-length and age (thin line) and the exponential mean (or preferred) solution (heavy line) for 250 acceptable thermal history solutions that are each compatible with both the measured FT age and the observed track length distributions. The predicted FT age from each modelling iteration must be compatible with the measured age at 2σ uncertainties to be accepted. The cumulative density function calculated from the observed track length distribution and probability density function generated during each modelling iteration were compared using the Kolmogorov-Smirnov test at 0.05 probability (Willett, 1992, 1997) to determine the acceptable fit. The preferred thermal history solutions in Figure 6 are bounded by an upper and lower temperature limit to define an envelope through which the 250 acceptable solutions pass; these are not in themselves, solutions. Shown below each thermal history is a histogram that illustrates the distribution of times at which the onset of track retention began for the 250 acceptable thermal history solutions. Models were run for 100 Ma and permitted cooling at rates up to 10°C/Ma (Willett, 1992, 1997). Average times for the onset of track retention is ca. 49 Ma for both FT-64 and FT-30. These values are somewhat older than the measured FT ages of the samples (45.3 Ma and 39.9 Ma), and this fact reflects the significance of the short tracks in the distributions. It must be emphasized that there are no real constraints provided by the FT data at temperatures above those at which fission tracks were retained during cooling (ca. 125°C). The cooling histories above that temperature are not meaningful, and do not imply that cooling of the rocks began during the Late Cretaceous. Yet we can infer from argon dates on biotite and FT dates on zircon in Andriessen and Reutter (1994) that the rocks presently at surface were still at temperatures above ca. 250°C in the Triassic. The 68 Ma K/Ar

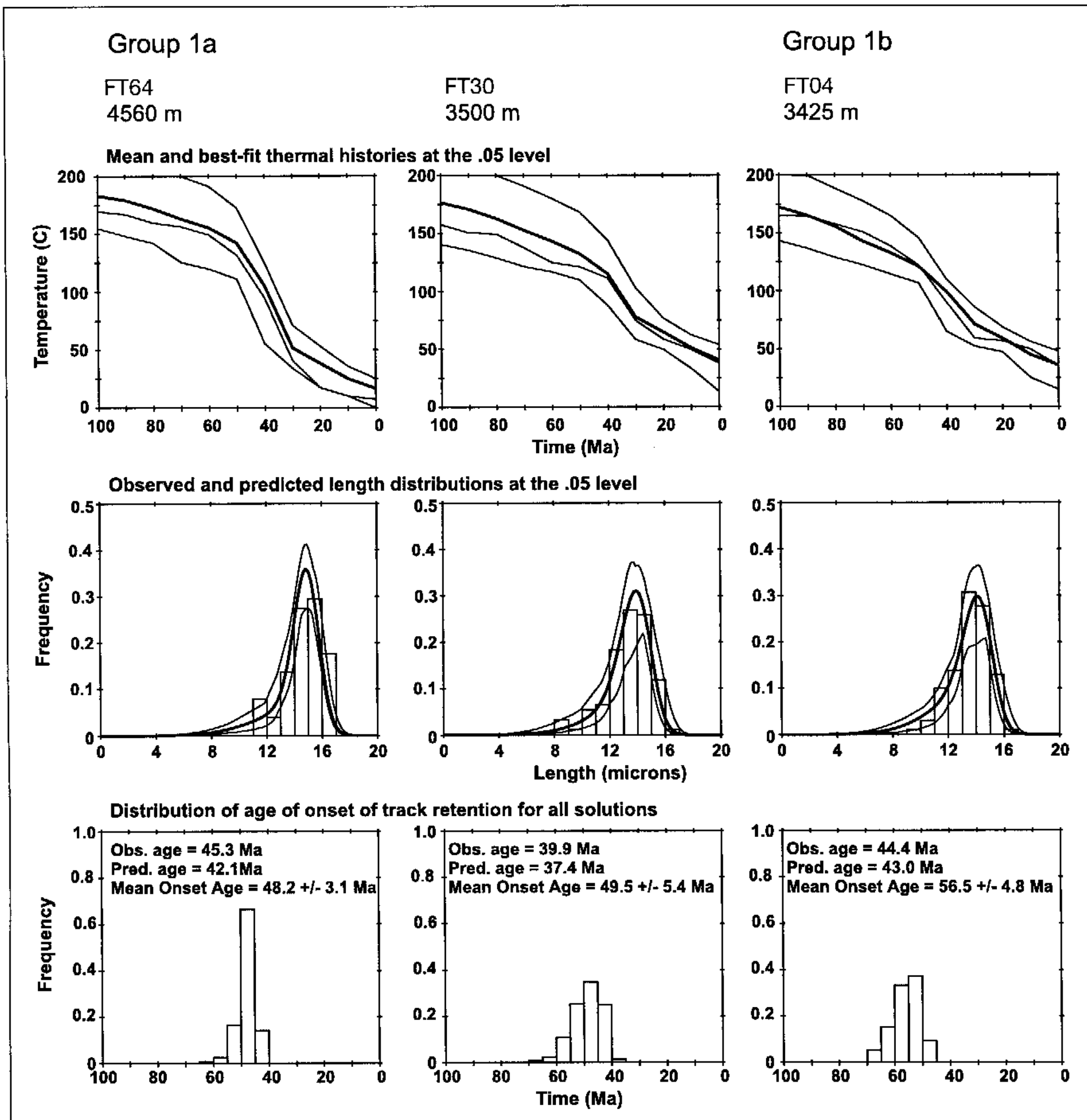


Fig. 6. Modelled thermal histories (cooling only model) of samples of granitoids from the Domeyko Cordillera from Figures 2 and 3 (Willett, 1992, 1997). Northern sub-group (1a: FT-64 and FT-30) and southern sub-group (1b: FT-04). The heavy line represents the probability distribution function for the exponential mean thermal history of 250 statistically acceptable solutions (preferred model), whereas the lighter line is the solution that best fits the track-length data. See text for discussion.

