Published online September 25, 2009; doi:10.1130/B26538.1

Geochemical variations in igneous rocks of the Central Andean orocline (13°S to 18°S): Tracing crustal thickening and magma generation through time and space

Mirian Mamani[1,†], Gerhard Wörner[1], and Thierry Sempere[2]

[1]*Abteilung Geochemie, Geowissenschaftlichen Zentrum der Universität Göttingen, Goldschmidtstrasse 1, D-37077 Göttingen, Germany*

[2]*Institut de Recherche pour le Développement and Université de Toulouse Paul Sabatier (SVT-OMP), Laboratoire Mécanismes de Transfert en Géologie, 14 avenue Edouard Belin, F-31400 Toulouse, France*

ABSTRACT

Compositional variations of Central Andean subduction-related igneous rocks reflect the plate-tectonic evolution of this active continental margin through time and space. In order to address the effect on magmatism of changing subduction geometry and crustal evolution of the upper continental plate during the Andean orogeny, we compiled more than 1500 major- and trace-element data points, and 650 Sr-, 610 Nd-, and 570 Pb-isotopic analyses of Mesozoic-Cenozoic (190–0 Ma) magmatic rocks in southern Peru and northern Chile (Central Andean orocline), mostly from new data and the literature. This data set documents compositional variations of magmas since Jurassic time, with a focus on the Neogene period, when major crustal thickening developed and its influence on magma composition was most pronounced. We relate the observed variations in Sr/Y, La/Yb, La/Sm, Sm/Yb, and Dy/Yb ratios, as well as in Sr-, Nd-, and Pb-isotopic ratios, to the crustal structure and evolution of the Central Andean orocline. In particular, the evolution of Dy/Yb and Sm/Yb ratios, which track the presence of the higher-pressure minerals amphibole and garnet, respectively, in the lower crust, documents that crustal thickness has grown through time. Spatial variations in trace elements and isotopic ratios further suggest that crustal domains of distinct composition and age have influenced magma composition through some assimilation. The crustal input in Quaternary magmas is quantified to have been between 7% and 18% by simple two-components mixing. When comparing our geochemical data set to the geological record of uplift and crustal thickening, we observe a correlation between the composition of magmatic rocks and the progression of Andean orogeny. In particular, our results support the interpretation that major crustal thickening and uplift were initiated in the mid-Oligocene (30 Ma) and that crustal thickness has kept increasing until present day. Our data do not support delamination as a general cause for major late Miocene uplift in the Central Andes and instead favor continued crustal thickening.

INTRODUCTION

The Andean Cordillera is a classic example of continental subduction-related magmatism. These magmas intrude and traverse a continental crust up to 70 km thick en route toward the surface. Because the composition of these magmas may be strongly affected by interaction with crustal material and because the thickness of the crust has changed through time, the Central Andes are an excellent natural observatory to study the interaction between crustal evolution and magma genesis. While the Andes span the entire length of the Pacific coast of South America, the Central Andes stand out as the segment of greatest orogenic volume (Fig. 1). Such a contrasted segmentation reflects the fact that the Andes have been built by a variety of processes that have changed along and across strike in nature, time, and intensity (Sempere et al., 2008). The Central Andes are themselves segmented into the northern Central Andes (5°30′S–~13°S), the Central Andean orocline (~13°S–28°S), and the southern Central Andes (28°S–37°S). Among these, the Central Andean orocline is characterized by the thickest continental crust of any subduction zone worldwide (Fig. 1), and it extends over southern Peru, Bolivia, northern Chile, and northwestern Argentina, covering an area of ~1,300,000 km^2. Its width, between the subduction trench and the sub-Andean front, is locally >850 km, and its crustal thickness reaches values >70 km, particularly along the main magmatic arc (James, 1971a, 1971b; Kono et al., 1989; Beck et al., 1996; Yuan et al., 2002). The Central Andean orocline thus appears to be an extreme case of crustal thickening among the various arc orogens of the Pacific Ocean margins. Because this segment does not result from continent-continent collision and displays significant tectonic shortening mainly along its eastern half, the origin of its anomalous crustal thickness remains uncertain (Sempere et al., 2008). The fact that the crust is thickest along the main arc (and in some other volcanic areas) is intriguing and has suggested that Central Andean magmatism participated in thickening the crust (James, 1971b; Kono et al., 1989; James and Sacks, 1999). Crustal thickening in the Central Andean orocline, however, is dominantly believed to have developed since ca. 30 Ma in association with tectonic shortening (e.g., Isacks, 1988; Sempere et al., 1990, 2008; Gregory-Wodzicki, 2000; Garzione et al., 2008). Magmatic addition to the arc has apparently remained largely constant, and available estimates are too low to explain the observed crustal thickness (Wörner et al., 2000, 2002; Trumbull et al., 2006).

The boundaries of the active Central volcanic zone coincide with those of the Central Andean orocline. The northern boundary also apparently coincides with the edge of the subducted Nazca Ridge and thus of the present-day "flat-slab" subduction (Hampel, 2002; Fig. 2). Previous studies of Central Andean magmatism have concluded that the thickened crust has strongly affected Andean arc magmas through crustal contamination (e.g., Wörner et al., 1988;

†E-mail: mmirian@gwdg.de

GSA Bulletin; January/February 2010; v. 122; no. 1/2; p. 162–182; doi: 10.1130/B26538.1; 14 figures; Data Repository item 2009161.

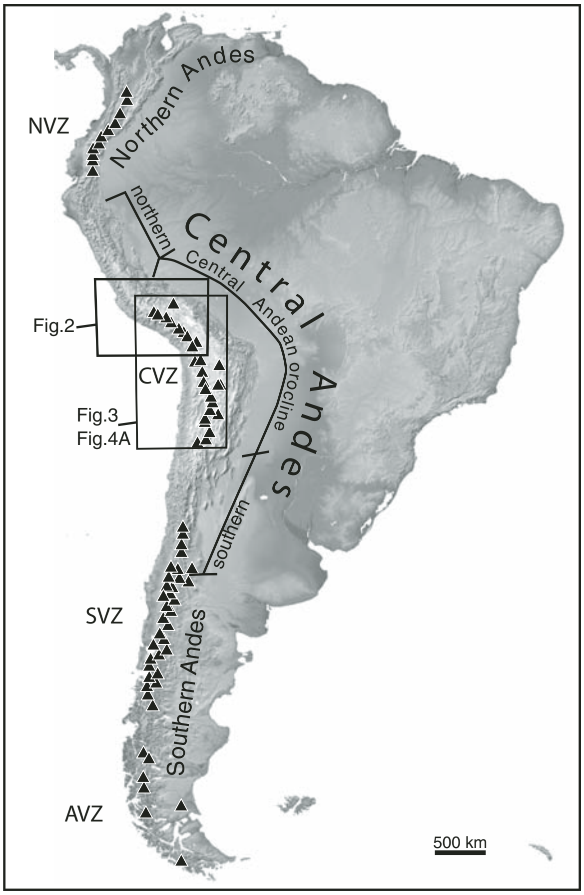


Figure 1. Segmentation of the Andes Cordillera (modified after Sempere et al., 2002). The Central Andean orocline is by far the segment of largest orogenic volume, which decreases northward and southward. Note that the boundaries of the Central volcanic zone coincide with those of the Central Andean orocline. Areas covered by Figures 2, 3, and 4A are indicated by boxes. Black triangles: active Andean volcanic zones (NVZ, CVZ, SVZ, AVZ—North, Central, South, and Austral volcanic zones, respectively). Topographic image is from http://photojournal.jpl.nasa.gov/catalog/PIA03388.

Davidson, 1991; Kay et al., 1994, 1999; Kay and Mpodozis, 2001). In particular, lavas younger than 5 Ma from the central and southern Central volcanic zone have compositional characteristics that suggest a higher degree of crustal contamination compared with similar andesitic magmas erupted in regions where, or times when, the crust was thinner (Davidson, 1991; Haschke et al., 2002). However, alternative mechanisms have been proposed, such as "source contamination" by subduction erosion of the forearc (Stern, 1991; Kay et al., 2005; Clift and Hartley, 2007), or influence of enriched subcontinental mantle lithosphere in the source region (Rogers and Hawkesworth, 1989). In this paper, we mainly focus on andesites erupted since ca. 30 Ma, i.e., during the main period of crustal thickening, but we also consider older rocks for comparisons. In particular, we use the Dy/Yb and Sm/Yb ratios, which are sensitive to the presence of the higher-pressure minerals amphibole and garnet, respectively, to document a signature of growing crustal thickness in Central Andean magmas.

To set the background for this discussion, we first present a summary of the Mesozoic-Cenozoic evolution of Central Andean arc magmatism, and, in order to correlate magma compositions to the evolution of the crust, we provide a consistent stratigraphic nomenclature based on absolute ages and correlations. This is needed because previous stratigraphic systems have relied on local and multiple names for units, which were often poorly dated and correlated (e.g., Instituto Geológico Minero y Metalúrgico Bulletins; Tosdal et al., 1981; Kaneoka and Guevara, 1984; Klinck et al., 1986; Mukasa, 1986a, 1986b; Clark et al., 1990; Fornari et al., 2002; Quang et al., 2005; Echavarria et al., 2006; Thouret et al., 2007; and references therein). These works illustrate recurrent debates and disagreements about names of stratigraphic units. Within our more consistent stratigraphy, we present new major- and trace-element and isotopic data for Cretaceous to Holocene igneous rocks, which we combine with previously published data to document in detail the variations in time and space of arc magma composition in southern Peru and northern Chile. This comprehensive data set is complemented by analyses from Proterozoic basement rocks in order to better define a mass balance for geochemical and isotopic variations in magmatic rocks erupted through time in different crustal domains (Mamani et al., 2008a).

GEOLOGIC EVOLUTION OF SOUTHERN PERU AND ADJACENT AREAS

It is widely accepted that the Mesozoic-Cenozoic magmatic and tectonic evolution of the Central Andean margin has strongly depended on a variety of subduction parameters. These include the rate and obliquity of plate convergence, the dip and rollback of the subducting plate, and their variations in time and space, as well as the existence of large geological, topographic, and/or thermal heterogeneities subducted with the oceanic plate. In addition, the structural and morphological heterogeneities of both plates, the geometrical and thermal evolution of the mantle wedge (e.g., Russo and Silver, 1996; Heuret et al., 2007; Schellart et al., 2007), and the growing thickness of the continental crust (Isacks, 1988; Somoza, 1998; Angermann et al., 1999; Oncken et al., 2006; Sempere et al., 2008) have all played important roles.

In southern Peru and northern Chile, arc magmatism migrated toward the continent at least from the Jurassic to the Paleogene (Pitcher et al., 1985; Clark et al., 1990; Trumbull et al., 2006). However, arc-related plutons, such as those forming the Coastal Batholith, are generally >1-km-thick tabular bodies (Haederle and Atherton, 2002; Pino et al., 2004; Jacay and Sempere, 2008) and, because of tectonic tilting, their present location may not reflect the position of the arc at the time of their emplacement.

In this section, we summarize, with emphasis on the magmatic arc, current knowledge regarding the geologic evolution of the study area, and we review Andean magmatic rocks before and during crustal thickening in southern Peru (Fig. 2). Information on the type, volumes, location, and timing of the magmatic arcs in northern Chile was mostly taken from Trumbull et al. (2006) and our own unpublished data. Because the timing of crustal thickening is crucial to understanding the evolution of the Central Andean volcanic units, we also provide a synthesis of current knowledge regarding this topic.

Overview: Migrating Magmatic Arcs and Evolving Backarc Basins

The history of the magmatic arc and neighboring regions provides insights into the evolution of the Central Andean subduction system. We propose hereafter a brief summary of the evolution of the arc region since the late Paleozoic on the basis of our analyses of the literature, and, in particular, of syntheses available for southern Peru (Sempere et al., 2002, 2008; Pino et al., 2004; Roperch et al., 2006). We used

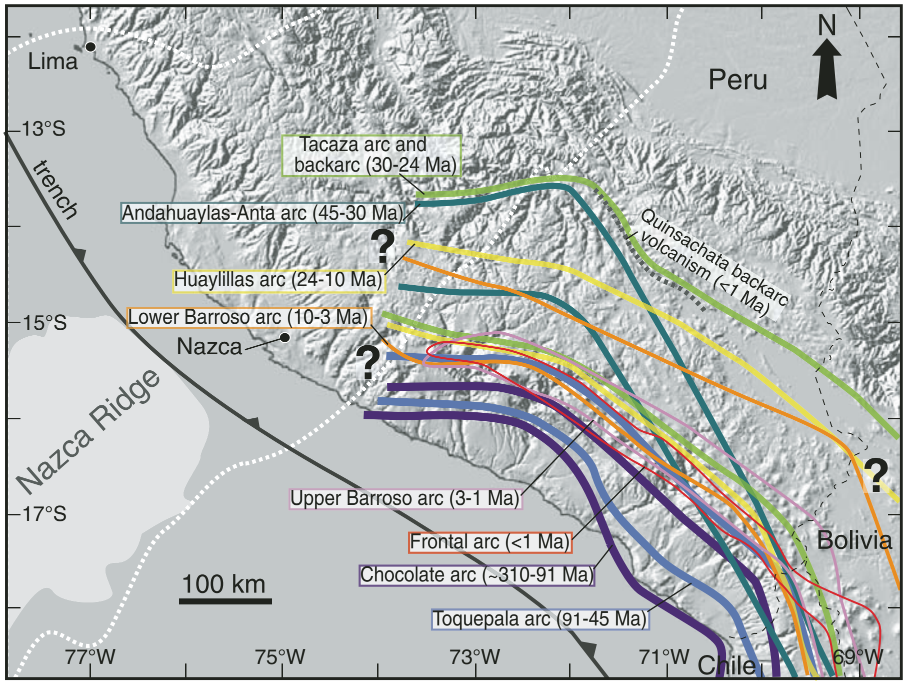


Figure 2. Location, extension, and age (Ma) of the volcanic arcs and backarc areas distinguished in southern Peru. The successive arcs approximately extended between the labeled lines of same color and thickness, drawn on the basis of dated outcrops and available geological maps. Extension of Nazca Ridge (white dashed lines) is after Hampel (2002).

numerous isotopic ages published for the region and other information from the literature, and observed contact relationships and morphological criteria in the field and on satellite images. A compilation of ages relative to the magmatic arcs recognized in this paper is given in the electronic GSA Data Repository item (Table DR1).[1]

Three main periods of unequal durations help to summarize the long-term evolution of the active margin corresponding to the present-day Central Andean orocline:

(1) From the late Paleozoic (ca. 400? Ma) to the mid-Cretaceous (91 Ma), the margin dominantly underwent tectonic stretching, leading to the formation of an overall marine backarc basin.

(2) From the Late Cretaceous (91 Ma) to the mid-Oligocene (30 Ma), the magmatic arc was large enough to form a significant, continuous relief, thus indicating incipient crustal thickening. In contrast with the previous evolution, the backarc basin was occupied by mostly continental environments. No clear and large-scale compressional structure is known from this period. It is noteworthy that ca. 45 Ma, the arc migrated to the north by up to 200 km in the western part of the study area, and it occupied this position until ca. 30 Ma.

(3) The major crustal thickening typical of the Andean orogeny has developed since the mid-Oligocene (30 Ma), while the main arc has migrated back toward the trench (see following). Transpressional to compressional deformation developed in the northeastern Altiplano, Eastern Cordillera, and Subandean belt; in contrast, extensional, transtensional, and transpressional deformations have occurred in the forearc, arc, and southwestern Altiplano of southern Peru (Fig. 3; Sempere and Jacay, 2007, 2008). This tectonic development created a number of basins in the backarc, arc, and forearc and fragmented or terminated others, resulting in a somewhat complex stratigraphic record in time and space. This evolution possibly developed in relation with variations in the convergence rate, but not in the orientation of convergence, which has changed little since ca. 49 Ma (Pardo-Casas and Molnar, 1987).

Stretching the Margin: Late Paleozoic to Mid-Cretaceous (ca. 400? to 91 Ma)

The geological record relevant to this study starts with sedimentary deposits of Devonian to mid-Carboniferous age, which onlap mainly 1-Ga-old basement rocks. These strata are conformably overlain by a thick but poorly known succession of dominantly basic volcanic and volcaniclastic rocks, referred to as the Chocolate Formation, which reflect coeval arc construction and probably also backarc extension (Martínez et al., 2005). The thickness of this unit increases from ≥1 km in the paleobackarc to >3 km along the present-day coast.

No direct dating of the Chocolate Formation lavas has been achieved so far due to their pervasive weathering. Stratigraphic relationships in the paleobackarc of southern Peru indicate that the Chocolate Formation conformably overlies strata of Middle to Late Carboniferous age (Pino et al., 2004; 2008, personal obs.), and it is conformably to gradually overlain by limestones that bear Sinemurian ammonites (ca. 195 Ma; Sempere et al., 2002). These relationships suggest that the unit spans at least the ca. 310–195 Ma interval, possibly with some internal hiatuses. However, this "Chocolate" magmatism continued well into the Jurassic along the present-day coastal area, reflecting persistent activity of the arc system, while the backarc northeast of this arc underwent lithospheric thinning and significant crustal downwarping. This backarc basin remained marine during most of the ca. 195–91 Ma time interval, but it turned terrestrial between ca. 135 and ca. 115 Ma (Sempere et al., 2002). Maximum water depths were reached by 180–170 Ma, as indicated by sedimentary facies (Pino et al., 2004), suggesting that lithospheric thinning and tectonic stretching of the margin culminated at about that time. A poorly known episode of oblique transcurrent deformation affected the backarc in the Early Cretaceous (Sempere et al., 2002).

In the Ilo area, the arc domain was the locus of a massive plutonic pulse between ca. 160 and ca. 150 Ma (Clark et al., 1990), leading to local growth of the Coastal Batholith; a further, major growth developed between 110 and 95 Ma by emplacement of more granitoids (Clark et al., 1990). The backarc record, however, testifies to only very limited volcanic activity during the ca. 195–91 time interval, in contrast with the subsequent period (Callot et al., 2008).

In northernmost Chile, the partly coeval La Negra Formation is locally up to 10 km thick and includes mainly andesitic lavas, as well as large plutonic bodies (Oliveros et al., 2006; Charrier et al., 2007). Whereas plutonism developed over the 190–100 Ma interval, volcanic activity is recorded starting at 176 Ma, but it was particularly voluminous between 165 and 150 Ma (Oliveros et al., 2006). As in Peru, the arc in which this volcanic unit and related plutons were produced, as well as the backarc basin, evolved in a markedly extensional context (Charrier et al., 2007).

[1]GSA Data Repository item 2009161, ages and chemical analyses of central Andean igneous rocks, is available at http://www.geosociety.org/pubs/ft2009.htm or by request to editing@geosociety.org.

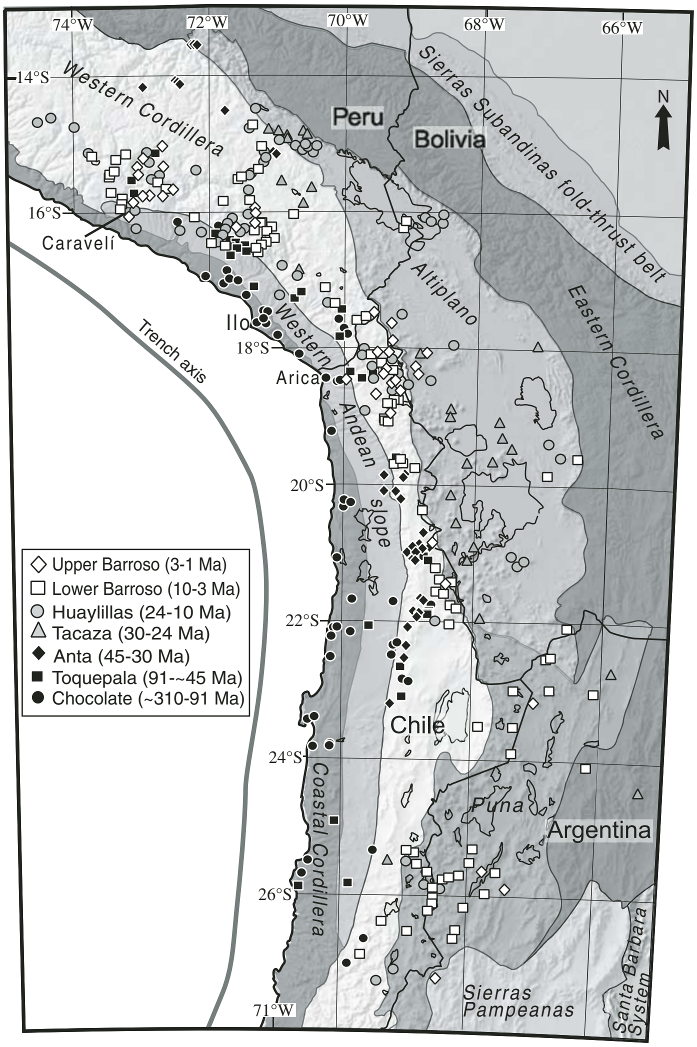


Figure 3. Main geomorphic units in the Central Andean orocline, and Chocolate to Upper Barroso localities sampled in this study (locations of younger samples are shown in Fig. 4). The Huaylillas and Barroso samples located in the western Andean slope are ignimbrites intercalated with Neogene strata.

Incipient Crustal Thickening: Late Cretaceous to Mid-Oligocene (91–30 Ma)

91–45 Ma: The Toquepala Arc

The ≥1.5-km-thick Toquepala Group is known only south of 16°40′S, and it mainly consists of volcanic rocks and subordinate volcaniclastic and fine-grained clastic deposits. In contrast to the Chocolate Formation volcanic rocks, which were erupted in submarine environments, the Toquepala volcanic rocks were erupted subaerially. North of 16°40′S, the Toquepala arc is represented by most plutons in the Coastal Batholith. Coeval erosional products of this arc crop out to the northeast, where their thickness may be >1.5 km.

Rapid initiation of the construction of the Toquepala arc is expressed by the abrupt invasion of the backarc basin by coarse volcaniclastic products and other continental deposits that sharply overlie the Albian-Turonian carbonate platform, which had occupied the backarc until then (Sempere et al., 2002; Callot et al., 2008). This sharp, well-dated turning point in the stratigraphic record was accompanied by a complete reversal of basin polarity, reflecting the emergence of a continuous emerged relief coinciding with the arc and thus suggesting incipient crustal thickening at the arc. It also coincides with the emplacement of the most voluminous units in the Coastal Batholith, which Mukasa (1986a) dated between 91 and 70 Ma. Numerous age dates have been obtained on Toquepala volcanic rocks (Pitcher et al., 1985; Boily et al., 1989; Clark et al., 1990, and references therein), but, due to significant alteration, no reliable ages are available for the part of the unit older than 75 Ma. Large ignimbrite sheets were erupted between ca. 75 and ca. 65 Ma (Martínez and Cervantes, 2003). Activity of the Toquepala arc apparently waned after ca. 50 Ma, and the youngest available ages are slightly older than ca. 45 Ma (Clark et al., 1990). In the Caravelí area of the present-day forearc, plutonic rocks of the Coastal Batholith are overlain by ~50- to 45-Ma-old, dominantly fine-grained continental strata ("Moquegua A" unit; Roperch et al., 2006), indicating that at least part of the batholith had been exhumed by this time.

In northern Chile, a similar evolution is described: plutonic, volcanic, and volcaniclastic rocks in the west define a >100-km-wide coeval arc, from which the backarc basin in the east was fed. This basin was filled by an ~6-m-thick accumulation of essentially continental, finer-grained red strata that unconformably overlie older units (Mpodozis et al., 2005; Charrier et al., 2007). Although interpretations of the coeval regional tectonic regime have varied from extensional to transpressional or compressional (see review by Mpodozis et al., 2005), it seems that the basin was dominated by extension during most of this interval (Charrier et al., 2007).

45–30 Ma: The Andahuaylas-Anta Arc

The interval from 45 to 30 Ma was characterized by a northward migration of the arc in the ~150-km-wide portion of southern Peru located between ~71.3°W and ~74.5°W (Fig. 2; Perelló et al., 2003). Arc rocks include the Andahuaylas-Yauri Batholith and the Anta volcanic rocks. Available geochronologic data consistently indicate that the arc occupied this anomalous position during the 45–30 Ma interval (Perelló et al., 2003). This arc migration must have been caused by a decrease in the slab dip in this particular area (Sandeman et al., 1995; James and Sacks, 1999), which did not suppress magmatic activity, but the cause of this relative flattening remains unknown.

In northern Chile, the coeval magmatism is mainly represented by intrusions along the Cordillera de Domeyko and the "Falla Oeste" and related structures (see age compilation by Trumbull et al., 2006), where it generated significant ore deposits (Cornejo, 2005). Transpressional tectonics caused protracted deformation and exhumation of the arc during this time interval (Maksaev and Zentilli, 1999; Charrier et al., 2007). Starting ca. 44 Ma, conglomerates derived from this "Incaic Range" accumulated in the basin to the east (Mpodozis et al., 2005; Charrier et al., 2007).

In southern Peru, the western area of the forearc records the abrupt arrival and accumulation of coarse conglomerates starting ca. 44.5 Ma (Roperch et al., 2006), indicating that some significant relief also began to be created at that time north of this area.

Major Crustal Thickening: Mid-Oligocene to Quaternary (30–0 Ma)

Overview: Synorogenic Sedimentary Deposits and Ignimbrites

The "modern" Andean orogenic period coincided with the back-migration of the main arc system toward the trench. This back-migration is evident when considering the succession of arcs recognized in this paper (Fig. 2). However, although we distinguish and name these successive arcs, we caution that there are overlaps in their geographic location as well as in the nature of their volcanic products. The back-migration was apparently inaugurated by the widespread eruption of the Tacaza mafic lavas and related volcanic rocks, starting ca. 30 Ma (Clark et al., 1990; Wasteneys, 1990; Sandeman et al., 1995).

The initiation of the arc back-migration is well expressed in the forearc sedimentary record of southern Peru. In the entire forearc, a major stratigraphic discontinuity well dated at 30 Ma (Roperch et al., 2006) separates the "Moquegua B" unit (45–30 Ma), which is nearly devoid of volcanic material, from the "Moquegua C" unit (30–15 Ma), which in contrast consists of a variety of reworked volcanic products (Roperch et al., 2006). Since 30 Ma, volcanic and other rocks have been eroded from reliefs located in the Western Cordillera, and the resulting material has accumulated in the forearc.

The "Moquegua D" unit (15–0 Ma) mainly consists of thick, very coarse, dominantly volcaniclastic conglomerates that were produced by intense erosion of extensive volcanic regions (Huaylillas and Lower Barroso volcanics). The predominance of volcanic clasts indicates intense volcanism and marked topographic gradients in their source region of the Western Cordillera. Dated ignimbrites above (14.2 Ma; Thouret et al., 2007) and below (16.2 Ma; Roperch et al., 2006) a marked erosional surface at the base of the "Moquegua D" unit constrain the age of this discontinuity to ca. 15 Ma. Deposition of coarse conglomerates continued in many areas of the forearc and on the pre–valley-incision paleosurface of the western Andean slope as late as ca. 4 Ma. In northern Chile, no equivalents of the Moquegua A and B are known on the western edge of the Altiplano, but coarse conglomerates with abundant volcanic detritus did accumulate on the western Andean slope from the mid-Oligocene until ca. 4 Ma (Wörner et al., 2000, 2002; Charrier et al., 2007), representing southern equivalents of the Moquegua C and D units. These sediments form a thick wedge-shaped deposit, which has led to the relatively smooth western slope of the Central Andes as still seen today.

Erosion resulted in the incision of deep canyons in southern Peru starting at ca. 9 Ma, but occurring more intensely around ca. 4 Ma (Thouret et al., 2007), and from before ca. 12 Ma to after 2.7 Ma in northern Chile (Wörner et al., 2000, 2002). This incision was caused by a combination of low sea levels during glacial periods, increased (but localized) run-off from the glaciated Western Cordillera, and late Pliocene and Quaternary uplift of the continental forearc and western Andean slope itself. For a more in-depth discussion on valley incision and timing of uplift, see Thouret et al. (2007) and Hoke et al. (2007).

Large volumes of ignimbrites have erupted since ca. 30 Ma in the Central Andean orocline. These ignimbrites are intercalated in alluvial to lacustrine strata deposited in a number of basins in the forearc, arc, or backarc (e.g., Wörner et al., 2000, 2002; Roperch et al., 2006; Charrier et al., 2007; Thouret et al., 2007). Along the western Andean slope of southern Peru, two major intervals of ignimbrite eruption are recognized, namely between ca. 26 and ca. 18 Ma, and between ca. 10 and ca. 1.5 Ma (Roperch et al., 2006; Thouret et al., 2007). In northern Chile, numerous isotopic ages obtained on the Oxaya ignimbrites date this unit between 25.6 ± 0.9 and 19.0 ± 0.6 Ma, and more ignimbrite eruptions occurred especially between ca. 10 and 2.7 Ma (Wörner et al., 2000; Charrier et al., 2007). It seems likely that large volumes of hot mantle-derived magmas and/or underplating of the crust by hot asthenosphere caused the generation of these crustal melts (Sandeman et al., 1995; Babeyko et al., 2002).

30–24 Ma: The Tacaza Arc

In southern Peru, a large expansion of magmatism took place in the mid-Oligocene, starting at ca. 30 Ma (Klinck et al., 1986; Wasteneys, 1990; Sandeman et al., 1995; Fornari et al., 2002; Mamani et al., 2004), reflecting initiation of the back-migration of the arc system. This Tacaza magmatism affected the southwestern Altiplano and Western Cordillera (particularly its northeastern edge) and thus encompassed a width of ~320–350 km. Coeval rocks similar to the Tacaza mafic lavas have been reported from nearby Bolivia (Jiménez et al., 1993; Aitcheson et al., 1995; Sandeman et al., 1995) and are likely to occur along the border area of northern Chile. Some ignimbrites were also erupted during this time interval in southern Peru, Bolivia, and northern Chile (e.g., Jiménez et al., 1993; Wörner et al., 2000; Roperch et al., 2006; Trumbull et al., 2006; Charrier et al., 2007).

Volcanic rocks erupted during the Tacaza interval mainly consist of voluminous shoshonitic to absarokitic (= high-K) mafic lavas, representing essentially primary mantle melts, and subordinate andesitic calc-alkaline magmas exhibiting some evidence of crustal contribution (Wasteneys, 1990; Sandeman et al., 1995). Some of the Tacaza mafic lavas have Mg, Ni, and Cr contents approaching primitive values (Wasteneys, 1990; Sandeman et al., 1995), suggesting that they are, at least locally, undifferentiated products of mantle melting. The generation of the voluminous Tacaza mafic lavas appears to have been a consequence of the arc back-migration: the steepening of the slab that caused this back-migration created space and thus induced mantle upwelling and decompression, which resulted in pervasive mantle melting. Eruption of ignimbrites starting ca. 29 Ma, and of voluminous ignimbrites starting ca. 26 Ma, can be explained as the result of crustal magmatic processes triggered by the heating of the lower crust by hot advected magmas and/or underplated asthenosphere, starting at 30 Ma. Inception of the Tacaza interval also coincided with a marked increase in the Dy/Yb ratio, indicating the rapid development of amphibole in the deep crust, and thus the onset of significant crustal thickening (see following).

Effects of this back-migration appear in the stratigraphic record as major sedimentary discontinuities at 30 Ma in forearc and backarc basins, coeval with the eruption of the oldest Tacaza lavas (see previous). Completion of tectonic rotations along the south Peru margin (Roperch et al., 2006) and onset of compressional to transpressional deformation in the Eastern Cordillera (Sempere et al., 1990) also occurred during the Tacaza interval.

A number of intrusions of Tacaza age occur northeast of the Altiplano. Some other Oligocene magmatic rocks (e.g., the macusanites described by Kontak et al., 1990) that occur northeast of the Altiplano have a completely different origin; they are not directly related to subduction processes in the mantle but rather represent intracrustal melting of sedimentary rocks in the thickened Eastern Cordillera.

24–10 Ma: The Huaylillas Arc

The arc active between ca. 24 and ca. 10 Ma was located mostly along and northeast of the present Western Cordillera and western Altiplano, coinciding with the southwestern edge of the 30–24 Ma arc, and thus illustrating the southwestward progression of the arc back-migration. The most extensive and voluminous plateau-forming ignimbrites found in southern Peru and northern Chile were erupted between ca. 26 and ca. 18 Ma (see previous; Wörner et al., 2000, 2002; Farías et al., 2005; Charrier et al., 2007). These ignimbrite sheets are typically intercalated in, or overlie the top of, a succession of Upper Oligocene to Lower Miocene sedimentary strata ("Moquegua C" unit in Peru, and Azapa Formation in Chile).

Large shield andesite volcanoes dated between 18 and 10 Ma were built immediately after this main period of ignimbrite eruption. These old volcanoes are located west of the present arc in northernmost Chile between 18°S and 22°S but more to the east farther south, suggesting that the curvature of the active arc increased after 10 Ma.

Stratovolcanoes of Huaylillas age are typically preserved as eroded circular structures that include abundant hydrothermal alteration and epigenetic ore deposits in their interior. Such intense alteration suggests that precipitation at the time was able to recharge the shallow hydrothermal systems inside the volcanic structures. Their central depressions are erosional features, and they are related to chemical weathering of these altered cores, though they are commonly erroneously interpreted as calderas (e.g., Echavarria et al., 2006). Most of these volcanoes consist of andesites and only minor dacites. The former are generally poorly phyric and almost devoid of amphibole. The exposed sections typically display shallow-dipping, thin (≤20 m) flows. These lavas are phenocryst-poor, have rather uniform compositions, and tend to form thin lava flows and low-angle shield volcanoes. These characteristics and the general absence of amphibole indicate relatively low water contents and the high eruption temperatures.

10–3 Ma: The Lower Barroso Arc

The arcs that have been active since 10 Ma occupy a relatively narrow zone in the axis of the Western Cordillera and make up the present Central volcanic zone (Fig. 1). They were produced during a period of relatively low convergence rates and obliquity (Pardo-Casas and Molnar, 1987; Somoza, 1998).

Between 10 and 3 Ma, the arc was located along the western portion, i.e., close to the present active front of the Western Cordillera arc. It is preserved as remnant stratovolcanoes displaying substantial glacial erosion. Volcanic rocks of this age, however, also extend to the east, suggesting significant widening of the Miocene arc (see age compilation in Trumbull et al., 2006). In contrast with the older Huaylillas rocks, these andesites tend to be more porphyritic and show larger amounts of amphibole.

Ignimbrites of this age are relatively small in volume, and they are intercalated with Moquegua D sedimentary strata or cover geomorphic surfaces particularly in the forearc of northern Chile (e.g., the Lauca and Ujina ignimbrites, and coeval equivalents). In southern Peru, the Sencca ignimbrites were erupted between 5 and 1 Ma, and were thus coeval with the Barroso volcanic activity. The ignimbrite that crops out near the Sencca village in southernmost Peru is identical to the 2.72 Ma Lauca-Pérez ignimbrites of northern Chile and western Bolivia (Wörner et al., 2000). The Sencca ignimbrites were deposited between the Western Cordillera and the northwestern Altiplano, filling basins and incised valleys.

3–1 Ma: The Upper Barroso Arc

The arc active between 3 and 1 Ma is represented by volcanoes located along the Western Cordillera. Volcanoes assigned to this time interval have either been dated or classified based on their morphology. Some of these stratocones and volcanic complexes were active only until the Pliocene and Pleistocene. Volcanic structures range from steeply flanked stratovolcanoes to smaller mafic lava fields and lava shields.

Active Volcanic Arc (<1 Ma)

The active volcanic front presently lies ~230 km northeast of the trench and is located 115 ± 5 km above the Benioff-Wadati plane (England et al., 2004). We sampled all active volcanoes between 16°S to 27°S in southern Peru (this study) and northern Chile (Wörner et al., 1992; Wörner, 1999–2008, personal commun.). These consist of relatively simple and youthful stratovolcanoes with a central crater, long-lived dome-and stratocone clusters, or simple dome and dacite flow centers. Many small monogenetic scoria cones associated with lava flows (<5000 yr B.P.) are also observed (Delacour et al., 2007; de Silva and Francis, 1991). Apart from the small-volume monogenetic centers, the lavas of the larger stratovolcanoes generally consist of porphyric andesites to dacites with a generally unaltered, partly glassy groundmass. Hornblende, clinopyroxene, and plagioclase are abundant major phenocryst phases.

Pliocene-Quaternary Ultrapotassic Magmatism in the Backarc

Very minor Pliocene-Quaternary ultrapotassic magmatism is associated with a major transcurrent fault system that prolongates into the Altiplano (Fig. 2) and was fed by melting of enriched portions of the mantle lithosphere in the deep parts of lithospheric-scale fractures, through which they were erupted (Carlier et al., 2005). It includes the Quinsachata volcanic rocks (Carlier et al., 2005), which consist of recent (younger than 1 Ma) glassy shoshonites. Similar rocks are also known in the Bolivian Altiplano (Carlier et al., 2005).

Central Andean Uplift and Crustal Thickening: A Debated Issue

Crustal thickness is crucial to the understanding of arc magma geochemistry because it determines the length of passage through the crust, affects the locations and probability of stagnation levels and differentiation, and thus bears upon the degree of crustal contamination undergone by magmas ascending from the mantle wedge. Furthermore, a key phenomenon in a thickening crust is that its lower part is submitted to increasing lithostatic pressure, leading to the development of garnet as a stable phase at the expense of plagioclase in the liquidus mineral assemblage of the magmas. In addition, any assimilated country rocks will also develop a garnet-bearing residue under these high-pressure conditions (Hildreth and Moorbath, 1988; Kay et al., 1999; Macpherson et al., 2006; Lee et al., 2006; Muntener et al., 2001). Given the contrasted affinities of heavy rare earth elements (HREE) for garnet, and of middle rare earth elements (MREE) over HREE for amphibole, as well as the trace-element signature of plagioclase (low La/Sr, Eu anomaly; see the Geochemical Earth Reference Model reference database, http://www.earthref.org/GERM/), the presence of garnet, amphibole, and plagioclase characteristically affects REE profiles of magmas interacting with, or generated in, the lower part of a thickened crust (Kay et al., 1999; McMillan et al., 1993; Macpherson et al., 2006), a phenomenon best recorded by the Sm/Yb and Dy/Yb ratios (as illustrated hereafter). The Sr/Y, La/Yb, and La/Sm ratios are particularly sensitive to the presence of garnet versus plagioclase since Sr and La have a marked affinity for plagioclase (and Y and Yb for garnet respectively): thus, high (low) Sr/Y, La/Yb and La/Sm ratios are indicative of absence (presence) of plagioclase and presence (absence) of garnet in magma genesis.