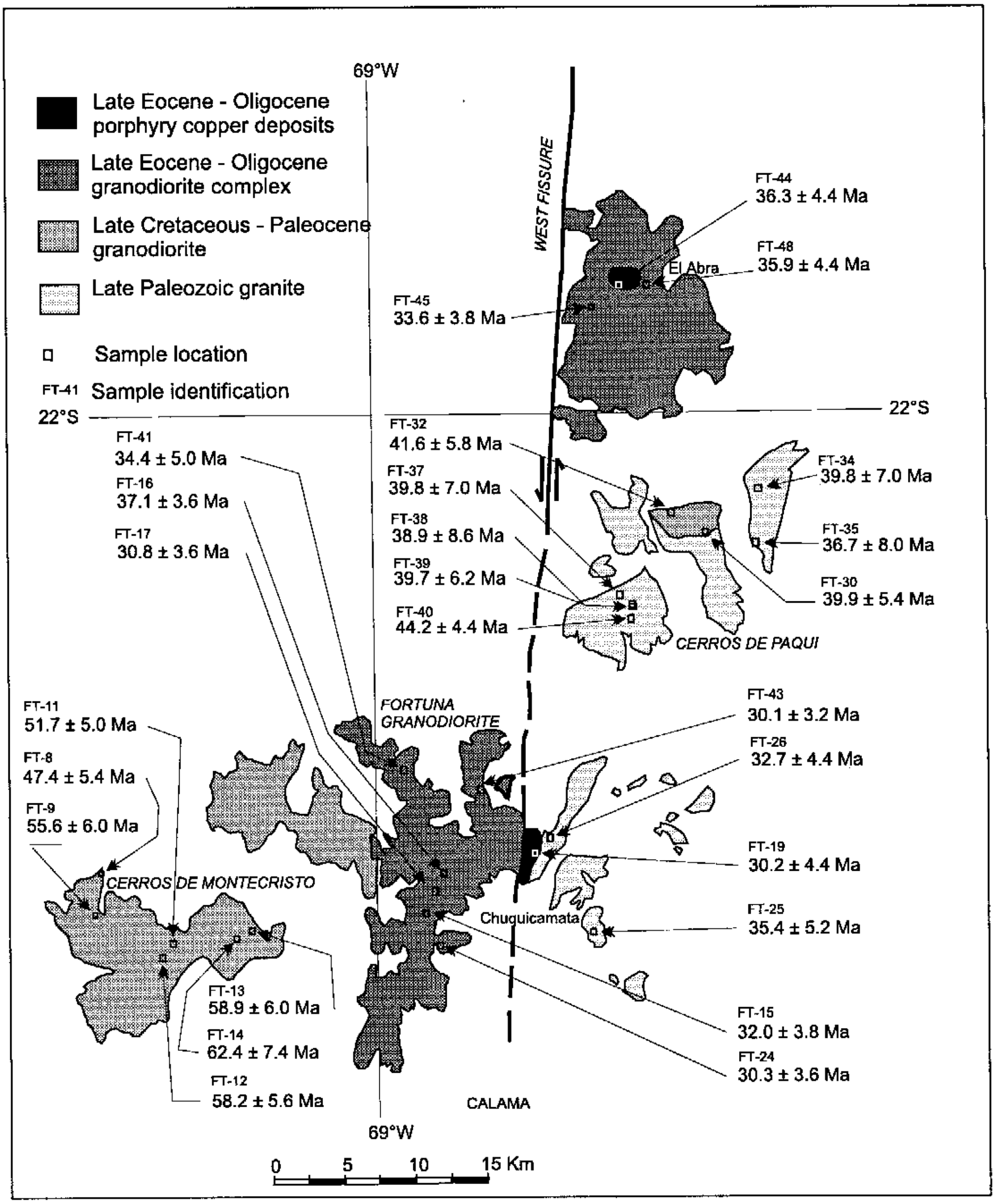


Fig. 7. Distribution of FT samples from the Chuquicamata-El Abra area (modified from Maksaev, 1990). A post-ore finite left-lateral displacement of ca. 35 km has been postulated for the strike-slip West Fault or West Fissure.

date on biotite for sample FT-30 (FT age of 40 Ma) suggests that the rocks now at 3500 m in Cerros de Paqui remained at a temperature above 125°C during the Cretaceous-Tertiary transition. We feel thus justified to consider that all the rocks in Group 1 were below the partial-annealing zone until the Eocene, and therefore the samples can be used to ascertain denudation rates (Fitzgerald et al., 1995; Gallagher et al., 1998).

Apatite FT data from the crystalline basement rocks place a Paleocene minimum timing for the initiation of denudation cooling in the Incaic event, with cooling continuing through to the Miocene. For the high-elevation sample FT-64, FT data are compatible with a faster rate of denudation cooling between 50 Ma and 30 Ma (ca. 170 to 140 m/My for a geothermal gradient of 30° or 25° per km, respectively), and then a slower rate of denudation (close to 50 m/My) between 30 Ma and 10 Ma, from which time the temperature has not varied significantly. However, due to the relatively high elevation of this sample in the range (4560 m), it could have resided at or close to the base of the partial annealing zone, and thus make the above rates less reliable.

The model for sample FT-30, taken 1000 m below the previous one, is compatible with the rocks having main-

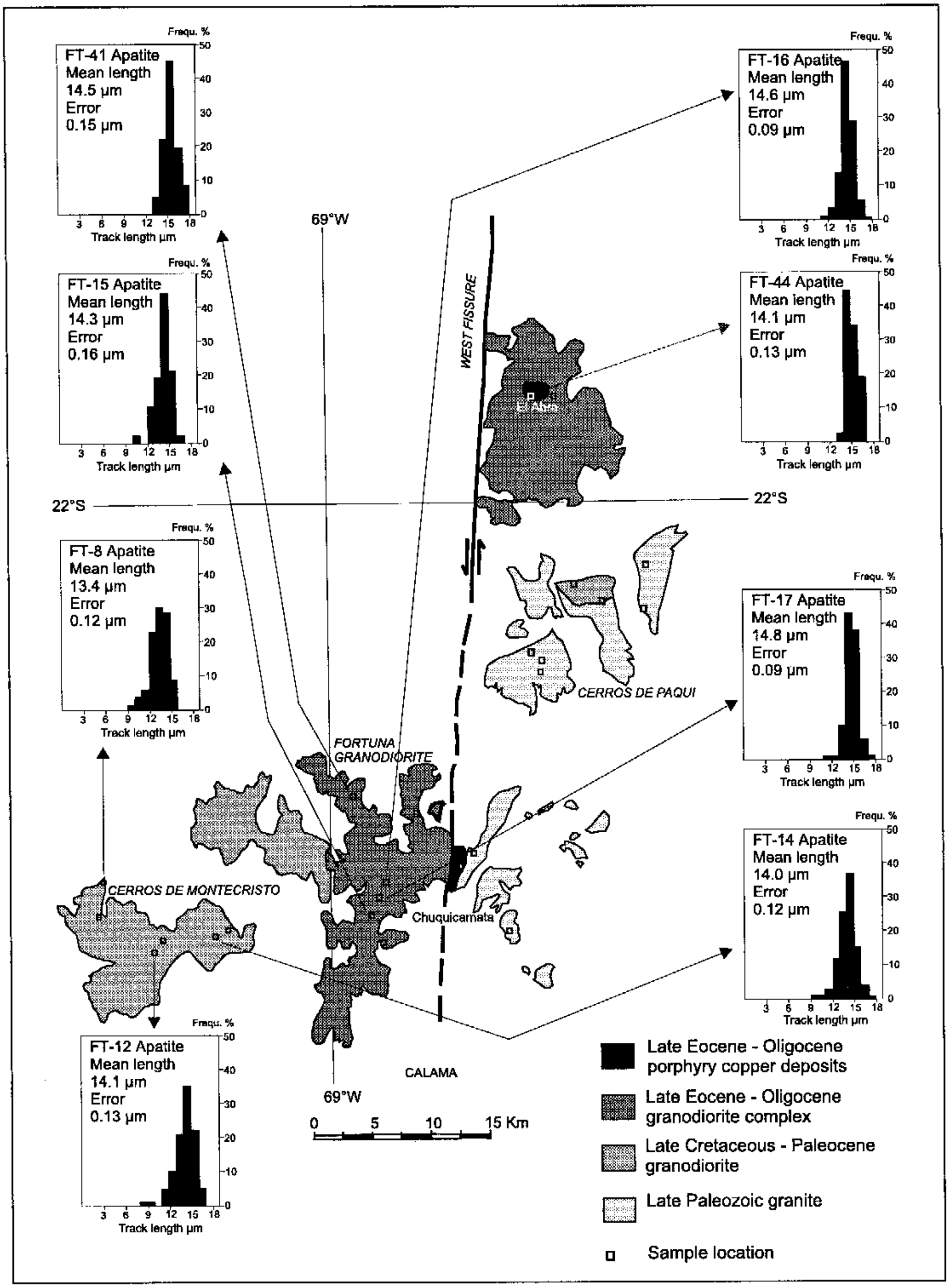


Fig. 8. Track-length histograms for apatite samples in the Chuquicamata-El Abra region (modified from Maksaev, 1990).

tained a higher temperature until about 45 Ma, then cooled rapidly (50°C in 15 My) to 30 Ma, when the rate of cooling decreased to the present day. The models for the other samples of Group 1 show a similar pattern of cooling. The track-length distributions suggest that these samples resided below the zone of partial annealing until the onset of the Incaic tectonic event. Thus a rate of denudation of ca. 110 m/My to 130 m/My may be suggested for these samples, assuming paleogeothermal gradients of 30°C and 25°C per km, respectively. These paleogeothermal gradients are speculative, but not unreasonable as a first approximation, since Giese (1994), on the basis of thermal measurements and calculations for the present-day upper crust of the Central Andes, estimates that a temperature of 141°C results at the depth of 5 km (average gradient of 28.2°C/km).

### *Group 1b*

Modelling for the crystalline basement of the Domeyko Cordillera in the more southern Pedernales region, east of Potrerillos and El Salvador (Fig. 2), yields a similar result (Sample FT-04, Fig. 6), but indicates that the onset of track retention occurred about 7 My earlier than in the Chuquicamata – El Abra region. This is compatible with FT ages of 44 Ma to 46 Ma obtained for that region. Track retention started at 70 Ma, cooling was fastest between 50 Ma and 30 Ma (equivalent to rates of ca. 110 to 130 m/My for a geothermal gradient of 30° or 25° per km, respectively).

## Group 2: Cerros de Montecristo Pluton

### *Age Data*

The Cerros de Montecristo granodioritic pluton (biotite $^{40}Ar$-$^{39}Ar$ plateau age of 63.1 ± 0.3 Ma; Maksaev, 1990), is located in the western foothills of the Domeyko Cordillera, 20 km west of Chuquicamata porphyry copper deposit (Figs. 7 and 8), and west of the Domeyko Fault System (locally West Fissure or West Fault). The apatite FT age of the sample (FT-14) collected at highest altitude (2880 m) yielded an age of 62.4 ± 7.4 Ma that is concordant with the biotite $^{40}Ar$-$^{39}Ar$ age of 63.1 ± 0.3 Ma (Maksaev, 1990). The other five