Knowledge of the timing and processes of surface uplift and crustal thickening in the Central Andes is necessary to better understand the variations in trace-element ratios presented in this paper. Because of isostasy, crustal thickening generally results in surface uplift (although it is not the only possible cause for it; see following), and therefore knowledge of the uplift history in the Central Andes provides an indirect but useful framework to compare the geochemical evolution of arc magmas with the evolution of crustal thickening. Andean uplift, however, is being investigated from two different approaches: (1) Thermochronology estimates the timing of rock uplift and exhumation of a volume of rock as it reaches Earth's surface through erosion. (2) Paleoaltitudes can be derived from a number of methods based on geochemical markers or morphological measurements of fossil leaves, which allow the amount and timing of surface uplift, i.e., the increase in elevation of Earth's surface, to be assessed (e.g., Kohn, 2007; Garzione et al., 2008, and references therein; Gregory-Wodzicki, 2000; Kowalski, 2002). Paleoaltitude estimates obtained by these latter methods are of local value only and generally have large uncertainties. More recently, a combined phylogeographic-phylochronologic approach was developed to address the issue of where and when high altitudes were acquired in the Andes at a regional scale (Picard et al., 2008), but this method can only deal with threshold elevation ranges, not paleoaltitudes.

In spite of the numerous studies performed during the last decades, the chronology of crustal thickening and related uplift in the Central Andes has remained debated until recently (e.g., Sempere et al., 2008, and references therein). Several independent methods and studies suggest that the first significant uplift occurred in the Oligocene–early Miocene interval (ca. 30–18 Ma), and a major episode of crustal thickening (e.g., Picard et al., 2008, and references therein) is inferred in relation with the major shortening initiated at ca. 26 Ma in the Eastern Cordillera of western Bolivia (Sempere et al., 1990). This time coincides with an increase in exhumation rates reported at ca. 25 ± 5 Ma in the same area (Kennan, 2000). Moreover, the magmatic record of southern Peru indicates that the increasing crustal thickness crossed a petrogenetic threshold due to phase changes in minerals associated with differentiation and assimilation of magmas with increasing pressure during the interval ca. 23 to ca. 17 Ma (Wasteneys, 1990; Sandeman et al., 1995). This is exactly the same time interval during which the critical 2.0–2.5 km altitude was reached in the same region according to Picard et al. (2008).

Many studies recognize the onset of a second major episode of uplift in southern Peru and Bolivia at ca. 10–9 Ma (e.g., Schildgen et al., 2007; Thouret et al., 2007; Garzione et al., 2008; Sempere et al., 2008, and references therein). However, disagreement exists about the cause of this uplift. One view favors massive delamination of dense lower lithospheric material into the mantle, rather than further crustal thickening, to have been the cause for uplift at this time (Garzione et al., 2006, 2007, 2008; Molnar and Garzione, 2007). This interpretation is, however, debated, because no magmatic products typical of a delamination process have been recognized in this region so far (e.g., Hartley et al., 2007), and significant crustal thickening should have preceded this process in order to allow for density instability and delamination to occur.

On the basis of their combined phylogeographic-phylochronologic approach to this issue, Picard et al. (2008) proposed that high altitudes and related crustal thickening remained essentially restricted to the Central Andean Orocline during the late Oligocene and most of the Miocene, until thermal weakening of the orogen in the late Miocene (Isacks, 1988) allowed rapid further crustal thickening and uplift to propagate northward into central and northern Peru, and presumably southward into central Chile and western Argentina. Such propagation may be explained by longitudinal flow of ductile lower-crustal material northward and southward from the overthickened Central Andean orocline (Husson and Sempere, 2003; Yang et al., 2003). This interpretation is consistent with the marked decrease in orogenic volume observed north and south of the orocline, and it ultimately suggests that the larger orogenic volume in the Central Andean orocline results from longer cumulative crustal thickening there.

Thus, the picture currently emerging is that surface uplift in the Central Andean orocline developed in two steps (Garzione et al., 2008; Picard et al., 2008; Sempere et al., 2008): the first one, in the Oligocene and early Miocene, reflected coeval crustal thickening; the second one, in the late Miocene, was apparently vigorous, but its relation to crustal thickening is debated.

As will be made clear later herein, our geochemical data support the timing of uplift, i.e., that crustal thickening in the Central Andean orocline started in the mid-Oligocene (30 Ma) and has kept increasing since then, during and after the major late Miocene surface uplift.

GEOCHEMICAL DATA

The geochemical data set contains nearly 392 new analyses from the seven volcanic arc systems recognized in southern Peru. These and 60 published data on Proterozoic and Paleozoic rocks of the Coastal Cordillera (from Tilton and Barreiro, 1980; Mukasa, 1986b; Loewy et al., 2004), Jurassic and Cretaceous rocks of the Chocolate and Toquepala arcs (52 samples; Mukasa, 1986b; Boyle, 1989; Clark et al., 1990), Oligocene rocks of the Tacaza arc (22 samples; Mamani et al., 2004, and references therein), and Miocene rocks from the Altiplano and Eastern Cordillera (30; Sandeman et al., 1995; Mamani et al., 2004) were combined with our previously unpublished data from northern Chile (395 unpublished samples). The data set also includes data from the literature and our own previous work in the region (Wörner et al., 1988, 1992; Thouret et al., 2005; Paquereau-Lebti et al., 2006; Delacour et al., 2007; Mamani et al., 2008a). With respect to the active (younger than 1 Ma) volcanic front, the data cover all active volcanoes from the entire Central volcanic zone (Fig. 1). Our data set thus comprehensively covers magmatism in the Central Andes from the early Mesozoic to the present. A compiled data set is published here in the electronic supplement (Table DR2 [see footnote 1]), in which the analytical methods are also described (Appendix).

RESULTS

The data were grouped according to age and geographic location (Fig. 3). The results are displayed in diagrams plotting geochemical parameters versus age. Age groups correspond to the succession of magmatic arcs described previously. Ages were obtained from K-Ar and Ar-Ar dating methods and inferred from stratigraphic relationships. The age of ~70% of our samples is tightly constrained by direct data. In addition, we used geomorphologic comparisons of the degree of erosion with dated stratovolcanoes in similar areas and at similar elevation (to account for differential erosion rates) to assign undated volcanic centers to appropriate age groups. Such age estimates are sufficiently reliable when the purpose is to assign volcanic edifices within age groups of several million years duration each.

Our data set is referenced (1) geographically by UTM coordinates (see electronic supplement, Table DR2 [see footnote 1]), and (2) according to their location in Mamani et al.'s (2008a) Arequipa, Cordillera, and Coastal crustal domains. These domains were defined on the basis of the isotopic composition of basement rocks and arc magmas because they were erupted through, and contaminated by, Andean crust of distinct chemical and isotopic composition (Fig. 4A). The isotope domains correlate to crustal structure as defined by a three-dimensional (3-D) density model as well as the structural grain of the central Andes (Mamani et al., 2008a; Ramos, 2008, and references therein). The Arequipa domain is a dominantly mafic Proterozoic terrane accreted to the Gondwana margin (Ramos, 2008), and it is characterized by mainly nonradiogenic Pb isotopes and low (between 16.083 and 18.453) $^{206}Pb/^{204}Pb$ ratios, typical of an ancient high-grade terrane. The Cordillera domain is here divided into the (northern) Paracas and (southern) Antofalla

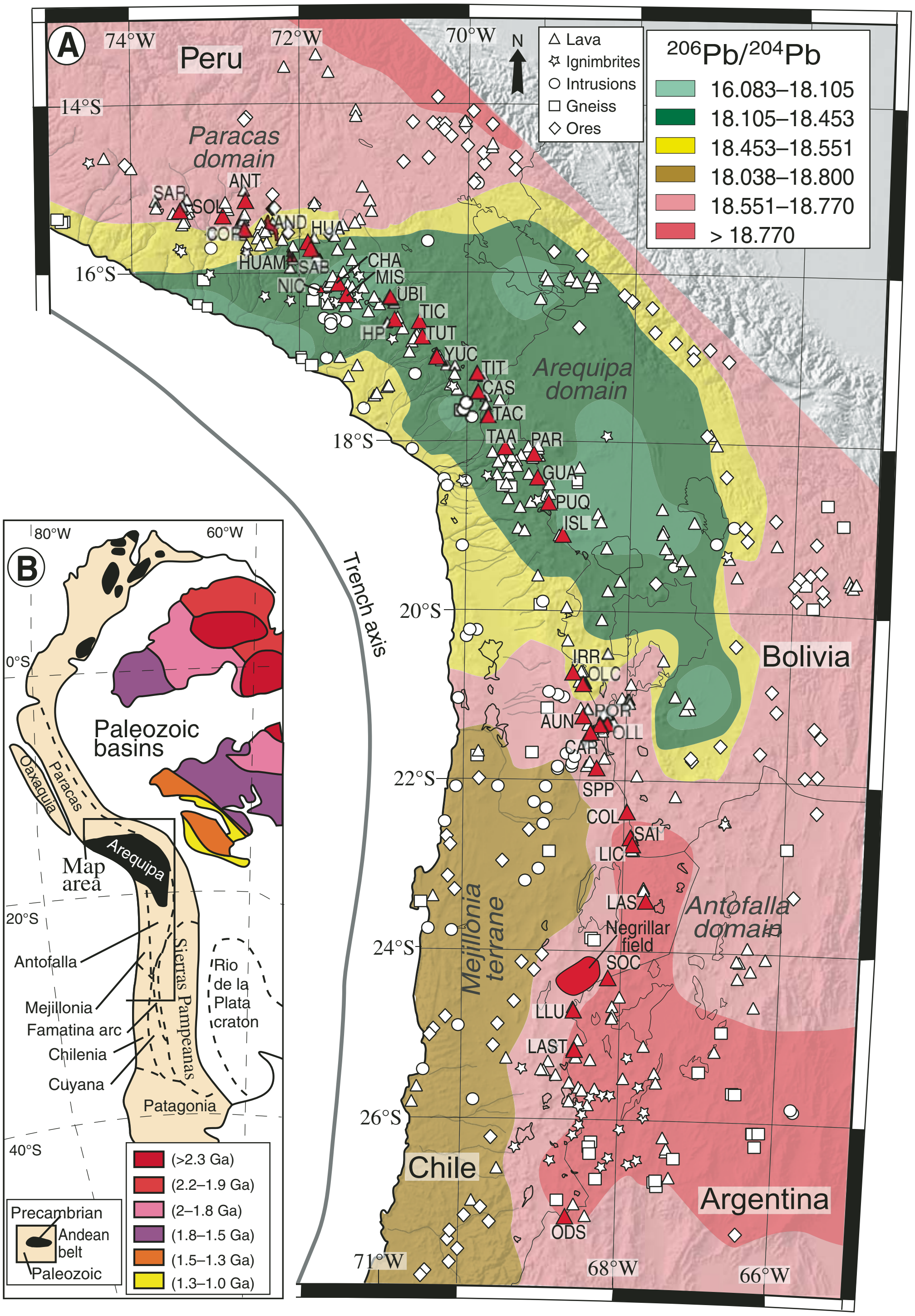


Figure 4. (A) Crustal domains distinguished on the basis of present-day $^{206}Pb/^{204}Pb$ ratios (modified after Mamani et al., 2008a). The Arequipa domain (16.083 < $^{206}Pb/^{204}Pb$ < 18.453) is in green, the Paracas and Antofalla domains ($^{206}Pb/^{204}Pb$ > 18.551) are in pink, and the transition zones (18.453 < $^{206}Pb/^{204}Pb$ < 18.551) are in yellow; the Mejillonia terrane (18.038 < $^{206}Pb/^{204}Pb$ < 18.800) is in brown. Red triangles are Quaternary volcanoes: SAR—Sara Sara, SOL—Solimana, ANT—Antapuna, AND—Andahua field, HUA—Hualca Hualca, HUAM—Huambo field, SAB—Sabancaya, CHA—Chachani, NIC—Nicholson, MIS—Misti, UBI—Ubinas, HP—Huayna Putina, TIC—Ticsani, TUT—Tutupaca, YUC—Yucamane, TIT—Titire, CAS—Casiri, TAC—Tacora, TAA—Taapaca, PAR—Parinacota, GUA—Guallatiri, PUQ—Puquintica, ISL—Isluga, IRR—Irruputuncu, OLC—Olca, AUN—Auncanquilcha, OLL—Ollagüe, POR—Porunita, CAR—Carcote, SPP—San Pedro–San Pablo, COL—Colorado, SAI—Sairecabur, LIC—Licancabur, LAS—Lascar, SOC—Socompa, LLU—Llullaillaco, LAST—Lastarria, ODS—Ojos del Salado. Samples used to define the crustal domains are located by open symbols according to rock types (see box). (B) Map illustrating the basement domains in western South America (adapted from Mamani et al., 2008a; Ramos, 2008).

domains: both are dominantly felsic terranes that were amalgamated to the western Gondwana margin (Ramos, 2008), and they have radiogenic Pb values, i.e., high (>18.551) $^{206}Pb/^{204}Pb$ ratios. The Coastal Cordillera domain reflects major mafic Mesozoic juvenile contribution to the Arequipa (Mamani et al., 2008a) and Mejillonia terranes (Lucassen et al., 2001), and it exhibits nonradiogenic and radiogenic Pb isotopes ($18.038 < {}^{206}Pb/^{204}Pb < 18.8$).

Major and Trace Elements

Lava types analyzed in this study are mostly andesites to dacites, along with smaller portions of ignimbrites such as rhyodacites and rhyolites. Intrusive rocks analyzed include gabbros, diorites, granodiorites, and granites. The vast majority of samples from all arcs plot in a typical subalkaline trend. However, some lava samples from the Toquepala (91–45 Ma), Andahuaylas-Anta (45–30 Ma), and northeastern Tacaza (30–24 Ma) and Huaylillas (24–10 Ma) arcs have higher K_2O contents relative to SiO_2 and Al_2O_3 (Fig. 5).

All studied lavas exhibit a typical and relatively uniform arc signature characterized by a marked enrichment in Cs, Rb, Ba, K, and Th, combined with a marked depletion in Nb and pronounced light (L) REE enrichment (La/Sm). In order to evaluate REE patterns, the Sr/Y, La/Yb, La/Sm, Sm/Yb, and Dy/Yb ratios were plotted versus SiO_2 by distinguishing age groups (Fig. 6). In order to avoid the effect of differentiation on trace-element systematics, we only discuss rocks with <65 wt% SiO_2. For these, maximum Sr/Y, Sm/Yb, and Dy/Yb ratios clearly increase through time (see also Haschke et al. [2002] and Haschke and Günther [2003] for a similar finding regarding rocks in northern Chile). An incipient increase is apparent for the 91–30-Ma-old rocks (Sr/Y changing from 3 to 40, Sm/Yb = 1–3, Dy/Yb = 1.6–2.2). These trace-element ratios further increase at 30–24 Ma (Sr/Y = 4–60, Sm/Yb = 2–5, Dy/Yb = 1.6–2.5) and reach their maximum values in <3-Ma-old lavas (Sr/Y = 30–100, Sm/Yb = 2–10, Dy/Yb = 1.9–3.6). Differentiated rocks (>65 wt% SiO_2) show lower Sr/Y, Sm/Yb, Dy/Yb ratios, and higher La/Sm and high La/Yb ratios, due to the effect of fractionation by feldspar and possibly accessory minerals.

Isotopic Compositions

Most $^{206}Pb/^{204}Pb$ ratios and ε_{Nd} values dealt with here (Fig. 7) are those used to delineate Mamani et al.'s (2008a) Arequipa and Paracas crustal domains (Fig. 4). The Arequipa domain has the lowest $^{206}Pb/^{204}Pb$ ratios (16.083–18.453) and ε_{Nd} values (–4 to –12). The Paracas domain is represented by the highest $^{206}Pb/^{204}Pb$ ratios (>18.551) and ε_{Nd} values (0 to –4). The $^{87}Sr/^{86}Sr$ ratios increase in more differentiated rock (i.e., ignimbrites). The lowest $^{87}Sr/^{86}Sr$ initial ratios (0.703–0.706) and highest ε_{Nd} values (2 to –1) correspond to the Chocolate, Toquepala, and Andahuaylas-Anta rocks, i.e., to rocks older than 30 Ma. High $^{87}Sr/^{86}Sr$ initial ratios (between 0.705 and 0.709) are exhibited by a few rocks within the 91–45 Ma interval and all rocks from 30 to 0 Ma. Samples from the Chocolate and Toquepala arcs have radiogenic mantle $^{206}Pb/^{204}Pb$ ratios (18.5–18.8) and nonradiogenic $^{206}Pb/^{204}Pb$ ratios (17.5–18).

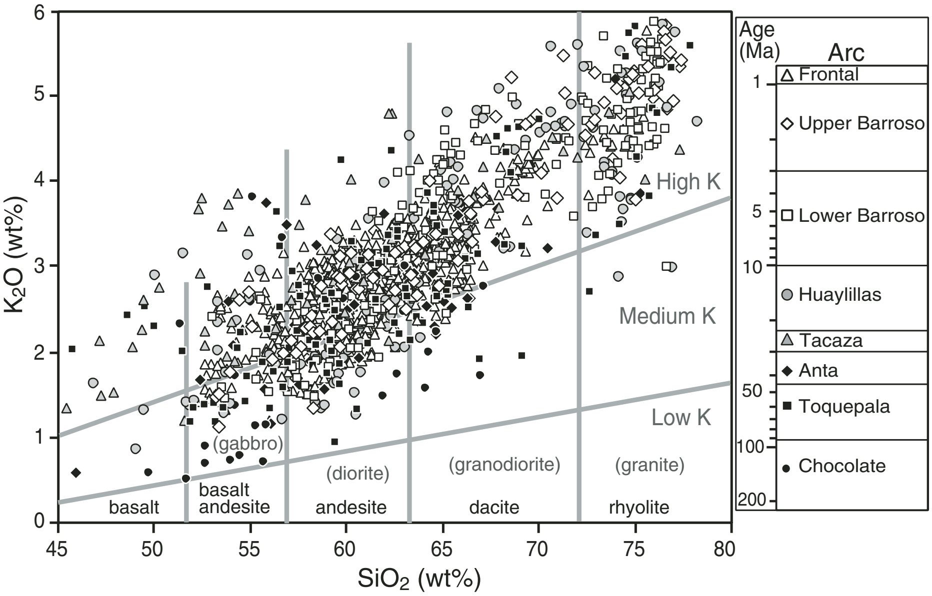


Figure 5. Plot of K_2O versus SiO_2 contents (wt%) for volcanic arcs in southern Peru. Calc-alkaline series classification is simplified after Le Maitre (1989). Most rocks plot in the high-K calc-alkaline field. Chocolate arc rocks plot in the medium-K field. Oxide contents are recalculated to 100% on a volatile-free basis and with all Fe as FeO.

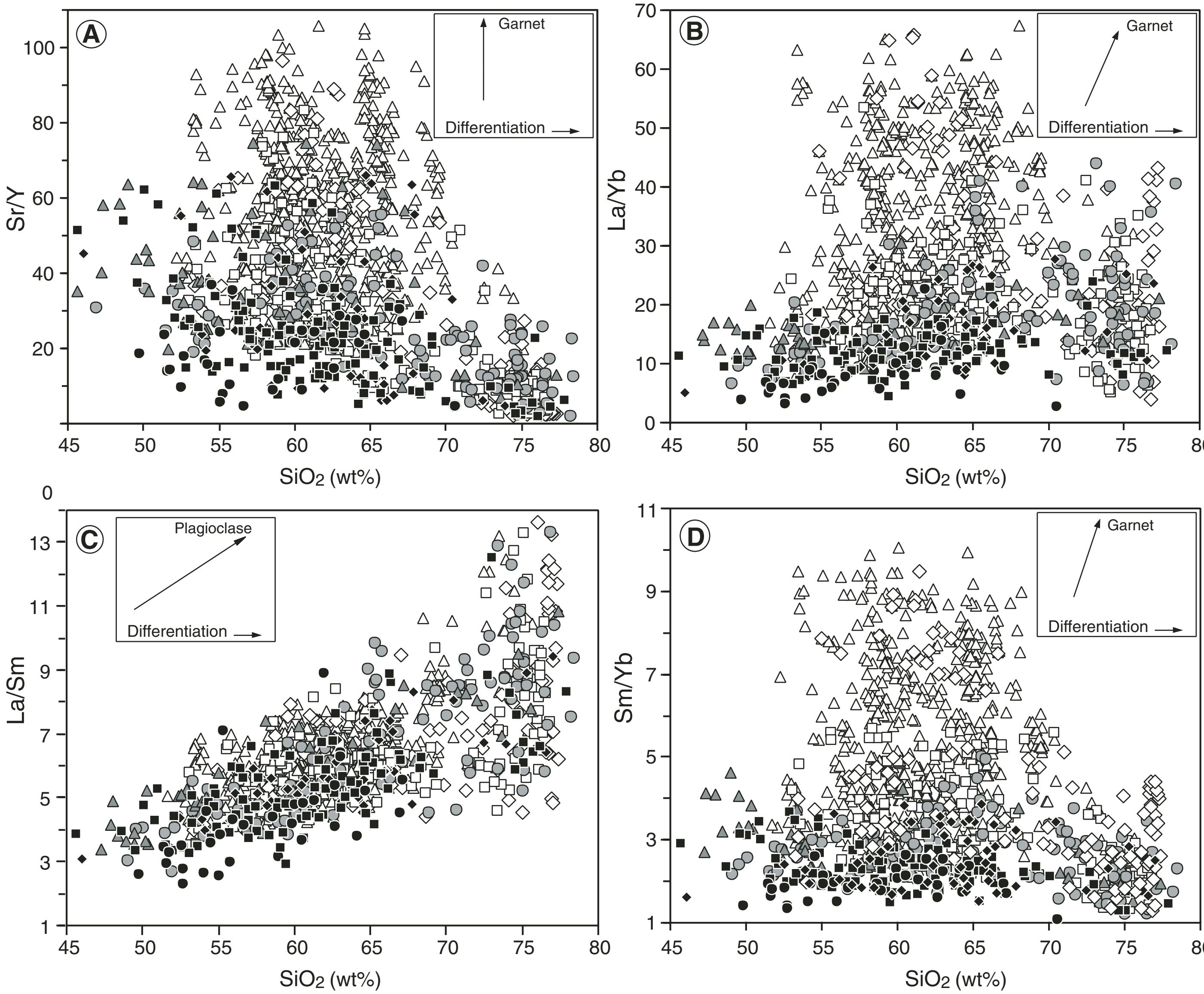
Summary of Results

Some major points directly emerge from the preliminary analysis of these data: (1) Significant variations in geochemical parameters have occurred through time (details and implications are discussed later herein). (2) In contrast, isotopic data, and in particular the Pb isotopic ratios, vary little through time but significantly between crustal domains. (3) Increases of parameters such as the Sr/Y, Sm/Yb, and Dy/Yb ratios are clearly apparent when the highest values are considered, whereas their lowest values show only a slight increase or remain more or less constant through time. This observation indicates that only higher values provide constraints on the extent of crustal thickening. (4) Variations in Dy/Yb ratios, indicative of amphibole involvement, show a significant increase at ca. 30 Ma, whereas high Sm/Yb ratios, indicative of garnet involvement, more markedly increase during the last 10 Ma.

DISCUSSION

Our data and results address two important issues for Central Andean evolution: the amount of crustal contribution to arc magmatism, and the timing of crustal thickening and related mineralogic changes involved in the generation and modification of the magmas.

Geochemical variations across or along the Andean margin (Fig. 7) have been investigated by numerous authors (James, 1982; Hildreth and Moorbath, 1988; Wörner et al., 1988, 1992; Rogers and Hawkesworth, 1989; Stern, 1990, 1991; Kay et al., 1994, 1999, 2005; Davidson et al., 1990; Trumbull et al., 1999; Haschke et al., 2002, 2003; Delacour et al., 2007; Mamani et al., 2008a, and references therein). Uncertainties about the source(s) and processes involved in Andean arc magma genesis result from the complex interplay of many factors, such as the thermal evolution of the descending slab, the release of mobile elements from sediments and hydrated basalt, the asthenospheric mantle, the lithospheric mantle, the lower and upper crust, as well as the age, composition, and thermal state of the crust. In addition, various processes are involved in the evolution of



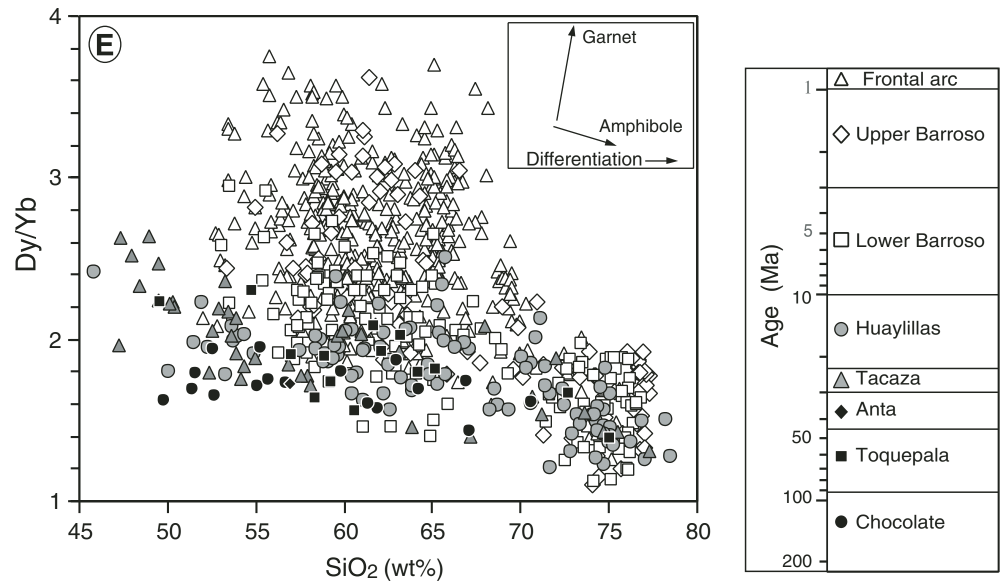


Figure 6. Plot of Sr/Y (A), La/Yb (B), La/Sm (C), Sm/Yb (D), and Dy/Yb (E) ratios versus SiO_2 content (wt%). Diagrams A, B, D, and E show for rocks with 50–65 wt% SiO_2; ratios of Sr/Y, La/Yb, Sm/Yb, and Dy/Yb ratios increase slightly from before 90 Ma into the 90–30 Ma interval, moderately from the latter into the 30–3 Ma interval, and strongly after 3 Ma. More differentiated rocks (>65 wt% SiO_2) show lower Sr/Y, Sm/Yb, and Dy/Yb ratios and higher La/Sm ratios, and thus these ratios are likely to have been affected by differentiation.

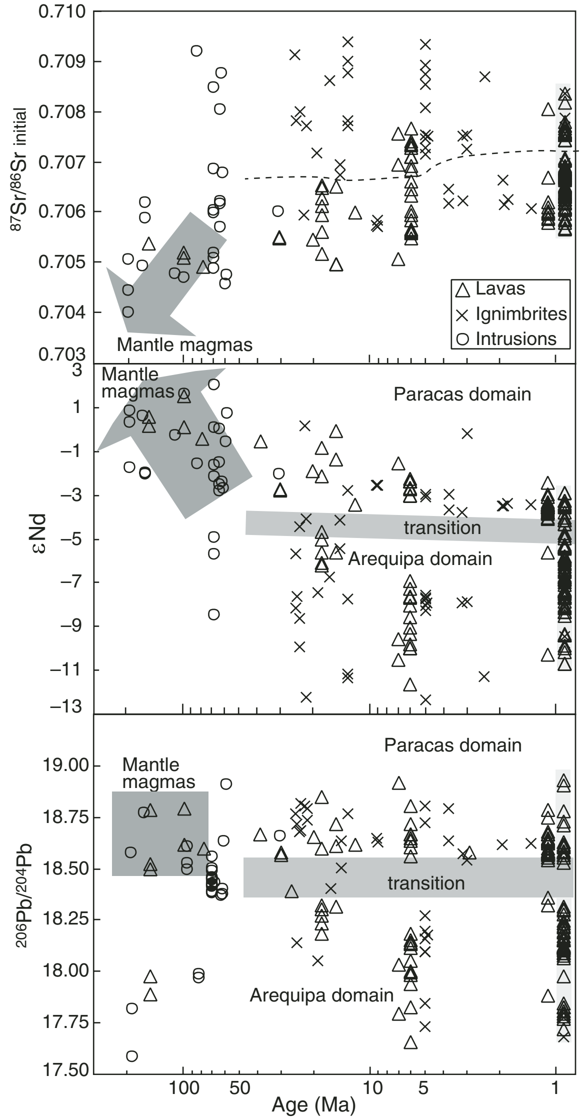


Figure 7. Plot of $^{87}Sr/^{86}Sr$ ratios, ε_{Nd} values, and $^{206}Pb/^{204}Pb$ ratios versus ages (Ma). Age ranges for volcanic arcs are as in Figure 5.

the magmas: partial mantle melting, assimilation, and fractional crystallization. This complexity has led to the development of conflicting models on Andean petrogenesis.

The following discussion assesses the different possible sources or processes thought to be responsible for the variations in major and trace elements and isotopic ratios observed along the Central Andean orocline through time and space before addressing the issue of crustal thickening.

Geochemical Enrichment: Source or Processes?

The most striking geochemical characteristic of Central Andean arc magmas is their highly variable and "enriched" isotopic composition. This could be explained by the following three hypotheses: (1) enrichment reflects variable contamination of the asthenospheric mantle wedge either by sediments subducted with the slab or by tectonic erosion of the forearc (Stern, 1990; Kay et al., 2005); (2) an ancient enriched lithospheric mantle below the Brazilian Shield was involved in Andean magmatism (Rogers and Hawkesworth, 1989); or (3) magmas generated in the mantle wedge were contaminated to different degrees through assimilation during their ascent through the thickening continental crust (e.g., James, 1982; Hildreth and Moorbath, 1988; Wörner et al., 1988, 1992; Davidson et al., 1990; Trumbull et al., 1999; Mamani et al., 2008a).

Subducted Sediments and Tectonic Erosion

There is little doubt that significant volumes of crustal rocks have been tectonically eroded from the Andean forearc (von Huene et al., 1999). The amount of this material that actually got subducted into the magma source region or was simply underplated below the forearc is a matter of debate (Stern, 1990, 1991; Clift and Hartley, 2007). Although subduction of sediments at high and variable rates has been envisaged (von Huene and Scholl, 1991), the trench facing the Central Andean orocline is almost devoid of sediments and probably has been so since onset of hyperaridity on the western Andean slope in the latest Oligocene or Miocene (Lamb and Davis, 2003; Dunai et al., 2005; Rech et al., 2006; Clift and Hartley, 2007), explaining why no accretionary prism is observed (von Huene et al., 1999). Moreover, the Coastal Cordillera has acted as a barrier limiting sediment transport into the ocean (Allmendinger et al., 2005). It is thus unlikely that significant amounts of sediments have been subducted into the source region of the "enriched" magmas.

Kay et al. (2005) proposed that tectonic erosion may have been particularly likely during arc migration and, at certain stages, could have been involved in magma genesis in the southern Central volcanic zone. Although this hypothesis is discussed later, it must be underlined here that arc migration has occurred many times during the history of the Andean margin (>400 Ma), whereas significant "enrichment" is only observed in a much more restricted geological past (≤30 Ma) and coincident with the timing of crustal thickening.

Rocks from the Central volcanic zone display neither the negative Ce anomalies nor the specific Pb compositions that would have resulted from modification of primitive arc basalts by Nazca plate sediments (Davidson et al., 1990). Entenmann (1994), Kiebala (2008), and Chang (2008) demonstrated that simple two-stage mixing is consistent with the observed variable large ion lithophile element (LILE) content and U-Th isotopes for the Parinacota and Taapaca volcanoes. According to their models, the subarc mantle wedge could first have been enriched by ~1% crustal material and fluids from the subducted oceanic crust, which produces typical arc magmas by fluid-flux melting, and these basaltic magmas would then have assimilated (lower?) crustal rocks during their storage in, and passage through, the thickened crust. This crustal assimilation is particularly well documented by high $\delta^{18}O$ values for the Parinacota, Taapaca, and El Misti volcanoes (Entenmann, 1994; Chang, 2008).

Plots of Rb/Cs and Ba/La versus age (Fig. 8) underline the apparently conflicting evidence for or against source contamination by subducted sediments, and they show that arguments based on simple trace-element ratios, which are valid for "clean" oceanic arcs, are not applicable to arc magmatism on a thick crust. Following Plank and Langmuir (1998), low Rb/Cs and Ba/La ratios should indicate involvement

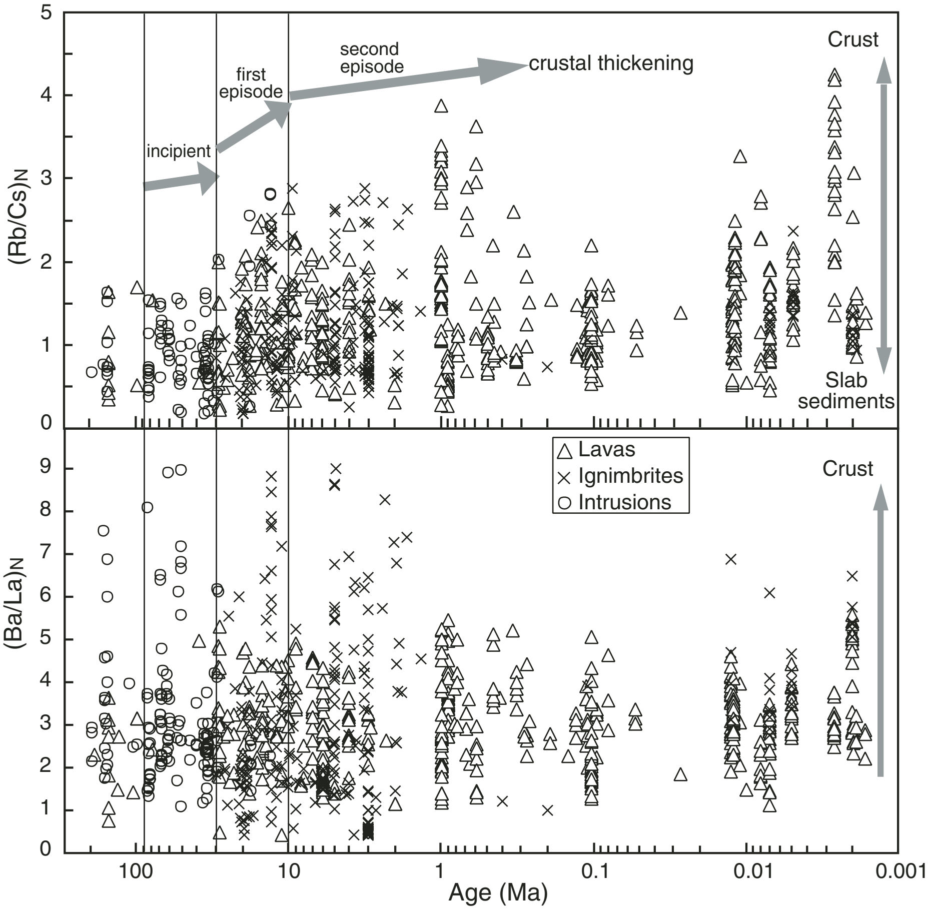


Figure 8. Log-scale plotting of the Rb/Cs and Ba/La ratios versus age (Ma) for lavas, ignimbrites, and intrusions. Arrows highlight the major intervals of changing magma compositions and increasing crustal thickness.

of Cs- and Ba-rich pelagic sediments. However, Ba/La ratios are very sensitive to alkali-feldspar fractionation. Therefore, the data concerning the most differentiated rocks (Fig. 8) have to be interpreted with caution, and thus they are not considered in this discussion. Intermediate lavas in our data set exhibit Ba/La ratios that remain approximately uniform through time. In comparisons of older, Cretaceous and Jurassic lavas and intrusions, which have low Rb/Cs ratios, to Oligocene and younger lavas, which have high Rb/Cs ratios, we see that the change from low to high Rb/Cs occurs at about the time when major crustal thickening developed (30 Ma). This was also the time of the onset of aridity (Lamb and Davis, 2003; Dunai et al., 2005, and references therein) and the decrease in the amount of sediments delivered to the trench. Prior to 30 Ma, more sediments should thus have been available for subduction into the mantle source region. The fact that Ba/La is higher in the older rocks therefore argues against the hypothesis that magmatic “enrichment” (i.e., increased Rb/Cs) was caused by involvement of subducted sediments.

Hildreth and Moorbath (1988) proposed that intracrustal contamination was the dominant process in controlling Ba/La and Rb/Cs ratios and, more generally, that variable amounts of crustal contributions due to varying crustal thickness account for the observed compositional diversity of magmas in the Southern volcanic zone of the Andes (Fig. 1). Our data confirm this conclusion in that the compositional changes we observe reflect the onset of large-scale crustal involvement starting at ca. 30 Ma, rather than the involvement of variable amounts of sedimentary components. Source contamination thus appears to have been volumetrically small in the Central Andes (Entenmann, 1994; Chang, 2008).

Enriched Subcontinental Lithosphere

Because in northern Chile at ~22°S, abundances of incompatible elements, Sr concentrations, and $^{87}Sr/^{86}Sr$ ratios all increase from west to east, Rogers and Hawkesworth (1989) proposed that Central Andean magmas have been mainly generated by partial melting of an ancient enriched subcontinental lithosphere. However, the fact that $^{87}Sr/^{86}Sr$ ratios and Sr contents increase with time at a rate similar to that of crustal thickening (Fig. 9) cannot be easily explained by this hypothesis. Furthermore, O and Pb isotopic ratios provide convincing evidence against the idea that at ~18°S, Andean rocks were dominantly generated by melting of an enriched mantle lithosphere (Davidson, 1991).