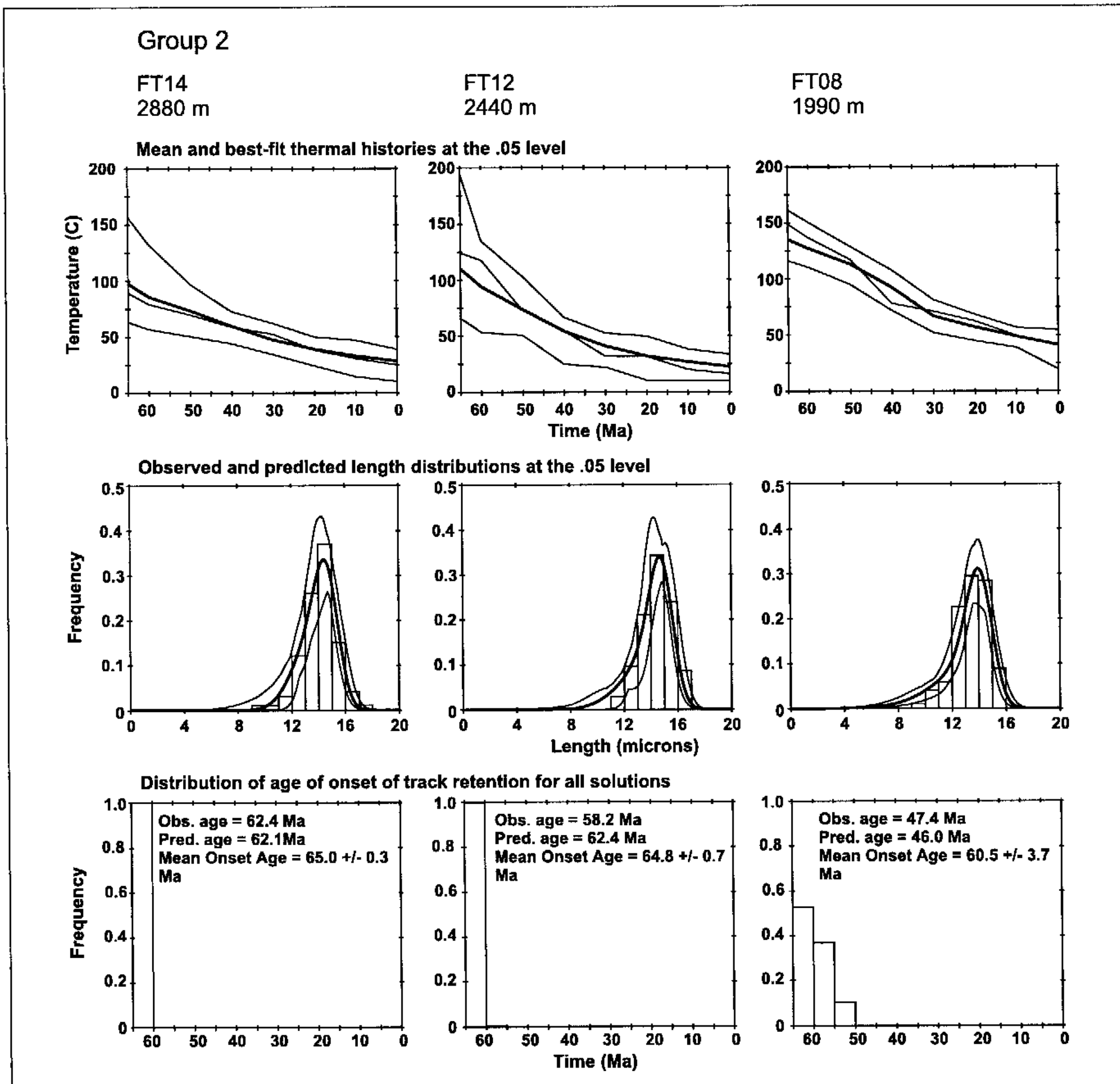


Fig. 9. Modelled thermal histories (cooling only model, Willett, 1992, 1997) of samples of granitoids from the Cerros de Montecristo. See text for discussion.

samples collected from this pluton show ages decreasing steadily with altitude down to 47.4 Ma at 1990 m (Fig. 5B). The concordance of the biotite $^{40}Ar$-$^{39}Ar$ and the apatite FT dates is taken to suggest that the sample from the highest elevation of the Cerros de Montecristo pluton has remained at shallow depth since the pluton cooled in the Paleocene, hence there has been no significant FT age reduction. The younger ages at lower altitudes (down to 1990 m of altitude) represent a progressively longer permanence of the rocks at slightly higher temperatures (deeper levels), but probably still below the closure temperature of fission tracks in apatite (partial annealing zone).

The measured mean confined track lengths in apatite from the Cerros de Montecristo Pluton range between 13.4 μm and 14.4 μm, with standard deviations from 0.99 μm to 1.37 μm (Table 4; Fig. 8). The track length distributions are narrow and unimodal, and except for the sample at lowest altitude, similar to that of apatite from "undisturbed volcanic" rocks of Gleadow et al. (1986). The preservation of the original cooling age of the pluton in the apatites from the sample collected at highest altitude of this Paleocene pluton is taken to indicate that the denudation at the top of the Cerros de Montecristo mountains has probably progressed less than about 1.5 km, and the total exhumation of the Cerros de Montecristo Pluton would be ca. 2.5 km since about 63 Ma (assuming a 30°C per km paleogeothermal gradient). Figure 5B shows the FT age versus elevation profile for the Cerros de Montecristo samples would suggest an overall very low rate of cooling/exhumation of this granodioritic pluton, less than half of that estimated for the Domeyko Cordillera just 15 km to the east, but this is probably an exhumed partial-annealing zone.

### *Thermal Modelling*

Figure 9 shows three models for samples FT-14, FT-12, and FT-8 taken at 2880 m, 2440 m, and 1990 m elevation, respectively. Concordant with the estimates from FT age versus elevation (Fig. 5B), the FT length modelling suggests the pluton cooled immediately after its emplacement at ca. 63 Ma, compatible with shallow emplacement and slow exhumation cooling at a rate of ca. 50 m/My, depending on the paleogeothermal gradient. Significantly, the shallowest sample indicates it maintained a higher temperature for longer, for instance, whereas the highest elevation sample reached 75°C about 55 Ma, the shallowest sample attained the same temperature only at 35 Ma.

Despite the uncertainty concerning the paleogeothermal gradient, the above demonstrates that the individual younger FT ages in this group simply represent longer residence in a temperature regime where fission tracks steadily shortened and partially disappeared. Significantly, there is no correlation between the younger FT ages and the proximity to the Eocene-Oligocene porphyries in the east. The data indicate a significantly different cooling history in the Sierra de Montecristo compared to the Domeyko Cordillera east of the Domeyko Fault system.

## Group 3: The Fortuna Granodiorite Complex

### *Age Data*

The Fortuna Granodiorite complex located immediately west of Chuquicamata (Figs. 7 and 8) gave an apatite FT age of 37.1 ± 3.6 Ma for the sample collected at highest altitude (3270 m). This age is concordant with a biotite $^{40}Ar$-$^{39}Ar$ age of 37.1 ± 0.5 and K-Ar dates of the same intrusive complex (Maksaev, 1990). The other five samples from the Fortuna Granodiorite complex collected at lower altitudes yielded younger ages that, although not consistently, show a decrease in age with decreasing elevation (Table 2). This may suggest that these apatites were partially annealed like those of Group 2. For some reason, their FT ages are also comparable to the various pulses of hydrothermal alteration and mineralization in the Chuquicamata porphyry copper deposit, the last of which took place at ca. 31, suggesting a possible resetting of FT ages. And yet, there is evidence that the West fault has had ca. 35 km of left-lateral displacement since the Oligocene (Dilles et al., 1997; Tomlinson and Blanco 1997a, 1997b), hence Fortuna should not have been near the Chuquicamata hydrothermal system until very recently. It is possible that the younging effect is related to proximity to the West Fault, which has been thermally active until the Pliocene (apatite FT age of ca 4.5 Ma in the MM mine; Zentilli et al., 1994a, 1994c), and electron spin resonance (ESR) in quartz indicate reactivation of the immediate fault zone into the Pleistocene (Lindsay et al., 1997).

### *Thermal Modelling*

The measured confined mean track lengths of the apatite from the Fortuna Granodiorite range from 14.3 μm to 14.8 μm, with errors from 0.09 μm to 0.16 μm. The distribution of track lengths is unimodal and very narrow (Fig. 8), similar to those of apatite standards or "undisturbed volcanic rocks" of Gleadow et al. (1986).

The model thermal evolution of samples FT-41 and FT-17 (Fig. 10) suggests a very rapid initial cooling followed by residence at low temperatures since the Miocene. Using the same assumptions of paleogeothermal gradients (above), as a first approximation this block has been exhumed at a rate between 200 and 270 m/My in the Eocene-Oligocene, quite differently from the Cerros de Montecristo to the west (Group 2) and the crystalline basement to the east (Group 1). Yet exhumation has not been more than 2 km to 4 km.

## Group 4: Chuquicamata and El Abra Porphyry Copper Deposits

### *Age Data*

An apatite sample (FT-19; Tables 1 and 2, Fig.7) from the Chuquicamata porphyry copper deposit, obtained from the mine pit some 400 m below the natural (pre-mine) surface, gave a FT age of 30.2 ± 4.4 Ma. This date is concordant with a K-feldspar $^{40}$Ar-$^{39}$Ar plateau age of 31.4 ± 0.2 Ma, and also with a biotite $^{40}$Ar-$^{39}$Ar plateau age of 31.7 ± 0.4 Ma obtained from the same sample (Maksaev, 1990). The concordance of these three dates, from minerals with significantly different closure temperatures, indicates fast cooling of the porphyry copper, at least within the range from about <300°C to 100°C. Not sufficient track lengths were measured to warrant modelling. The data for FT-19 compare well with apatite FT ages at Chuquicamata by Barry Kohn reported in McInnes et al. (1999): 28.0 ± 2.7, 29.7 ± 2.2, 32.0 ± 2.8, 33.3 ± 2.4 Ma (average 30.75 Ma).

A mineralized dacitic porphyry from El Abra porphyry copper deposit (Ambrus, 1977) yielded an apatite FT age of 36.3 ± 4.4 Ma (FT-44; Tables 1 and 2) that is concordant with the 36.8 Ma $^{40}$Ar-$^{39}$Ar higher step (irregular $^{40}$Ar-$^{39}$Ar spectra with a gradient of ages) of the biotite from the same sample (Maksaev, 1990).

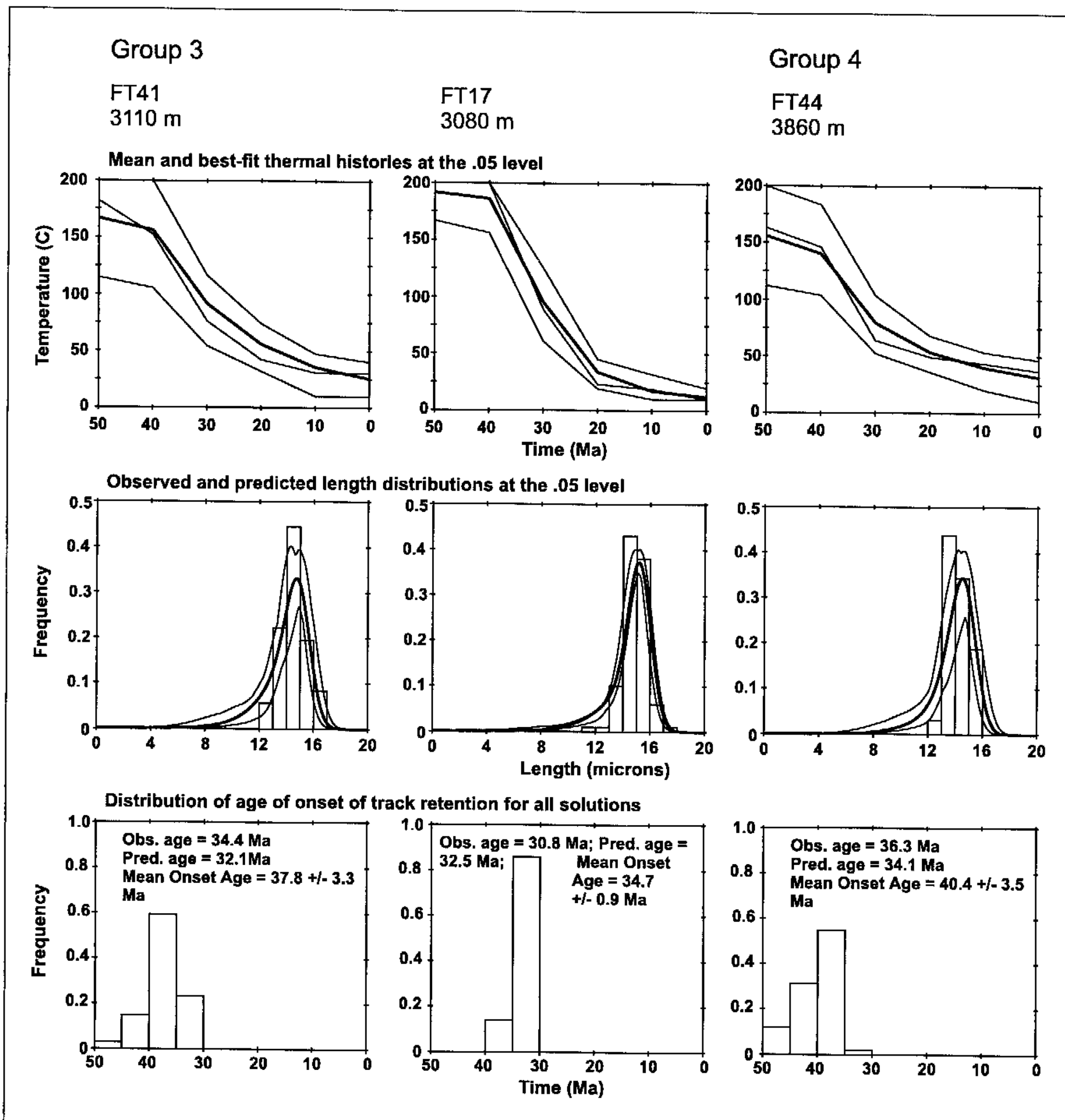


Fig. 10. Modelled thermal histories (cooling only model, Willett, 1992, 1997) of samples of granitoids from the Fortuna Granodiorite Complex (Group 3) and El Abra (Group 4). See text for discussion.

Apatite samples from the unmineralized Southern Granodiorite of El Abra yielded ages of 33.6 ± 3.8 and 35.9 ± 4.4 Ma that are also concordant with the biotite $^{40}Ar$-$^{39}Ar$ plateau age of 36.7 ± 0.8 Ma obtained for this intrusion (Maksaev, 1990). These FT ages are consistent with the fast cooling of El Abra porphyry copper deposit below about 300°C, and its residence since about 36 Ma at temperatures under about 50°C.

*Thermal Modelling*

Only sample FT-44 of this group had sufficient track lengths measured to allow for modelling. The confined track length of the apatite separate from this sample is 14.1 μm with an error of 0.13 μm (Table 2). The distribution is narrow and unimodal (Fig. 8) comparable to "undisturbed volcanic" rocks of Gleadow et al. (1986).

The model (Group 4 in Fig. 10) indicates that the mean age of onset of track retention was 40.4 ± 3.5 Ma, well within error but somewhat older than the 36.8 Ma (biotite $^{40}Ar/^{39}Ar$) deduced by Maksaev (1990), suggesting that the pluton may be older and the argon age is a cooling age. The discrepancy could also result from slight changes in apatite geochemistry, which has not been tested. Nevertheless, the rocks would have cooled at a rate of ca. 50°C between 40 Ma and 30 Ma, which in terms of denudation cooling could represent 200 to 170 m/My. After 30 Ma, cooling has been much slower, possibly reflecting denudation rates of less than 50 m/My. As such, FT-44 has a cooling history quite similar to that of the highest elevation Fortuna sample (FT-64; Group 3).