In the eastern Altiplano of southern Peru and Bolivia, i.e., in a marked backarc position, Pliocene-Quaternary mantle-derived ultrapotassic mafic rocks exhibit highly variable bulk compositions and are likely to have originated by melting of a heterogeneous lithospheric mantle (Carlier et al., 2005). However, this kind of lithospheric mantle does not appear to be present below the main arc (Allmendinger et al., 1997).

Crustal Thickening and Assimilation through Time

In most analyzed rocks, La/Yb and Sr/Y ratios are higher for rocks younger than 30 Ma, and they are related to crustal isotopic signatures, strongly suggesting that crustal thickness played a key role in the processes that fractionated REE (Hildreth and Moorbath, 1988). Similar temporal variations were observed by Haschke et al. (2003). Wörner et al. (1988), Davidson et al. (1991), and McMillan et al. (1993) showed that at 18°S, arc lavas younger than 7 Ma have unusually high Sr concentrations and show a depletion in HREEs compared to the older (>7 Ma) magmatic products. These authors took this observation to indicate higher pressure of assimilation in the crust resulting from crustal thickening prior to that time, but they did not distinguish the details of the changing HREE depletion over the interval 30–7 Ma. Their interpretation is in general agreement with our conclusions here and with views expressed by Hildreth and Moorbath (1988), Kay et al. (1994, 1999, 2005), and Haschke et al. (2002, 2003).

The Sr/Y, La/Yb, and La/Sm ratios are sensitive to plagioclase fractionation, whereas the Dy/Yb and Sm/Yb ratios are less affected (Fig. 6). Therefore, we concentrate here on Dy/Yb and Sm/Yb to constrain the amount and timing of crustal involvement before and during thickening, since these parameters, which respectively track the presence of amphibole and garnet residues, show particularly marked variations with time (Fig. 10).

The marked increases in Dy/Yb and Sm/Yb ratios displayed by lavas younger than 30 Ma (Figs. 9 and 10) thus reflect increasing lithostatic pressure in zones of differentiation and

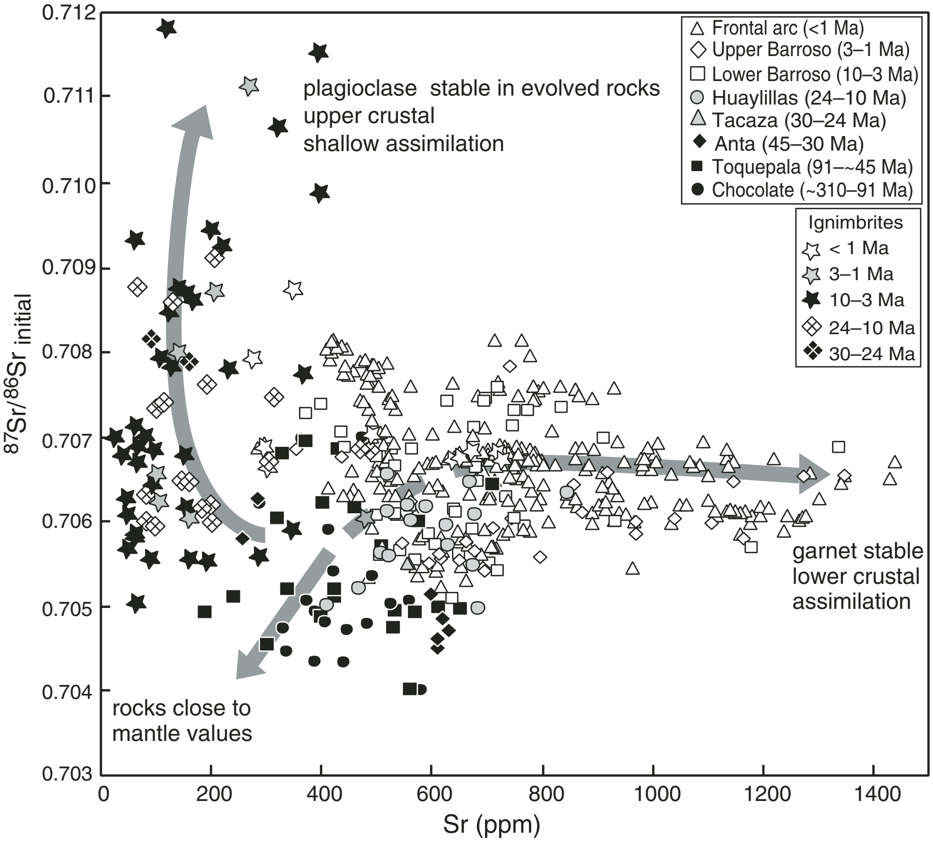


Figure 9. Initial $^{87}Sr/^{86}Sr$ ratios versus Sr concentrations of magmatic rocks in the Central Andean orocline.

crust-magma interaction (at depths where amphibole and garnet develop), as a consequence of crustal thickening (see also Davidson et al., 1991). In contrast, rocks older than 30 Ma display lower $^{87}Sr/^{86}Sr$, Dy/Yb, and Sm/Yb ratios. Rocks older than 91 Ma have even lower $^{87}Sr/^{86}Sr$, Dy/Yb, and Sm/Yb ratios, suggesting that the Andean crust was relatively thin before that time. Rocks emplaced or erupted between 91 and 30 Ma show moderate Dy/Yb fractionation and low values but little change in Sm/Yb, thus reflecting a possible involvement of amphibole but not of garnet, confirming that only minor thickening of the crust can have occurred during this time interval. In contrast, lavas erupted between 30 and 3 Ma show both Dy/Yb and Sm/Yb fractionation, indicating development of garnet in the lower crust and thus confirming they were generated during major crustal thickening.

Lavas younger than 3 Ma exhibit even higher Dy/Yb and Sm/Yb ratios, thus suggesting further increased crustal thickening or a migration of magma storage and assimilation to deeper levels in the crust. Because the present crustal thickness in the arc region of the Central Andean orocline (≥65 km; Yuan et al., 2002) implies pressures of ~2.1 GPa near the base of the crust, garnet should be stable in the lower crust, and arc magmas interacting with it may acquire a "garnet signature." However, as stated before, magmas with relatively low Sm/Yb and Dy/Yb ratios existed at any given time, indicating that low-pressure signatures persisted after the crust had significantly thickened, and thus that the significant parameters for crustal thickness are in fact the maximum Dy/Yb and Sm/Yb ratios displayed by an age group. Persistence of low-pressure signatures in magmas only reflects that the assimilation and differentiation of these particular magmas took place at shallow crustal levels, and that they did not significantly interact with the higher-pressure parts of the crust dominated by garnet or amphibole. Therefore, isolated observations of low Sm/Yb (or Dy/Yb) ratios in a given magmatic suite cannot be taken as evidence for a thin crust. More generally, isolated high Dy/Yb and Sm/Yb ratios cannot just be taken as simple and direct proxies of crustal thickness (see details in the following).

High Sm/Yb and Dy/Yb ratios are observed dominantly in the intermediate compositional range (andesites), whereas more mafic and more differentiated magmas show less extreme values (Fig. 6). Furthermore, differentiation can be assumed to have mainly taken place within the continental crust, and the garnet signature (high Sm/Yb and Dy/Yb) is associated with a shift toward crustal isotope composition. It is thus unlikely that the observed garnet signatures result from melting garnet-bearing lithologies in the mantle and/or subducted slab.

We emphasize that the garnet and amphibole residues may reflect either residues in crustal rocks during high-pressure melting and assimilation or crystallization of garnet and amphibole directly from basaltic to andesitic magmas (e.g., Macpherson et al., 2006, and references therein). However, since there is a correlation between increasing "crustal" isotopic signatures and HREE depletion, it is argued that these two processes should be strongly related (Kay et al., 1999).

Concurrent with the observed shifts in trace-element signatures in arc andesites, we observe that the largest ignimbrite volumes (representing more differentiated rocks, >65 wt% SiO_2) erupted at specific times (mainly ca. 24 Ma, ca. 20 Ma, ca. 14 Ma, ca. 10 Ma). Only smaller, medium-sized (typically <300 km^3) ignimbrite volumes were erupted between 5 and 1 Ma. At any given time, however, ignimbrites are distinct from spatially associated andesites in major and trace elements (Figs. 5, 6, 8, 9, and 10) and are more radiogenic in Sr isotopes (Fig. 9). Maximum Sr-isotope values (up to 0.709) and lower Sr concentrations are observed in the younger ignimbrites, which suggest plagioclase (i.e., low-pressure) fractionation. Only a few more recent and volumetrically minor (<5 km^3) ignimbrites have relatively low Sr-isotope values and other geochemical characteristics. Because the older large plateau-forming ignimbrites (24–10 Ma) display low Sr-isotope values, they reflect crustal melting of less enriched (lower-crustal) lithologies and/or involvement of a larger component of mantle-derived magma. Ignimbrites with higher Sm/Yb ratios also have higher Sr-isotope values, suggesting that isotopic compositions and trace-element signatures are correlated.

The marked increases in maximum Sm/Yb ratios exhibited by the Tacaza (30–24 Ma) and Upper Barroso (3–1 Ma) lavas, respectively, reflect the two successive changes in magma genesis during crustal thickening. The first change at ca. 30 Ma indicates a significant involvement of amphibole as a major mineral phase in the lower-crustal zone where magmas differentiated and were modified by crustal assimilation. This change was coeval with the uplift documented in southern Peru and Bolivia (e.g., Picard et al., 2008, and references therein). The second change, documenting involvement of garnet in the lower crust partly at the expense of amphibole in the Pliocene, indicates that crustal thickening continued into the late Neogene. However, the time at which this second mineralogical threshold was crossed, as documented by the Sm/Yb maximum ratios, is younger by ~7 Ma than the onset of the second major episode of crustal thickening as documented by thermochronologic and paleoaltimetric studies (e.g., Garzione et al., 2006, 2008; Schildgen et al., 2007; Thouret et al., 2007). This observation has two general implications and a third implication that is specific to the evolution of the magma systems of individual stratovolcanoes: First, the temporal delay between thickening and the increase of the (maximum) Sm/Yb ratios in Andean andesites may indicate either that the zone where magmas are modified migrated to deeper crustal levels after the second stage of crustal thickening, or, alternatively, that heating of the lower crust to allow assimilation there required this duration of time. Second, the fact that magmatic products remained geochemically relatively uniform when a major episode of crustal thickening and uplift took place is a strong argument against delamination as a cause for this uplift (see discussion in Hartley et al., 2007), since this would have led to a major change in the chemistry of magmas derived from the mantle. Third, and more specifically, the observation that many stratovolcanoes still do not show elevated Sm/Yb ratios, even in the youngest rocks, and thus at maximum crustal thickness, can be exploited to identify the depth of their principal level of crust-magma interaction (see further discussion on this later herein).

The production of ignimbrites that results from progressive thermal and mechanical maturation of the crust (de Silva et al., 2006) indicates that sufficient heat to induce crustal fusion could have been achieved by invasion of the crust by hot basic magmas, as was the case during the Tacaza interval (see previous discussion), and thus ultimately by the invasion of asthenospheric mantle into the mantle wedge during slab steepening. In the Central Andean orocline, large volumes of ignimbrites were erupted between ca. 75 and ca. 65 Ma (Toquepala arc), i.e., during incipient crustal thickening, and after initiation of major crustal thickening (mainly ca. 24 Ma, ca. 20 Ma, ca. 14 Ma, ca. 10 Ma), suggesting that crustal thickening might have been partly related to the production of voluminous mantle-derived magmas and their transfer into the crust. Since plateau-forming ignimbrites have younger ages toward the south in northern Chile (19–11 Ma at 17–20°S, but 14–16 Ma at 22°S), crustal thickening, or at least the processes that resulted in massive crustal melting, may have developed later in the south.

No Simple Links between REE Patterns and Crustal Thickness

From the previous discussion, it should not be concluded that any magma ascending through a thickened crust necessarily displays HREE depletion. Magma-crust interaction in "Mixing, Assimilation, Storage, and Homogenization"-zones (Hildreth and Moorbath, 1988) is not restricted to deep crustal sections, even, and perhaps especially, in the case of a thick crust. Magmas traversing a thick crust do not necessarily interact extensively with deep garnet- or amphibole-bearing rocks. Our data clearly demonstrate that many magmas from epochs of incipient (90–30 Ma) or major (<30 Ma) crustal thickening do *not* exhibit high Sr/Y, La/Yb, La/Sm, Sm/Yb, and/or Dy/Yb ratios (Fig. 10). In fact, it is only the increase in the *maximum* values of these ratios that constitutes a clear indication of progressive crustal thickening, whereas low values from coeval rocks do not preclude the existence of a thick crust. Hence, confident conclusions regarding crustal thickness drawn from such ratios can only be based on a significant number of samples. Whereas only one sample with a high Sm/Yb or Dy/Yb ratio is sufficient to indicate a somewhat thickened crust, one sample with low values of these ratios cannot be interpreted as indicative of an unthickened crust.

Concentrating thus on the evolution of maximum trace-element ratios, we observe that these exhibit significant increases at ca. 30 Ma, ca. 3 Ma, and <1 Ma. This finding is similar to results reported by Haschke et al. (2002) and Haschke and Günther (2003), and it is thus unlikely to result from an artifact of the data set or sampling. However, Haschke et al. (2002) and Haschke and Günther (2003) used HREE depletion as a proxy for crustal thickness and thus suggested that crustal thickness may have increased and decreased repeatedly through time in the Central Andes. Because we document here that many recent lavas show no pronounced HREE depletion, in spite of having undoubtedly traversed a thick crust (Fig. 11), we underline that directly relating REE ratios in individual rocks to crustal thickness at the time of eruption is not warranted.

In a W-E traverse of the southernmost Central volcanic zone, where crustal thickness remained constant or even increased, Kay et al. (2005) reported an increase in La/Yb and Sm/Yb ratios followed by a subsequent decrease. Because they observed that the age and location of HREE-depleted magmas were spatially related, they proposed that a migration of the arc, induced by tectonic erosion of the leading edge of the continental margin, had caused the zone of magma-crust interaction to shift from deep crustal sections, which had been depleted by earlier crustal melting, to "fertile" crust that allowed increased assimilation at high pressure. Unlike Kay et al. (2005), we find in the Central Andean orocline that spatial and temporal variations in geochemistry have been decoupled. We observe, however, that some regions of thick crust show stronger maximum HREE depletions than others at a given time (Figs. 10 and 11). Rather than crustal thickness or tectonic erosion alone, we therefore propose that three factors may actually control HREE depletion: (1) a thick (>~45 km) crust, which is a prerequisite for the formation of HREE-depleted magmas; (2) the depth of magma-crust interaction, which may be deep or shallow, even in a "thick crust" setting; and (3) the bulk composition of the crust and the capacity to yield HREE-depleted partial melts as assimilants. This is because garnet stability and abundance and the degree of crustal melting also depend on bulk rock composition and not only on pressure and temperature. This combination of factors may explain why recent arc lavas erupted through the Arequipa domain crust (Fig. 4) are generally much more HREE-depleted (higher Sr/Y) than those to the south and north (Fig. 11). The Arequipa domain is characterized by a thicker crust (Yuan et al., 2002) and may be more mafic than that of the Paracas and Antofalla domains (Mamani et al., 2008a). However, there is no doubt that the Paracas and Antofalla crust is thick enough to stabilize garnet in its lower part.

These factors are schematically illustrated in Figure 11 with respect to variable Sr/Y and Sm/Yb ratios. Only Quaternary rocks with <65 wt% SiO_2 are plotted, and thus the effect of plagioclase fractionation is excluded. Although trace-element ratios are high and the crust is sufficiently thick to stabilize garnet in its deeper section, the highest Sr/Y and Sm/Yb ratios are observed in young volcanic rocks that erupted through the (mafic) Arequipa domain. This diagram thus distinguishes between the effect imposed by the bulk composition of the crust on the amphibole to garnet transition (affecting Sm/Yb) and the presence or absence of plagioclase (affecting only Sr/Y). High Sm/Yb and Sr/Y ratios thus represent garnet-bearing and plagioclase-free (i.e., eclogitic) residues for rocks from the Arequipa domain, whereas high Sm/Yb and low Sr/Y ratios reflect stability of both garnet and amphibole, i.e., garnet-amphibolite residues in the Paracas and Antofalla domains. The fact that the Sm/Yb ratio increases earlier than the Sr/Y ratio in the Arequipa domain may reflect an evolution from (1) garnet-free to (2) garnet-amphibolite and (3) eclogitic residues.

Although the Lower Barroso lavas were erupted during rapid surface uplift between 10 and 7 Ma, these lavas exhibit REE patterns and

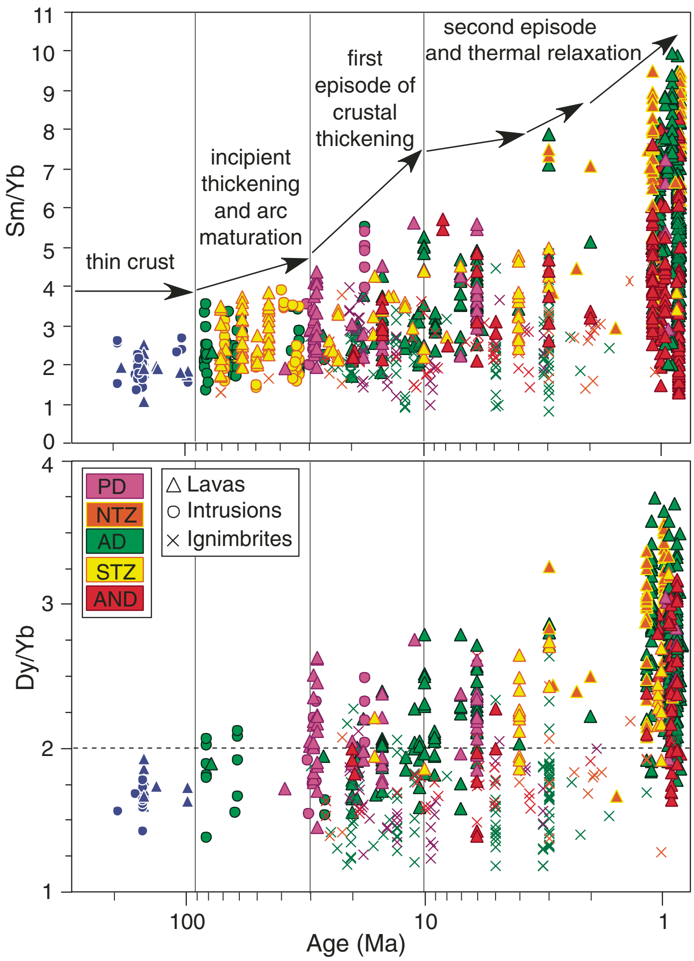


Figure 10. Plot of Sm/Yb and Dy/Yb ratios versus rock ages (note the logarithmic age axis). Arrows in upper diagram highlight the different episodes of crustal thickening. Legend: PD—Paracas domain, NTZ—Northern transition zone, AD—Arequipa domain, STZ—Southern transition zone, and AND—Antofalla domain. The samples from Chocolate and La Negra Formations are in blue. The dashed horizontal line in the Dy/Yb diagram marks the boundary between garnet-dominated (>2) and amphibole-dominated (<2) compositions. Low values of the Dy/Yb ratio indicative of amphibole involvement, show a significant increase at ca. 30 Ma and afterward, whereas higher values of the Sm/Yb ratio, which is indicative of garnet involvement, have notably increased during the last few million years.