## Discussion

The purpose of this study was to better understand the spatial and temporal association of giant porphyry copper formation with the tectonic uplift of the Domeyko Cordillera. The FT data presented above indicate that the region as a whole experienced exhumation cooling in the Eocene-Oligocene, concurrent with formation of the porphyries between 40 Ma and 31 Ma. However, the study detects clear differences between different blocks represented by the various groups of samples within the Domeyko Cordillera domain. Even if the absolute values of denudation rates computed here are somewhat speculative, the treatment and the data are internally consistent, and allow for comparison.

The crystalline basement rocks presently at the surface in the Domeyko Cordillera, represented by Group 1 (Fig. 2), cooled in the late Paleozoic to Triassic to temperatures cooler than the retention temperature for argon in K-bearing minerals (ca. 300°C to 200°C; Reynolds et al., 1998) and FT in zircon (ca. 350°C, Gallagher et al., 1998). Late Cretaceous diorite intrusions in the Paleozoic basement, such as in the region of Cerros de Paqui (68 Ma K/Ar in biotite, blocking temperature ca. 300°C) had no regional overprinting effect. Hence, at the end of the Cretaceous, rocks now at the surface were temperatures of more than ca. 125°C (total annealing of FT in apatite), but less than ca. 350°C, representing a depth range of between 12 km and 4 km for an average paleogeothermal gradient of 30° km. Our FT data for apatite in these rocks indicates that they resided in the total annealing zone for apatite, and that significant exhumation took place between 50 Ma and 30 Ma. Speculative exhumation rates of ca. 170 to 140 m/My can be suggested. Since then, exhumation has continued at much reduced rate, ca. 50 m/My.

The data presented above are comparable to those of Andriessen and Reutter (1994) who obtained an apatite FT age of 50.8 ± 7.6 Ma (mean track length 12.74 ± 2.76 μm) for a late Paleozoic granitic pluton of the Domeyko Cordillera at 23°30' Lat. S (Fig. 2; CHI-6, midway between groups 1a and 1b). This pluton yielded a biotite K-Ar age of 279 ± 11 Ma and a zircon FT age of 213 ± 34 Ma. Andriessen and Reutter (1994) agreed with Maksaev (1990) in interpreting the FT data to indicate denudation cooling of the late Paleozoic granite during the Eocene as a result of the Incaic tectonic pulse. However, it must be noted that the track-length distribution histogram for CHI-6 in Andriessen and Reutter (1994) is bimodal, suggesting that the rocks resided near the base of the partial annealing zone rather than the total annealing zone, denoting less overall depth of erosion than the samples in the present study. Alternatively, the rocks would have a more complex thermal history, which could only be tested by modelling of the raw data.

In contrast to Group 1, rocks in the Sierra de Montecristo (Group 2) west of the highest ridges of the Domeyko Cordillera (Fig. 7) show no evidence of the post-30 Ma break in cooling rates. The pluton was shallowly intruded in the Paleocene, cooled very rapidly, and continued to cool at a slow monotonous rate, equivalent to an exhumation rate of ca. 50 m/My for a constant paleogeothermal gradient. This contrast confirms that Group 1 rocks experienced a time-temperature history that was relatively localized in a longitudinal, narrow belt.

Furthermore, the time-temperature history experienced by the Fortuna complex (Group 3, Fig. 7), adjacent to Sierra de Montecristo, was not shared either, and there is no evidence in these apatites of a thermal effect by the intrusion (merely 10 km away) of the Fortuna plutons at ca. 40 Ma.

The Fortuna complex data provide interesting insights into the important controversy concerning the finite (post-ore) left-lateral displacement of the strike-slip West Fault or West Fissure (Fig. 7), which has lead to tireless exploration for the "missing half" of the Chuquicamata deposit. Ambrus (1979) gives credit to previous workers (A.V. Taylor, A. Thomas) for proposing a correlation between the intrusives of Fortuna (Group 3) and El Abra (Group 4), implying a finite displacement of ca. 40 km. Other authors have estimated the left-lateral displacement to be only 25 km (Reutter et al., 1996), but recent field mapping indicates 35 km (Dilles et al., 1997; Tomlinson and Blanco, 1997a, 1997b). In contrast, Sillitoe (1973) proposed that the Fortuna intrusives represent the uplifted deeper roots of the Chuquicamata porphyry system, and some workers still question the field evidence for 35 km of left-lateral displacement.

Late in the preparation of this manuscript we learned of new data of apatite (U-Th)/He and FT for the Fortuna complex (McInnes et al., 1999). Apatite (U-Th)/He ages range from 33.6 to 17.1 Ma, reflecting the low closure temperature of the (U-Th)/He method (ca. 75°C for cooling rates of ca. 10°C/My). The (U-Th)/He ages for Chuquicamata are indistinguishable from the FT dates ((U-Th)/He: 30.9 Ma versus FT: 30.8 Ma). Their novel data add credence to the shallow emplacement, and rapid cooling rates of the Chuquicamata hydrothermal system after ca. 35 Ma (Maksaev, 1990). We totally disagree with McInnes et al. (1999) on their tectonic conclusions on the relationship of the Fortuna complex and Chuquicamata, which are not supported by field evidence.

The time-temperature FT modelling data for the El Abra (FT44) sample is quite similar to the high-elevation Fortuna sample (FT41), suggesting that both groups have had a similar exhumation cooling history, compatible with the recent geological mapping and correlation (Dilles et al. (1997), with which we agree.

The major porphyry copper deposits of northern Chile were formed during the Late Eocene to Early Oligocene (41 Ma to 31 Ma). Fission track analyses from Chuquicamata and El Abra (Group 4) are compatible with the relatively shallow emplacement of the porphyry systems (ca. 2 km to 3 km), followed by rapid cooling and exhumation. Such relatively shallow porphyry mineralization and fast cooling is consistent with the conceptual models developed for this type of deposit (Lowell and Guilbert, 1970; Burnham, 1997; Candela and Blevin, 1995; Hedenquist, 1995). However, many of these deposits still preserve extensive zones of supergene enrichment and exotic deposits (Mortimer et al., 1977, 1978), which in many cases add significantly to their economic feasibility (Ossandon and Zentilli, 1997). This supergene enrichment took place after the end of hypogene alteration mineralization (ca. 31 Ma, Zentilli et al., 1994a; Reynolds et al., 1998), in the Miocene (19 Ma to 15 Ma, Sillitoe and McKee, 1996). Erosion was reduced to a minimum during the Miocene, when hyper-arid conditions prevailed in the region (Alpers and Brimhall, 1988). The FT modelling presented reflects this drastic reduction in denudation rates since the Miocene.

The movement across the Domeyko Fault System has inverted to become left-lateral, and its timing is constrained between the age of quartz-sericitic alteration at Chuquicamata (31 Ma; Lindsay, 1998; Reynolds et al., 1998) and the development of transtensional basins with interbedded 25 Ma tuffs north of Chuquicamata; its finite left-lateral displacement is accepted to be ca. 35 ± 1 km (Dilles et al. 1997; Tomlinson and Blanco, 1997a, 1997b).

## The Model

The present high (5000 m) elevation of the Domeyko Cordillera has been generally considered to have been attained in the Miocene, during the Miocene Quechua compression, through thrusting that resulted in crustal thickening and shortening (Mégard, 1987; Maksaev, 1979, 1990). Mpodozis and Ramos (1989) infer that the Domeyko Cordillera is limited in the west by west-verging thrusts and in the east by east-verging reverse faults. In a discussion of fore-arc tectonics of the region, Buddin et al. (1993) clearly show the Domeyko Cordillera bound by west- and east-verging thrusts, but as most authors do, they imply that the deformation was Oligocene or younger. Could this structure have started to develop earlier, in the Eocene-Oligocene, when the magmatic arc was axial to the Domeyko Cordillera?

Most relevant is the definition by Tomlinson et al. (1996) of the Challo Fault in the Chitigua Cuadrangle (Maksaev, 1978), a west-verging reverse structure that bounds the western margin of the Cordon de Millo-Sierra del Medio Paleozoic block, with a NS extension of over 110 km. Field relationships mapped by Tomlinson et al. (1996) indicate that the Challo Fault is an Eocene structure that has been locally reactivated by the Domeyko Fault system in the Oligocene. Hence, the Cordon de Millo and Quchuita samples of Group 1 lie in the hangingwall of the Eocene Challo reverse fault, which must have had several kilometers of vertical displacement, dated for the first time by the FT data presented here.

Another piece of evidence comes from the Cordillera de Domeyko south extension into the area of El Salvador and Potrerillos, west of Salar de Pedernales (Fig. 2). Perelló and Müller (1984) described NS reverse faults (Sierra Castillo in the west and Barrancas in the east). These faults, marked by mylonites, bound an uplifted horst of Paleozoic crystalline basement rocks of the Sierra Castillo Batholith. These authors proposed the faults, which would have 3000 m to 3500 m vertical displacement, were active after the Late Cretaceous, whose rocks are faulted, and before the deposition of Miocene gravels, which are not. We suggest that these reverse faults are equivalent to the Challo Fault (above): they were compressive, reverse structures in the Eocene, and were probably later reactivated as strike-slip faults since the Oligocene (Tomlinson et al., 1994).

In summary, the FT data necessitate that considerable exhumation took place between as early as ca. 50 Ma, most definitely between 40 Ma and 30 Ma, when as much as 4 km to 5 km of the crust was eroded away, and if erosion was not keeping up with uplift of the block (England and Molnar, 1990), considerable tectonic (surface) uplift may have resulted. A regional unconformity confirms this extensive pre-Oligocene exhumation (Maksaev, 1990). The FT modelling presented here is compelling evidence that the Incaic compressive event started in the early Eocene.

In terms of plate interaction, 50 to 40 Ma to 30 Ma coincides with a time of increased convergence rates and oblique incidence between the Nazca and South American Plates (Fig. 1; Pilger, 1984; Pardo-Casas and Molnar, 1987; Maksaev, 1990). The porphyry copper deposits of the Central Andes developed within a large active right-lateral strike slip system during this period (Maksaev and Zentilli, 1988; Maksaev 1990). Lindsay (1998) demonstrated that vein arrays and mineralized stockworks at Chuquicamata developed within the active (first ductile, during porphyry emplacement, and later brittle) shear domain of the extensive Domeyko Fault system. Other porphyry copper deposits do not occur within the present expression of the faults, but normally within a short distance of them (Fig. 1). Maksaev (1990) and Zentilli et al. (1994a, 1994b, 1995) have argued that isotopic and geochemical data require that the magmas responsible for the genesis of the giant porphyries were generated at a greater depth, and their ascent resulted in less crustal contamination, than earlier and later magmas in the same regions. The depth of magma generation was based on extremely steep REE patterns, suggesting magmas were generated in a zone of (residual) garnet stability. The lesser contamination of the magmas was based on the observation by Zentilli et al. (1988, 1994b, 1995), Maksaev (1990) that Sr, Nd and Pb isotopes in Chuquicamata were compatible with a mantle source with less contamination by the upper crust than pre-porphyry igneous rocks in the region. The porphyry intrusions have initial Sr values of ca. 0.704 and slightly positive Nd isotopic values (+1 to +3) indicative of derivation from an enriched mantle source. Pb compositions, which are identical to those of the Miocene and Pliocene porphyry copper deposits of Central Chile, (Zentilli et al., 1988) were interpreted to be coming from an enriched mantle source. However, the presence of inherited zircons of Paleozoic age in the porphyry suites are evidence of some crustal interaction (Zentilli et al., 1994b). Tosdal (1995) and Cornejo et al., (1997) suggested that these magmas were probably derived from or contaminated by late Paleozoic rocks residing in the Lower crust. It is possible that a similar deep, isotopically homogeneous source rock has been tapped in both northern and central Chile porphyries following crustal thickening events.

An effective mechanism to avoid contamination is rapid ascent of magmas through pre-contaminated conduits to upper crustal levels. The development of tall magma columns with high-aspect cupolas under these circumstances would enhance enrichment in volatiles and metals (Candela and Blevin, 1995). Lindsay (1998) proposed such a process for Chuquicamata, based on his field work and a review of studies on syn-kinematic emplacement of plutons within extensional domains in strike-slip fault zones (Guineberteau et al., 1987; D'Lemos et al., 1992).