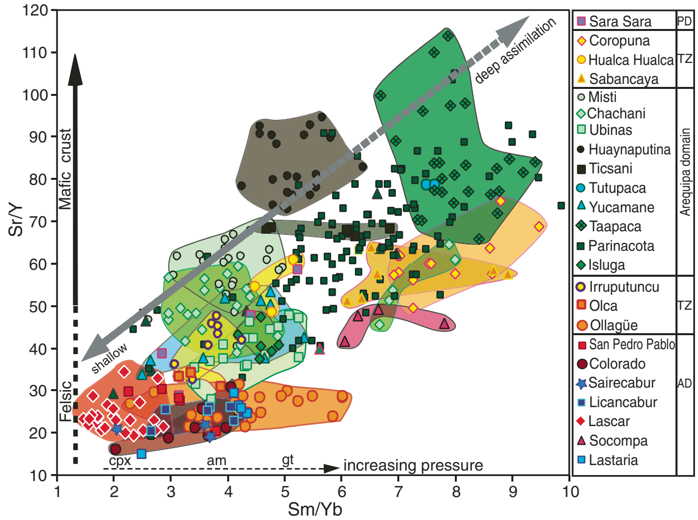


Figure 11. Sm/Yb versus Sr/Y variations for Quaternary volcanoes in the Central Andean orocline. Abbreviations for crustal domains are as in Figure 10. See Figure 4 for location of volcanoes. Data plotted are from lavas with <65 wt% SiO_2 (Figs. 6A and 6D). The Sr-, Nd-, and Pb-isotope ratios for each volcano appear in Figure 12, and the amount of crustal contamination is estimated in Figure 13.

maximum HREE depletions similar to those of the 24–10 Ma lavas (Fig. 10). If, as we argue, this surface uplift resulted from crustal thickening rather than delamination, this means that the magmatic expression of a trace-element signature of further crustal thickening was delayed. As detailed already, there are several possible (or combined) explanations for this, including: (1) changing magma conduits and stagnation reservoirs during thickening did not allow deep, high-pressure magma evolution, or, more simply, (2) a thermal relaxation time in the middle to deep crust after crustal thickening must be overcome before the garnet signature by assimilation is formed.

Assimilation Reflects Compositional Domains of the Andean Crust

In addition to the overwhelming time control on trace-element signals, regional isotopic patterns in crustal domains are detected (see also Macfarlane et al., 1990; Wörner et al., 1992; Aitcheson et al., 1995; Tosdal, 1996; Lucassen et al., 2001; Loewy et al., 2004; Mamani et al., 2008a).

The most striking boundary is defined by isotopic contrasts between the Andahua and Huambo volcanic fields, which are located north and south, respectively, of the northern boundary of the Arequipa domain (Figs. 4, 12, and 13). These fields are both characterized by an abundance of monogenetic cinder cones and lava flows, many of which are relatively mafic and probably erupted through deep-reaching crustal faults (Delacour et al., 2007). The

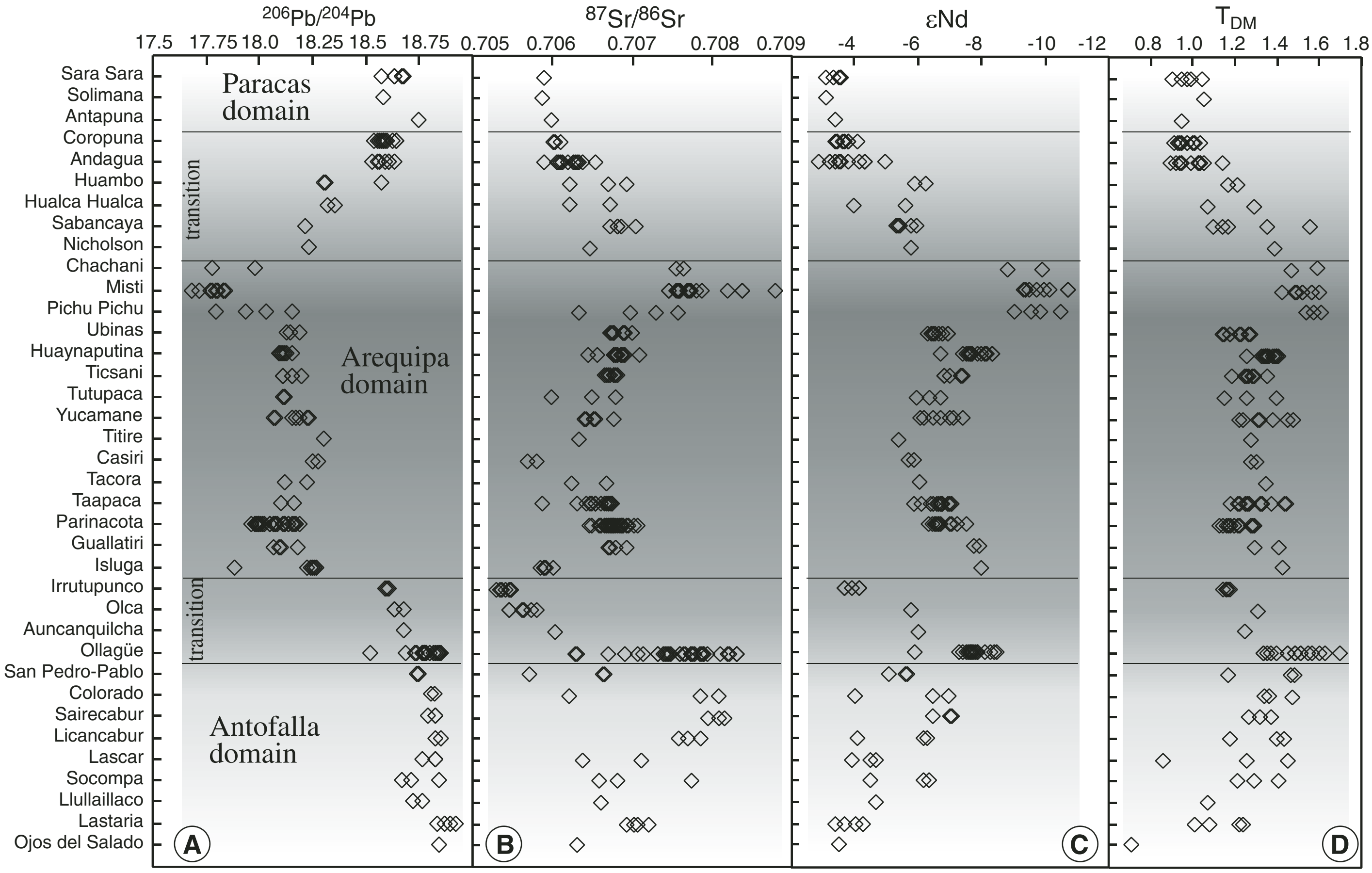


Figure 12. Variations in $^{206}Pb/^{204}Pb$, $^{87}Sr/^{86}Sr$, ε_{Nd}, and Sm-Nd model ages (T_{DM}) for rocks from Quaternary volcanoes, listed from north (top) to south (bottom). These variations reflect the influence of crustal domains along strike. The crust of the Paracas and Antofalla domains (light-gray fields) is dominantly felsic, whereas that of the Arequipa domain (dark-gray field) is dominantly mafic (see also Fig. 4).

compositional change of monogenetic centers across this boundary coincides at the surface with the Iquipi-Trigal crustal fault system, which is seismically active (David, 2007). The boundary delineated by distinct Pb isotopic compositions is remarkably abrupt (within a distance of 60 km, over a >60-km-thick crust; Mamani et al., 2008b), suggesting that the boundary is rather steep. It is therefore likely that this boundary represents a major and active suture between distinct crustal blocks.

A series of minor volcanic centers also occurs along the southern boundary of the Arequipa domain. Other examples include monogenetic centers along the West-Fissure fault in northern Chile (Deruelle, 1982), which marks the eastern boundary of the Coastal Cordillera domain (Mamani et al., 2008a). Minor volcanic centers possibly preferentially formed along crustal boundaries in the Central Andean orocline.

The depleted mantle (T_{DM}) ages of andesites from Quaternary volcanoes confirm the definition of crustal domains (Fig. 12F). North of 16°S, T_{DM} ages vary between 0.8 and 1.3 Ga; between 16°S and 21°S, T_{DM} ages range from 1.0 to 1.8 Ga; and south of 21°S, T_{DM} ages vary from 0.7 to 1.4 Ga. The Arequipa Proterozoic basement has T_{DM} ages between 1.9 and 2.3 Ga (Loewy et al., 2004). In the southern Central Andean orocline, the Paleozoic basement has T_{DM} ages between 0.9 and 1.8 Ga (Lucassen et al., 2001). The good correlation among Sm-Nd model ages, Pb isotopic ratios, and trace-element signatures of Andean magmas indicates that these have derived their assimilated component from crustal sections of different age and composition (Fig. 4).

Quantifying Crustal Contributions

Although many uncertainties exist about the composition and structure of the continental crust, all data presented here are most consistent with models involving intracrustal assimilation in MASH zones, as proposed by Hildreth and Moorbath (1988), Wörner et al. (1988, 1994), Davidson et al. (1990, 1991), and Mamani et al. (2008a), among others. To estimate the amount of crustal material assimilated by Andean magmas, we calculated mixing models (following Albarède, 1996) between a mantle-derived magma and local crust, based only on the Sr-, Nd-, and Pb-isotopic ratios. The composition of the mantle-derived magma is difficult to assess. In the absence of tighter constraints, we chose a juvenile basaltic andesite of Jurassic age simply because no primitive lava compositions are observed in the younger rock suites. Assimilants are represented by compositional data from the Mollendo Proterozoic basement (Arequipa domain) and the Paleozoic Cotahuasi (Paracas domain) and Choja (Antofalla domain) basements.

We find that the amounts of crustal contamination in andesites range from 7% to 18%. In the Arequipa domain, the most-contaminated lavas are from the Chachani, El Misti, and Pichu Pichu volcanoes. In the Antofalla domain, they are from the Sairecabur, Licancabur, and Socompa volcanoes (Fig. 12). The least-

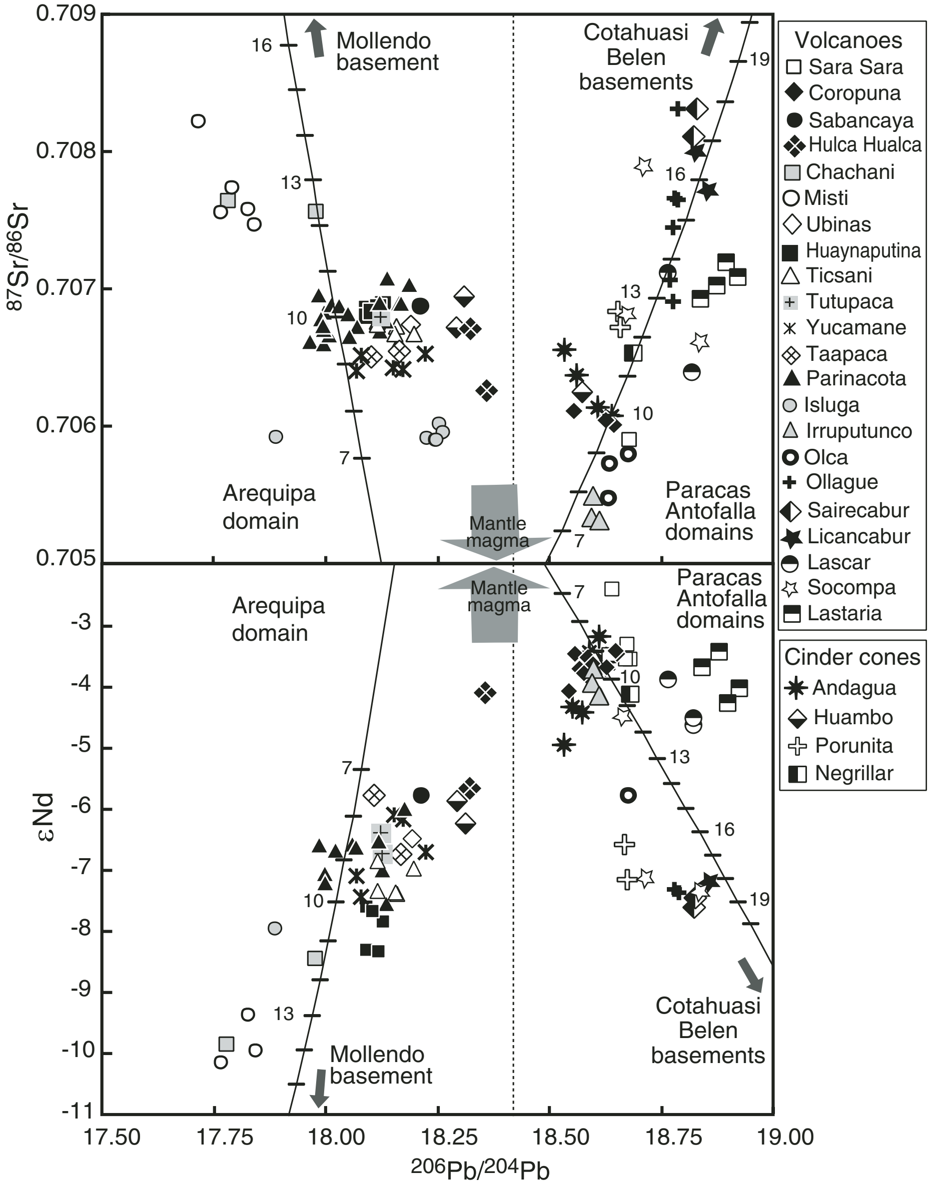


Figure 13. ^{87}Sr/^{86}Sr ratios, ε_{Nd} values, and ^{206}Pb/^{204}Pb ratios for Quaternary volcanoes from different domains and cinder cones from transition zones. Lines show result of a simple mixing model (Albarède, 1996) of a Jurassic lava close to a mantle source (Sr, Nd, and Pb contents [ppm] and isotopic ratios are 111/0.7033, 13/0.5127, 4/18.221, respectively) with the Paleozoic Cotahuasi basement (Sr, Nd, and Pb content/ratios are 96/0.735, 31/0.51145, 11/20, respectively) and the Proterozoic Mollendo basement (Sr, Nd, and Pb content/ratios are 150/0.73, 54/0.51128, 7/17, respectively). Tic marks are percentages of crustal contamination.

contaminated lavas occur in the transition zones (e.g., Coropuna, Hualca Hualca, Irruputunco, Olca, and Auncanquilcha volcanoes).

To better constrain the assimilation process, we calculated the La/Yb and Sm/Yb ratios of crustal melts (Albarède, 1996) using mafic assimilants with different proportions of residual minerals (plagioclase, amphibole, garnet) at variable degrees of partial melting. As a potential source, we used a Jurassic basaltic andesite from the Chocolate arc (La = 10.5, Sm = 2.3, Yb = 1.4, Y = 10.6). Source mineralogies vary from a garnet-free amphibolite (amphibole/plagioclase = 75/25) to a garnet-bearing amphibolite (amphibole/garnet = 90/10) and eclogite (amphibole/garnet/clinopyroxene = 10/40/50) in order to account for increasing crustal thickness. Distribution coefficients (K_D) were taken from Rollinson (1993). Such rocks have been postulated to make up a large part of the lower arc crust in parts of southern Peru (Petford and Atherton, 1996). Taken at face value, the observed high La/Yb and Sm/Yb compositions observed in young Andean rocks (Fig. 14) could be explained simply by partial melting of garnet-bearing basaltic lower crust. However, the degree of melting would have to be at least 75% from a plagioclase-free residue reflecting ~2.1 GPa, i.e., ~60 km crustal thickness. As such a high degree of melting is unlikely, and as the absence of any mantle-derived magma in any of the younger volcanic rocks in this arc setting is not reasonable, we infer from this modeling that high La/Yb and Sm/Yb ratios were not achieved by direct partial crustal melting but rather resulted from mixing of mantle-derived magmas with up to 15%–18% of low-degree (<20%) partial melts from mafic crustal lithologies at high pressure.

CONCLUSION

Central Andean arc magmas are characterized by an overall enriched geochemical composition with significant variations in time and space. Because neither subducted sediments and tectonic erosion nor the presence of an enriched subcontinental lithosphere are likely causes of the observed enrichment and variations, these are best explained by interaction of mantle-derived arc magmas with a mature composite crust during different stages of crustal thickening. However, even though the changes in magma composition with time are significant, the geochemical signal is by no means simply and directly related to crustal thickness.

Because a growing lithostatic pressure in the lower crust produces mineralogical changes that induce fractionations of trace elements in melts with which the residue interacts, our analysis of a comprehensive set of geochemical data on Andean magmatic rocks sheds an independent light on the timing of crustal thickening and thus advances key issues concerning the evolution of the Central Andean orocline. High Dy/Yb ratios indicate interaction of magmas with an amphibole-rich crust, which significantly increased at ca. 30 Ma in the Central Andean orocline. High Sm/Yb ratios indicate interaction with a garnet-rich crust, and these have more markedly increased since ca. 10 Ma.

However, HREE depletion occurs only in the case of significant magma-crust interaction with a lower crust that has a composition suitable for growing amphibole and/or garnet with increasing pressure: the relatively unfractionated REE profile of many volcanic rocks erupted through a thick crust indicate that the

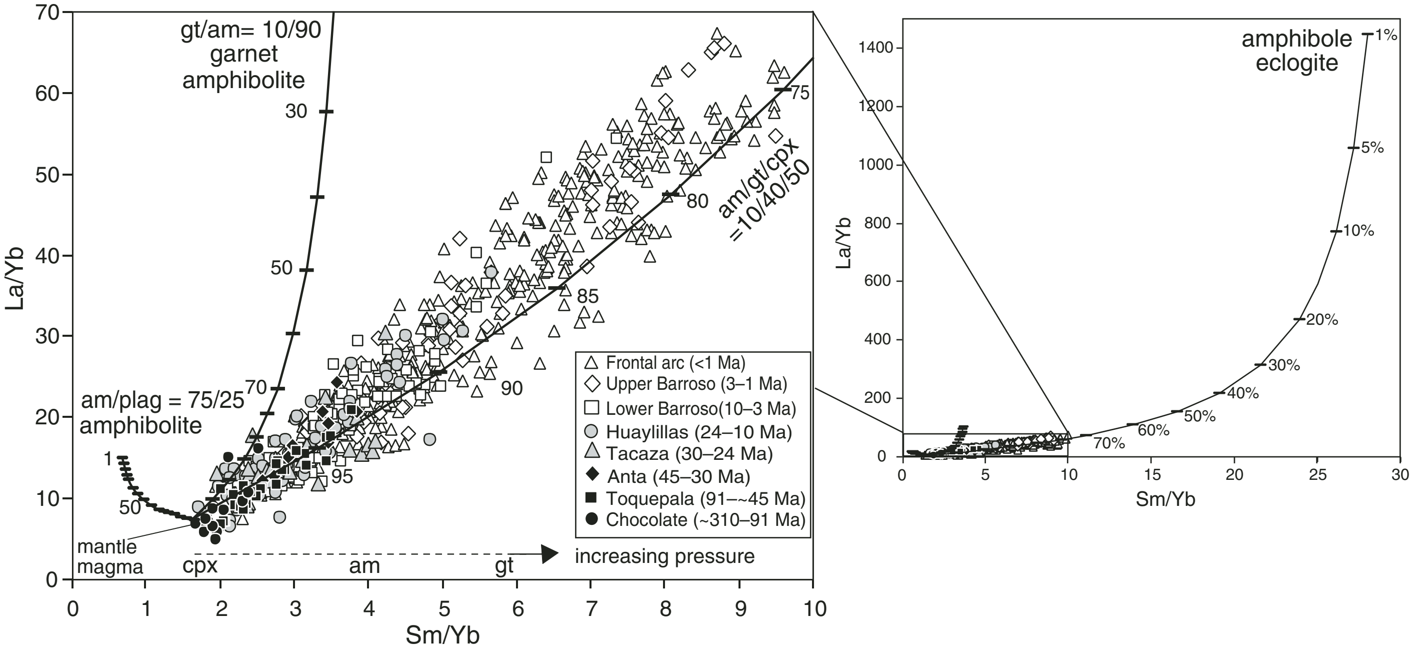


Figure 14. La/Yb versus Sm/Yb diagram showing the evolution of arc rocks (<65 wt% SiO_2); cpx—clinopyroxene, am—amphibole, gt—garnet. Curves show result of batch-melting model (Albarède, 1996) of amphibolite, garnet amphibolite, and amphibole eclogite.

corresponding magmas did not interact significantly with lower-crustal lithologies. The relation between crustal thickness and trace-element signatures is thus only general and not simple and direct, and only variations of the maximum values of parameters such as the Sr/Y, Sm/Yb, and Dy/Yb ratios are informative on crustal thickening.