Oblique convergence in the Eocene-Oligocene may have partitioned into an arc-normal component, which enhanced crustal shortening and thickening by reverse faults

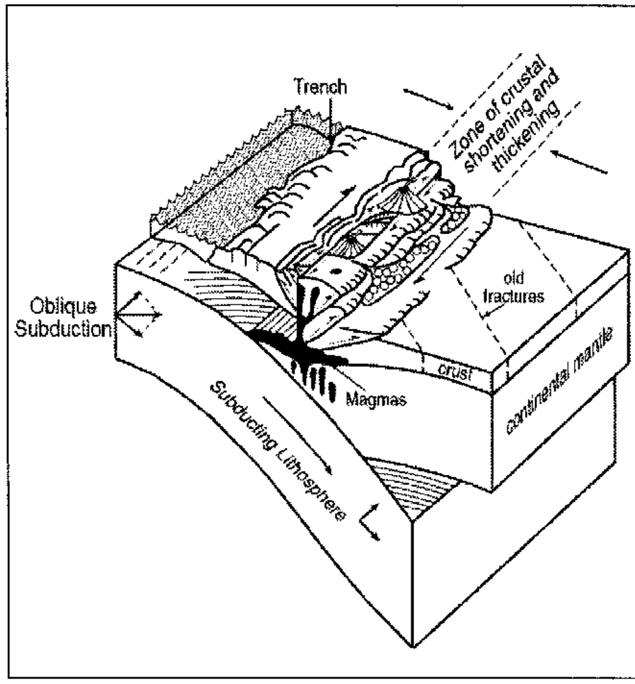


Fig. 11. Cartoon depicting a proposed model for emplacement of giant porphyry copper deposits in the Chilean Central Andes in a belt of crustal shortening and thickening along the 41 Ma to 30 Ma magmatic front. Oblique subduction and accelerated rates of incidence (Pardo-Casas and Molnar, 1987) leads to intra-arc transpression partitioned into thrusts of opposing vergence and a right-lateral shear system (the Domeyko Shear System). Porphyry magmas are allowed to rise rapidly in localized extensional domains within the deeply rooted shear system. This model borrows from Maksaev and Zentilli (1988), Maksaev (1990), Lindsay (1997), Saint Blanquat et al. (1998), Cembrano (1998), Braun and Beaumont (1995), D'Lemos et al. (1992) and references therein. Not to scale.

(e.g., Challo Fault), and an arc-parallel component, which concentrated shear stresses within the continental plate, leading to the development of the right-parallel Domeyko Fault System. The shear stresses would have localized in the thermally weakened lithosphere coinciding with the magmatic front (Woodcock, 1986). Transtensional regimes may lead to the emplacement of plutons in the upper crust (Grocott and Wilson, 1997). The kinematics of deformation at plate boundaries is primarily controlled by the direction and magnitude of the relative slip vector with respect to the trend of the plate boundary zone (McKenzie and Jackson, 1986; Tikoff and Teyssier, 1994; Braun and Beaumont, 1995; Saint Blanquat et al., 1998; Burbridge and Braun, 1998). Cembrano (1998) has proposed a tectonic model to explain complex structural relationships and crustal thickening and shortening in the active Liquiñe-Ofqui transpressive fault zone in the southern Andes (Cembrano et al., 1996), which exerts strong controls on Quaternary volcanism (Cembrano and Moreno, 1994). It is fair to indicate, however, that in the area of Potrerillos-El Salvador, Tomlinson et al. (1994) have documented Eocene left-lateral displacements in structures that they consider equivalent to the Domeyko Fault system further north where Eocene right-lateral displacements predominate, hence the system may be more longitudinally complex than envisaged here.

Figure 11, which borrows from all the above studies, summarizes schematically a speculative model for the localization of porphyry copper deposits in the Central Andes of Chile during the Eocene-Oligocene metallogenetic event. Magmas pool in the lower crust unless permitted to rise relatively rapidly through localized extensional domains in the ductile right-lateral, strike-slip Domeyko Fault System. Overlying crust undergoes shortening and plate-boundary-parallel shearing, concentrated above the magmatic front, overlying where near-rigid lithospheric mantles converge and the oceanic plate subducts. The path of the magmas to the upper crust may be diverted by brittle structural inhomogeneities, such as rejuvenated old crustal faults (Salfity, 1985; Mpodozis et al., 1994). Although porphyry systems developed in the upper few kilometers of the brittle crust, some of the felsic magmas were extruded explosively, as demonstrated by volcanic tuffs of ca. 38 Ma age (Hammerschmidt et al., 1992), thus coeval with porphyry emplacement. The model accounts for higher exhumation rates in the immediate area of the magmatic front, thus none of the volcanic edifices were preserved, and for the formation of clastic accumulations in the adjacent basins.

This model may also have application to the belt of Miocene to Pliocene giant porphyry copper deposits in central Chile (e.g., Los Pelambres, Los Bronces-Rio Blanco, El Teniente). There too, crustal thickening and shortening preceded and accompanied porphyry emplacement (Skewes and Holmgren, 1993; Skewes and Stern, 1995, Kurtz et al., 1997). Rates of convergence were high at the time (Pardo-Casas and Molnar, 1987) and also right-oblique, although at a larger angle than during the Eocene-Oligocene, hence less likely to develop large arc-parallel shear systems. Crustal shortening was dominated by thrusting (Almendinger et al., 1990), and, Skewes and Holmgren (1993), on the basis of fluid inclusion studies, estimated denudation rates of 260 to 150 m/My during the last 11 Ma. No major strike-slip faults have been identified so far on the surface. However, as pointed out by Lindsay (1998), arc volcanoes such as Mount St. Helens in western North America tend to occur in tensile zones above deep, seismically active fault zones that are oriented to accommodate strike slip displacements within the magmatic arc (Weaver and Hill, 1978; Weaver et al., 1987). Pre-existing fault systems that also control the distribution of Quaternary volcanism in the southern Andes (Cembrano and Moreno, 1994) must have played an important role in localizing porphyry intrusions.

## Conclusions

Apatite FT thermochronology has demonstrated its effectiveness as a tool in evaluating times and rates of denudation-cooling in the Domeyko Cordillera of the Chilean Central Andes. The study shows that different tectonic blocks within the region experienced contrasting time-temperature histories, indicating several kilometers of rock were eroded from the whole region in the Eocene. Rates of exhumation of 200 to 100 m/My prevailed between ca. 50 Ma and 30 Ma, whereas since then, the rate has been ca. 50 m/My, probably reflecting hyper-arid conditions. The lack of erosion during sustained crustal thickening since the Eocene has resulted in net surface uplift of the Domeyko Cordillera to its present elevations of ca. 5000 m. The FT data are compatible with a model whereby high convergence rates between the Nazca and South American Plates in the Eocene-Oligocene lead to crustal thickening and shortening, and to the deepening of the zone of magma generation. The right-oblique convergence lead to the development of a trench-parallel shear zone along the weakened lithosphere along the magmatic front. Magmas rose rapidly and with minimal contamination in localized transtensional domains in the Domeyko Shear zone, to be emplaced in the near surface.

Porphyry copper deposits were emplaced during this period of crustal thickening and active exhumation, between 41 Ma and 31 Ma. The porphyry systems, emplaced at a few kilometers of the surface into the active shear system, cooled rapidly and were progressively exhumed, developing important supergene blankets. The low erosion rates since the Oligocene-Miocene have permitted the preservation of many of these supergene blankets.

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