For the Chocolate interval (ca. 310–91 Ma), low $^{87}Sr/^{86}Sr$ ratios and high ε_{Nd} values indicate that magmas were primarily derived from the mantle, and they either traversed a juvenile crust or suffered only a very small degree of assimilation of mature crust. Low Dy/Yb and Sm/Yb ratios suggest that the crust was thin. Distinct Pb-isotopic ratios indicate either derivation from a mantle source and/or contamination by Arequipa or Mejillonia basement rocks. These data confirm that the Central Andean margin dominantly underwent tectonic stretching and lithospheric thinning during this time interval.

A slight increase in Dy/Yb and Sm/Yb ratios corroborates incipient crustal thickening developing from the Late Cretaceous to the mid-Oligocene (91–30 Ma). Higher $^{87}Sr/^{86}Sr$ ratios and lower ε_{Nd} values reflect a higher degree of assimilation and contamination by more mature crustal material. Pb-isotopic ratios suggest that magmas were derived from mantle sources and were contaminated by local basement. It is noteworthy that at ca. 45 Ma, the arc migrated north by up to 200 km in western southern Peru.

The major crustal thickening typical of the present Central Andes has developed since 30 Ma, while, in southern Peru, the main arc has migrated back toward the trench. The steepening of the slab associated with this back-migration induced mantle upwelling and decompression, which resulted in pervasive mantle melting and widespread, voluminous high-K mafic magmatism during the interval 30–24 Ma. The influx of hot asthenosphere into the mantle wedge and the related invasion of the crust by large amounts of hot mantle-derived melts triggered crustal melting, which resulted in massive ignimbrite eruptions, especially between ca. 26 and ca. 18 Ma. The marked increase of Sm/Yb and Dy/Yb ratios displayed by 30–24-Ma-old rocks indicates that magmas interacted and thermally equilibrated with a lower crust in which garnet and amphibole were a major stable residual phase, coincident with the onset of significant crustal thickening and related to uplift in the late Oligocene.

A second and major episode of surface uplift developed in the late Miocene (ca. 10–6 Ma; Garzione et al., 2006, 2008; Schildgen et al., 2007; Thouret et al., 2007), and our analysis of the data provides insights into its cause. Although Dy/Yb and Sm/Yb ratios of lavas that erupted during the 10–3 Ma interval display no clear increase, volcanic rocks younger than 3 Ma show very high Dy/Yb and Sm/Yb ratios, pointing to garnet as the only major residual phase controlling REE, and thus indicating that crustal thickening had continued to the point of imposing very high lithostatic pressure in the lower crust. The delay observed between the inferred resumption of crustal thickening and the eruption of lavas with a marked garnet signature is possibly due to a relaxation time of a few million years before this signature was formed, and/or to magma stagnation at shallower crustal levels before final eruption. Initiation of this uplift episode coincided with the resumption of ignimbrite flare-ups at ca. 10 Ma, reflecting the initiation of a new major episode of crustal melting and thus suggesting intrusion of hot magmas into the crust, in agreement with the eruption of high-K mafic magmas southwest of Lake Titicaca between 6 and 5 Ma (Kaneoka and Guevara, 1984) and more regionally between ca. 9 and ca. 4 Ma (M. Fornari and T. Sempere, 2008, personal commun.).

The $^{87}Sr/^{86}Sr$ and $^{206}Pb/^{204}Pb$ ratios and ε_{Nd} values of volcanic rocks younger than 30 Ma document that assimilation and contamination by the Central Andean basement markedly increased during crustal thickening. Volcanic rocks erupted in the Arequipa domain (and the Paracas and Antofalla domains) have been contaminated by a crust with T_{DM} ages between 1.8 and 1.0 Ga (between 1.4 and 0.7 Ga for the Paracas and Antofalla domains, respectively). The amount of continental crustal contribution in recent arc andesites is estimated to range between 7% and 18%. The least-contaminated lavas are from volcanoes located near the crustal boundaries delineated by Mamani et al. (2008a), where minor volcanic centers preferentially formed, documenting that these boundaries, which are seismically active, represent major sutures between crustal blocks.

The association of "crustal" isotopic signatures and HREE depletion increasing with time indicates that major crustal thickening was initiated in the mid-Oligocene and has continued until the present. Although tectonic shortening is widely accepted as the main cause of crustal thickening in the Central Andean orocline, little or no shortening has occurred in the arc region since >10 Ma (Sempere and Jacay, 2007, 2008). Since no shortening could have caused the major and rapid uplift documented between ca. 10 and ca. 6 Ma (Garzione et al., 2006, 2007; Sempere et al., 2008; Hartley et al., 2007), it has been proposed that uplift was triggered by delamination of dense lithospheric material into the asthenosphere (Garzione et al., 2006, 2008; Garzione et al., 2007). However, no magmatism typical of delamination process is known in the concerned region (this study; Hartley et al., 2007). Furthermore, delamination cannot thicken the crust—and may even thin it—and this hypothesis is thus clearly inconsistent with the data discussed here. Because tectonic shortening is an unlikely cause in this particular case, our work thus raises again the issue of the process(es) responsible for the latest stage of crustal thickening observed in the Central Andean orocline. Because arc magmatism appears to be insufficient to significantly thicken the crust (Wörner et al., 2000, 2002; Trumbull et al., 2006), we suggest that large-scale lateral flow of ductile lower crust from tectonically overthickened regions (Husson and Sempere, 2003) may have contributed to crustal thickening in this part of the Andes at this particular time.

ACKNOWLEDGMENTS

This work is based on >20 a of sampling by Wörner and his students during many expeditions to the Central Andes of northern Chile, Bolivia, and southern Peru. Jean-Claude Thouret introduced Wörner to the volcanoes of southern Peru and helped sample the Holocene centers and ignimbrites there. His insight into the stratigraphy and volcanology of southern Peru was a great help. We also thank Michel Fornari for fruitful discussions and insights. We are grateful to G. Hartmann and K. Simons for providing analytical support, and to B. Hansen for access to the isotope laboratory. Mamani and Wörner were supported by the German Science Foundation grant Wo362/31-1 and a grant by the Deutscher Akademischer Austausch Dienst. Sempere was supported by the Institut de Recherche pour le Développement (IRD).

REFERENCES CITED

Aitcheson, S.J., Harmon, R.S., Moorbath, S., Schneider, A., Soler, P., Soria, E.E., Steele, G., Swainbank, I., and Wörner, G., 1995, Pb isotopes define basement domains of the Altiplano, Central Andes: Geology, v. 23, p. 555–558, doi: 10.1130/0091-7613(1995)023<0555:PIDBDO>2.3.CO;2.

Albarède, F., 1996, Introduction to Geochemical Modelling: New York, Cambridge University Press, 563 p.

Allmendinger, R.W., Jordan, T.E., Kay, S.M., and Isacks, B.L., 1997, The evolution of the Altiplano-Puna Plateau of the Central Andes: Annual Review of Earth and Planetary Sciences, v. 25, p. 139–174, doi: 10.1146/annurev.earth.25.1.139.

Allmendinger, R.W., González, G., Yu, J., Hoke, G., and Isacks, B., 2005, Trench-parallel shortening in the northern Chilean forearc: Tectonic and climatic implications: Geological Society of America Bulletin, v. 117, p. 89–104, doi: 10.1130/B25505.1.

Angermann, D., Klotz, J., and Reigber, C., 1999, Space-geodetic estimation of the Nazca–South America Euler vector: Earth and Planetary Science Letters, v. 171, p. 329–334, doi: 10.1016/S0012-821X(99)00173-9.

Babeyko, A.Yu., Sobolev, S.V., Trumbull, R.B., Oncken, O., and Lavier, L.L., 2002, Numerical models of crustal scale convection and partial melting beneath the Altiplano-Puna Plateau: Earth and Planetary Science Letters, v. 199, p. 373–388, doi: 10.1016/S0012-821X(02)00597-6.

Beck, S.L., Zandt, G., Myers, S.C., Stephen, C., Wallace, T.C., and Silver, P.G., 1996, Crustal thickness variations in the central Andes: Geology, v. 24, p. 407–411, doi: 10.1130/0091-7613(1996)024<0407:CTVITC>2.3.CO;2.

Boily, M., Brooks, C., and Ludden, J.N., 1989, Chemical and isotopic evolution of the Coastal Batholith of southern Peru: Journal of Geophysical Research, v. 94, p. 483–498, doi: 10.1029/JB094iB09p12483.

Callot, P., Sempere, T., Odonne, F., and Robert, E., 2008, Giant submarine collapse of a carbonate platform at the Turonian-Coniacian transition: The Ayabacas Formation, southern Peru: Basin Research, v. 20, p. 333–357, doi: 10.1111/j.1365-2117.2008.00358.x.

Carlier, G., Lorand, J.-P., Liégeois, J.-P., Fornari, M., Soler, P., Carlotto, V., and Cardenas, J., 2005, Potassic-ultrapotassic mafic rocks delineate two lithospheric mantle blocks beneath the southern Peruvian Altiplano: Geology, v. 33, p. 601–604, doi: 10.1130/G21643.1.

Chang, Y.-H., 2008, O-Isotopes as Tracer for Assimilation Processes in Different Magmatic Regimes (El Misti, Southern Peru and Taapaca, Northern Chile) [Master's thesis]: Göttingen, Germany, Universität Göttingen, 51 p.

Charrier, R., Pinto, L., and Rodríguez, M.P., 2007, Tectonostratigraphic evolution of the Andean orogen in Chile, *in* Moreno, T., and Gibbons, W., eds., The Geology of Chile: London, The Geological Society of London, p. 21–114.

Clark, A.H., Farrar, E., Kontak, D.J., Langridge, R.J., Arenas, M.J., France, L.J., McBride, S.L., Woodman, P.L., Wasteneys, H.A., Sandeman, H.A., and Archibald, D.A., 1990, Geologic and geochronologic constraints on the metallogenic evolution of the Andes of southeastern Peru: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 85, p. 1520–1583.

Clift, D., and Hartley, A.J., 2007, Slow rates of subduction erosion and coastal underplating along the Andean margin of Chile and Peru: Geology, v. 35, p. 503–506, doi: 10.1130/G23584A.1.

Cornejo, P., 2005, Geológico de los sistemas de pórfidos cuprífero en Chile: Santiago, Chile: Revista del Servicio Nacional de Geología y Minería, v. 2, p. 10–18.

David, C., 2007, Comportement Actuel de l'Avant-Arc et de l'Arc du Coude d'Arica dans l'Orogenèse des Andes Centrales [Ph.D. thesis]: Toulouse, France, Université Paul Sabatier, 290 p.

Davidson, J.P., 1991, "Role of subduction erosion in the generation of Andean magmas" by C.R. Stern: Comment: Geology, v. 19, p. 1054–1055, doi: 10.1130/0091-7613(1991)019<1054:CARORO>2.3.CO;2.

Davidson, J.P., McMillan, N.J., Moorbath, S., Wörner, G., Harmon, R.S., and López-Escóbar, L., 1990, The Nevados de Payachata volcanic region (18°S/69°W, northern Chile). II: Evidence for widespread crustal involvement in Andean magmatism: Contributions to Mineralogy and Petrology, v. 105, p. 412–432, doi: 10.1007/BF00286829.

Delacour, A., Gerbe, M.-C., Thouret, J.-C., Wörner, G., and Paquereau-Lebti, P., 2007, Magma evolution of Quaternary minor volcanic centres in southern Peru: Bulletin of Volcanology, v. 69, p. 581–608, doi: 10.1007/s00445-006-0096-z.

Deruelle, B., 1982, Petrology of Plio-Quaternary volcanism of the south-central and meridional Andes: Journal of Volcanology and Geothermal Research, v. 14, p. 77–124, doi: 10.1016/0377-0273(82)90044-0.

de Silva, S.L., and Francis, P.E., 1991, Volcanoes of the Andes: Berlin, Springer-Verlag, 216 p.

de Silva, S.L., Zandt, G., Trumbull, R., Viramonte, J., Salas, G., and Jiménez, N., 2006, Large ignimbrite eruptions and volcanotectonic depressions in the Central Andes: A thermomechanical perspective, *in* de Natale, G., Troise, C., and Kilburn, C., eds., Mechanisms of Activity and Unrests at Large Calderas: Geological Society of London Special Publication 269, p. 47–63.

Dunai, T.J., González, G., and Juez-Larré, J., 2005, Oligocene-Miocene age of aridity in the Atacama Desert revealed by exposure dating of erosion-sensitive landforms: Geology, v. 33, p. 321–324, doi: 10.1130/G21184.1.

Echavarria, L., Nelson, E., Humphrey, J., Chávez, J., Escobedo, L., and Iriondo, A., 2006, Geologic evolution of the Caylloma epithermal vein district, southern Peru: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 101, p. 843–864.

England, P., Engdahl, R., and Thatcher, W., 2004, Systematic variations in the depths of slabs beneath arc volcanoes: Geophysical Journal International, v. 156, p. 377–408, doi: 10.1111/j.1365-246X.2003.02132.x.

Entenmann, J., 1994, Magmatic Evolution of the Nevados de Payachata Complex and the Petrogenesis of Basaltic Andesites in the Central Volcanic Zone of Northern Chile [Ph.D. thesis]: Mainz, Germany, Universität Mainz, 184 p.

Farías, M., Charrier, R., Comte, D., Martinod, J., and Hérail, G., 2005, Late Cenozoic deformation and uplift of the western flank of the Altiplano: Evidence from the depositional, tectonic, and geomorphic evolution and shallow seismic activity (northern Chile at 19°30′S): Tectonics, v. 24, p. TC4001, doi: 10.1029/2004TC001667.

Fornari, M., Baldellón, E., Espinoza, F., Ibarra, I., Jiménez, N., and Mamani, M., 2002, Ar-Ar dating of late Oligocene–early Miocene volcanism in the Altiplano, *in* 5th International Symposium on Andean Geodynamics: Paris, Institut de recherché pour le développement and Université Paul Sabatier, Extended abstract, p. 223–226.

Garzione, C.N., Molnar, P., Libarkin, J.C., and MacFadden, B.J., 2006, Rapid late Miocene rise of the Bolivian Altiplano: Evidence for removal of mantle lithosphere: Earth and Planetary Science Letters, v. 241, p. 543–556, doi: 10.1016/j.epsl.2005.11.026.

Garzione, C.N., Molnar, P., Libarkin, J.C., and MacFadden, B.J., 2007, Reply to Comment on "Rapid late Miocene rise of the Andean plateau: Evidence for removal of mantle lithosphere" by Garzione et al. (2006): Earth and Planetary Science Letters, v. 259, p. 630–633, doi: 10.1016/j.epsl.2007.04.054.

Garzione, C.N., Hoke, G.D., Libarkin, J.C., Withers, S., MacFadden, B.J., Eiler, J., Ghosh, P., and Mulch, A., 2008, Rise of the Andes: Science, v. 320, p. 1304–1307, doi: 10.1126/science.1148615.

Gregory-Wodzicki, K.M., 2000, Uplift history of the Central and Northern Andes: A review: Geological Society of America Bulletin, v. 112, p. 1091–1105, doi: 10.1130/0016-7606(2000)112<1091:UHOTCA>2.3.CO;2.

Haederle, M., and Atherton, M.P., 2002, Shape and intrusion style of the Coastal Batholith, Peru: Tectonophysics, v. 345, p. 17–28, doi: 10.1016/S0040-1951(01)00204-9.

Hampel, A., 2002, The migration history of the Nazca Ridge along the Peruvian active margin: A re-evaluation: Earth and Planetary Science Letters, v. 203, p. 665–679, doi: 10.1016/S0012-821X(02)00859-2.

Hartley, A.J., Sempere, T., and Wörner, G., 2007, A comment on "Rapid late Miocene rise of the Bolivian Altiplano: Evidence for removal of mantle lithosphere" by C.N. Garzione et al.: Earth and Planetary Science Letters, v. 259, p. 625–629, doi: 10.1016/j.epsl.2007.04.012.

Haschke, M., and Günther, A., 2003, Balancing crustal thickening in arcs by tectonic vs. magmatic means: Geology, v. 31, p. 933–936, doi: 10.1130/G19945.1.

Haschke, M.R., Siebel, W., Günther, A., and Scheuber, E., 2002, Repeated crustal thickening and recycling during the Andean orogeny in north Chile (21°–26°S): Journal of Geophysical Research, v. 107, p. 1–18, doi: 10.1029/2001JB000328.

Heuret, A., Funiciello, F., Faccenna, C., and Lallemand, S., 2007, Plate kinematics, slab shape and back-arc stress: A comparison between laboratory models and current subduction zones: Earth and Planetary Science Letters, v. 256, p. 473–483, doi: 10.1016/j.epsl.2007.02.004.

Hildreth, W., and Moorbath, S., 1988, Crustal contributions to arc magmatism in the Andes of central Chile: Contributions to Mineralogy and Petrology, v. 98, p. 455–489, doi: 10.1007/BF00372365.

Hoke, G., Isacks, B.L., Jordan, T.E., Blanco, N., Tomlinson, A.J., and Ramezani, J., 2007, Geomorphic evidence for post–10 Ma uplift of the western flank of the central Andes, 18°30′–22°S: Tectonics, v. 26, p. TC5021, doi: 10.1029/2006TC002082.

Husson, L., and Sempere, T., 2003, Thickening the Altiplano crust by gravity-driven crustal channel flow: Geophysical Research Letters, v. 30, p. 1243, doi: 10.1029/2002GL016877.

Isacks, B.L., 1988, Uplift of the Central Andean plateau and bending of the Bolivian orocline: Journal of Geophysical Research, v. 93, p. 3211–3231, doi: 10.1029/JB093iB04p03211.

James, D.E., 1971a, Andean crustal and upper mantle structure: Journal of Geophysical Research, v. 76, p. 3246–3271, doi: 10.1029/JB076i014p03246.

James, D.E., 1971b, Plate tectonic model for the evolution of the Central Andes: Geological Society of America Bulletin, v. 82, p. 3325–3346, doi: 10.1130/0016-7606(1971)82[3325:PTMFTE]2.0.CO;2.

James, D.E., 1982, A combined O, Sr, Nd and Pb isotopic and trace element study of crustal contamination in central Andean lavas. I: Local geochemical variations: Earth and Planetary Science Letters, v. 57, p. 47–62, doi: 10.1016/0012-821X(82)90172-8.

James, D.E., and Sacks, I.S., 1999, Cenozoic formation of the Central Andes: A geophysical perspective, *in* Skinner, B.J., ed., Geology and Ore Deposits of the Central Andes: Society of Economic Geologists Special Publication 7, p. 1–25.

Jiménez, N., Barrera, L., Flores, O., Lizeca, J.L., Murillo, S., Sanjinés, O., Hardyman, R.F., Tosdal, R.M., and Wallace, A.R., 1993, Magmatic evolution of the Berengüela-Charaña region, northwestern Altiplano, Bolivia, *in* 2nd International Symposium on Andean Geodynamics: Paris, Extended abstract, p. 377–380.

Kaneoka, I., and Guevara, C., 1984, K-Ar age determinations of late Tertiary and Quaternary Andean volcanic rocks, southern Peru: Geochemical Journal, v. 18, p. 233–239.

Kay, S.M., and Mpodozis, C., 2001, Central Andean ore deposits linked to evolving shallow subduction systems and thickening crust: GSA Today, v. 11, p. 4–9, doi: 10.1130/1052-5173(2001)011<0004:CAODLT>2.0.CO;2.

Kay, S.M., Coira, B., and Viramonte, J., 1994, Young mafic back arc volcanic rocks as indicators of continental lithospheric delamination beneath the Argentine Puna Plateau, Central Andes: Journal of Geophysical Research, v. 99, p. 24,323–24,339, doi: 10.1029/94JB00896.

Kay, S.M., Mpodozis, C., and Coira, B., 1999, Neogene magmatism, tectonism and mineral deposits of the Central Andes (22° to 33°S latitude), *in* Skinner, B.J., ed., Geology and Ore Deposits of the Central Andes: Society of Economic Geologists Special Publication 7, p. 27–59.

Kay, S.M., Godoy, E., and Kurtz, A., 2005, Episodic arc migration, crustal thickening, subduction erosion, and magmatism in the south-central Andes: Geological Society of America Bulletin, v. 117, p. 67–88, doi: 10.1130/B25431.1.

Kennan, L., 2000, Large-scale geomorphology of the Andes: Interrelationships of tectonics, magmatism and climate, *in* Summerfield, M.A., ed., Geomorphology and Global Tectonics: New York, Wiley, p. 165–199.

Kiebala, A., 2008, Magmatic Processes by U-Th Disequilibria: Comparison of Two Andean Magmatic Systems: El Misti (Southern Peru) and Taapaca (Northern Chile) [Ph.D. thesis]: Göttingen, Germany, Universität Göttingen, 82 p.

Klinck, B.A., Ellison, R.A., and Hawkins, M.P., 1986, The Geology of the Cordillera Occidental and Altiplano West of Lake Titicaca, Southern Peru: Lima, Peru, Instituto de Geología Minería y Metalurgía, preliminary report, 353 p.

Kohn, M.J., ed., 2007, Paleoaltimetry: Geochemical and Thermodynamic Approaches: Reviews in Mineralogy and Geochemistry, v. 66, 278 p.

Kono, M., Fukao, Y., and Yamamoto, A., 1989, Mountain building in the Central Andes: Journal of Geophysical Research, v. 94, p. 3891–3905, doi: 10.1029/JB094iB04p03891.

Kontak, D.J., Farrar, E., Clark, A.H., and Archibald, D.A., 1990, Eocene tectono-thermal rejuvenation of an Upper Paleozoic–Lower Mesozoic terrane in the Cordillera de Carabaya, Puno, southeastern Peru, revealed by K-Ar and $^{40}Ar/^{39}Ar$ dating: Journal of South American Earth Sciences, v. 3, p. 231–246, doi: 10.1016/0895-9811(90)90005-L.

Kowalski, E.A., 2002, Mean annual temperature estimation based on leaf morphology: A test from tropical South America: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 188, p. 141–165, doi: 10.1016/S0031-0182(02)00550-3.

Lamb, S., and Davis, P., 2003, Cenozoic climate change as a possible cause for the rise of the Andes: Nature, v. 425, p. 792–797, doi: 10.1038/nature02049.

Lee, C.T.A., Cheng, X., and Horodyskyj, U., 2006, The development and refinement of continental arcs by primary basaltic magmatism, garnet pyroxenite accumulation, basaltic recharge and delamination: Insights from the Sierra Nevada, California: Contributions to Mineralogy and Petrology, v. 151, p. 222–242, doi: 10.1007/s00410-005-0056-1.

Le Maitre, R.W., ed., 1989, A Classification of Igneous Rocks and Glossary of Terms: Recommendations of the International Union of Geological Sciences Subcommission on the Systematics of Igneous Rocks: Oxford, UK, Blackwell, 193 p.

Loewy, S.L., Connelly, J.N., and Dalziel, I.W.D., 2004, An orphaned basement block: The Arequipa-Antofalla basement of the Central Andean margin of South America: Geological Society of America Bulletin, v. 116, p. 171–187, doi: 10.1130/B25226.1.

Lucassen, F., Becchio, R., Harmon, R., Kasemann, S., Franz, G., Trumbull, R., Wilke, H., Romer, R., and Dulski, P., 2001, Composition and density model of the continental crust at an active continental margin—The Central Andes between 21°S and 27°S: Tectonophysics, v. 341, p. 195–223, doi: 10.1016/S0040-1951(01)00188-3.

Macfarlane, A.W., Marcet, P., LeHuray, A.P., and Petersen, U., 1990, Lead isotope provinces of the Central Andes inferred from ores and crustal rocks: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 85, p. 1857–1880.

Macpherson, C.G., Dreher, S.T., and Thirlwall, M.F., 2006, Adakites without slab melting: High pressure differentiation of island arc magma, Mindanao, the Philippines: Earth and Planetary Science Letters, v. 243, p. 581–593, doi: 10.1016/j.epsl.2005.12.034.

Maksaev, V., and Zentilli, M., 1999, Fission-track thermochronology of the Domeyko Cordillera, Chile: Metallogenetic implications for Andean porphyry copper metallogenesis: Exploration and Mining Geology, v. 8, p. 65–89.

Mamani, M., Ibarra, I., Carlier, G., and Fornari, M., 2004, Petrología y geoquímica del magmatismo alcalino de la zona noroeste del Altiplano peruano (departamento de Puno), *in* Jacay, J., and Sempere, T., eds., Nuevas Contribuciones del IRD y sus Contrapartes al Conocimiento Geológico del Sur del Perú: Sociedad Geológica del Perú Publicación Especial 5, p. 157–174.

Mamani, M., Tassara, A., and Wörner, G., 2008a, Composition and structural control of crustal domains in the Central Andes: Geochemistry, Geophysics, Geosystems, v. 9, doi: 10.1029/2007GC001925.

Mamani, M., Wörner, G., and Thouret, J.-C., 2008b, Tracing a major crustal domain boundary based on the geochemistry of minor volcanic centres in southern Peru, *in* 7th International Symposium on Andean Geodynamics: Paris, Institut de recherché pour le développement and Université Nice Sophia Antipolis, Extended abstract, p. 298–301.

Martínez, W., and Cervantes, J., 2003, Rocas ígneas en el sur del Perú: Nuevos datos geocronométricos, geoquímicos y estructurales entre los paralelos 16° y 18°30′S: Lima, Peru, INGEMMET, Boletínes, serie D, v. 26, 140 p.

Martínez, W., Cervantes, J., Romero, D., and Sempere, T., 2005, The late Paleozoic–early Mesozoic Chocolate Formation of southern Peru: New data and interpretations, *in* 6th International Symposium on Andean Geodynamics: Paris, Institut de recherché pour le développement and Universidad de Barcelona, Extended abstracts, p. 490–492.

McMillan, N., Davidson, J., Wörner, G., Harmon, R.S., López-Escober, L., and Moorbath, S., 1993, Mechanism of trace element enrichment related to crustal thickening: The Nevados de Payachata region, Northern Chile: Geology, v. 21, p. 467–470, doi: 10.1130/0091-7613(1993)021<0467:IOCTOA>2.3.CO;2.

Molnar, P., and Garzione, C.N., 2007, Bounds on the viscosity coefficient of continental lithosphere from removal of mantle lithosphere beneath the Altiplano and Eastern Cordillera: Tectonics, v. 26, p. TC2013, doi: 10.1029/2006TC001964.

Mpodozis, C., Arriagada, C., Basso, M., Roperch, P., Cobbold, P., and Reich, M., 2005, Late Mesozoic to Paleogene stratigraphy of the Salar de Atacama Basin, Antofagasta, northern Chile: Implications for the tectonic evolution of the Central Andes: Tectonophysics, v. 399, p. 125–154, doi: 10.1016/j.tecto.2004.12.019.

Mukasa, S.B., 1986a, Zircon U-Pb ages of super-units in the Coastal Batholith, Peru: Implications for magmatic and tectonic processes: Geological Society of America Bulletin, v. 97, p. 241–254, doi: 10.1130/0016-7606(1986)97<241:ZUAOSI>2.0.CO;2.

Mukasa, S.B., 1986b, Common Pb isotopic compositions of the Lima, Arequipa and Toquepala segments in the Coastal Batholith, Peru: Implication for magma genesis: Geochimica et Cosmochimica Acta, v. 50, p. 771–782, doi: 10.1016/0016-7037(86)90353-4.

Muntener, O., Kelemen, P.B., and Grove, T.L., 2001, The role of H_2O during crystallization of primitive arc magmas under uppermost mantle conditions and genesis of igneous pyroxenites: An experimental study: Contributions to Mineralogy and Petrology, v. 141, p. 643–658.

Oliveros, V., Féraud, G., Aguirre, L., Fornari, M., and Morata, D., 2006, The Early Andean magmatic province (EAMP): $^{40}Ar/^{39}Ar$ dating on Mesozoic volcanic and plutonic rocks from the Coastal Cordillera, northern Chile: Journal of Volcanology and Geothermal Research, v. 157, p. 311–330, doi: 10.1016/j.jvolgeores.2006.04.007.

Oncken, O., Hindle, D., Kley, J., Elger, K., Victor, P., and Schemmann, K., 2006, Deformation of the Central Andean upper plate system: Facts, fiction, and constraints for plateau models, *in* Oncken, O., et al., eds., The Andes: Active Subduction Orogeny: Berlin, Springer-Verlag, Frontiers in Earth Sciences, v. 1, p. 3–28.

Paquereau-Lebti, P., Thouret, J.-C., Wörner, G., and Fornari, M., 2006, Neogene and Quaternary ignimbrites in the area of Arequipa, southern Peru: Stratigraphical and petrological correlations: Journal of Volcanology and Geothermal Research, v. 154, p. 251–275, doi: 10.1016/j.jvolgeores.2006.02.014.

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, doi: 10.1029/TC006i003p00233.

Perelló, J., Carlotto, V., Zarate, A., Ramos, P., Posso, H., and Caballero, A., 2003, Porphyry-style alteration and mineralization of the middle Eocene to early Oligocene Andahuaylas-Yauri Belt, Cusco region, Peru: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 98, p. 1575–1605.

Petford, N., and Atherton, M., 1996, Na-rich partial melts from newly underplated basaltic crust: The Cordillera Blanca Batholith, Peru: Journal of Petrology, v. 37, p. 1491–1521, doi: 10.1093/petrology/37.6.1491.

Picard, D., Sempere, T., and Plantard, O., 2008, Direction and timing of uplift propagation in the Peruvian Andes deduced from the molecular phylogeny of highland biotaxa: Earth and Planetary Science Letters, v. 271, p. 326–336, doi: 10.1016/j.epsl.2008.04.024.

Pino, A., Sempere, T., Jacay, J., and Fornari, M., 2004, Estratigrafía, paleogeografía y paleotectónica del intervalo Paleozoico superior-Cretáceo inferior en el área de Mal Paso-Palca (Tacna), *in* Jacay, J., and Sempere, T., eds., Nuevas Contribuciones del IRD y sus Contrapartes al Conocimiento Geológico del Sur del Perú: Sociedad Geológica del Perú, Publicación Especial 5, p. 15–44.

Pitcher, W.S., Atherton, M.P., Cobbing, E.J., and Beckinsale, R.D., 1985, Magmatism at a Plate Edge: The Peruvian Andes: Glasgow, Blackie & Son, and New York, Halsted Press, 328 p.

Plank, T., and Langmuir, C.H., 1998, The chemical composition of subducting sediment and its consequences for the crust and mantle: Chemical Geology, v. 145, p. 325–394, doi: 10.1016/S0009-2541(97)00150-2.

Quang, C.X., Clark, A.H., Lee, J.K.W., and Hawkes, N., 2005, Response of supergene processes to episodic Cenozoic uplift, pediment erosion, and ignimbrite eruption in the Porphyry Copper Province of southern Peru: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 100, p. 87–114.

Ramos, V.A., 2008, The basement of the Central Andes: The Arequipa and related terranes: Annual Review of Earth and Planetary Sciences, v. 36, p. 289–324, doi: 10.1146/annurev.earth.36.031207.124304.

Rech, J.A., Currie, B.S., Michalski, G., and Cowan, A.M., 2006, Neogene climate change and uplift in the Atacama Desert, Chile: Geology, v. 34, p. 761–764, doi: 10.1130/G22444.1.

Rogers, G., and Hawkesworth, C.J., 1989, A geochemical traverse across the north Chilean Andes: Evidence for crustal generation from the mantle wedge: Earth and Planetary Science Letters, v. 91, p. 271–285, doi: 10.1016/0012-821X(89)90003-4.

Rollinson, H., 1993, Using Geochemical Data: Evaluation, Presentation, Interpretation: Harlow, UK, Longman Scientific & Technical, 352 p.

Roperch, P., Sempere, T., Macedo, O., Arriagada, C., Fornari, F., Tapia, C., García, M., and Laj, C., 2006, Counterclockwise rotation of late Eocene–Oligocene fore-arc deposits in southern Peru and its significance for oroclinal bending in the Central Andes: Tectonics, v. 25, p. TC3010, doi: 10.1029/2005TC001882.

Russo, R.M., and Silver, P.G., 1996, Cordillera formation, mantle dynamics, and the Wilson cycle: Geology, v. 24, p. 511–514, doi: 10.1130/0091-7613(1996)024<0511:CFMDAT>2.3.CO;2.

Sandeman, H.A., Clark, A.H., and Farrar, E., 1995, An integrated tectono-magmatic model for the evolution of the southern Peruvian Andes (13°–20° S) since 55 Ma: International Geology Review, v. 37, p. 1039–1073.

Schellart, W.P., Freeman, J., Stegman, D.R., Moresi, L., and May, D., 2007, Evolution and diversity of subduction zones controlled by slab width: Nature, v. 446, p. 308–311, doi: 10.1038/nature05615.

Schildgen, T.F., Hodges, K.V., Whipple, K.X, Reiners, P.W., and Pringle, M.S., 2007, Uplift of the western margin of the Andean plateau revealed from canyon incision history, southern Peru: Geology, v. 35, p. 523–526, doi: 10.1130/G23532A.1.

Sempere, T., and Jacay, J., 2007, Synorogenic extensional tectonics in the forearc, arc and southwest Altiplano of southern Peru: Eos (Transactions, American Geophysical Union), v. 88, no. 23, Joint Assembly supplement, abstract U51B–04.

Sempere, T., and Jacay, J., 2008, Anatomy of the Central Andes: Distinguishing between western, magmatic Andes and eastern, tectonic Andes, *in* 7th International Symposium on Andean Geodynamics: Paris, Institut de recherché pour le développement and Université Nice Sophia Antipolis, Extended abstracts, p. 504–507.

Sempere, T., Hérail, G., Oller, J., and Bonhomme, M.G., 1990, Late Oligocene–early Miocene major tectonic crisis and related basins in Bolivia: Geology, v. 18, p. 946–949, doi: 10.1130/0091-7613(1990)018<0946:LOEMMT>2.3.CO;2.

Sempere, T., Carlier, G., Soler, P., Fornari, M., Carlotto, V., Jacay, J., Arispe, O., Néraudeau, D., Cárdenas, J., Rosas, S., and Jiménez, N., 2002, Late Permian–Middle Jurassic lithospheric thinning in Peru and Bolivia, and its bearing on Andean-age tectonics: Tectonophysics, v. 345, p. 153–181, doi: 10.1016/S0040-1951(01)00211-6.

Sempere, T., Folguera, A., and Gerbault, M., 2008, New insights into the Andean evolution: An introduction to contributions from the 6th ISAG symposium (Barcelona, 2005): Tectonophysics, v. 459, p. 1–13, doi: 10.1016/j.tecto.2008.03.011.

Somoza, R., 1998, Updated Nazca (Farallon)–South America relative motions during the last 40 My: Implications for the mountain building in the Central Andean region: Journal of South American Earth Sciences, v. 11, p. 211–215, doi: 10.1016/S0895-9811(98)00012-1.

Stern, C.R., 1990, Comment on "A geochemical traverse across the North Chilean Andes: Evidence for crust generation from the mantle wedge" by G. Rogers and C.J. Hawkesworth: Earth and Planetary Science Letters, v. 101, p. 129–133, doi: 10.1016/0012-821X(90)90134-J.

Stern, C.R., 1991, Role of subduction erosion in the generation of Andean magmas: Geology, v. 19, p. 78–81, doi: 10.1130/0091-7613(1991)019<0078:ROSEIT>2.3.CO;2.

Thouret, J.-C., Rivera, M., Wörner, G., Gerbe, M.-C., Finizola, A., Fornari, M., and Gonzales, K., 2005, Ubinas: Evolution of the historically most active volcano in southern Peru: Bulletin of Volcanology, v. 67, p. 557–589, doi: 10.1007/s00445-004-0396-0.

Thouret, J.-C., Wörner, G., Gunnell, Y., Singer, B., Zhang, X., and Souriot, T., 2007, Geochronologic and stratigraphic constraints on canyon incision and Miocene uplift of the Central Andes in Peru: Earth and Planetary Science Letters, v. 263, p. 151–166, doi: 10.1016/j.epsl.2007.07.023.

Tilton, G.R., and Barreiro, B.A., 1980, Origin of lead in Andean calc-alkaline lavas, southern Peru: Science, v. 210, p. 1245–1247, doi: 10.1126/science.210.4475.1245.

Tosdal, R.M., 1996, The Amazon-Laurentian connection as viewed from the Middle Proterozoic rocks in the central Andes, western Bolivia and northern Chile: Tectonics, v. 15, p. 827–842.

Tosdal, R.M., Farrar, E., and Clark, A.H., 1981, K-Ar geochronology of the late Cenozoic volcanic rocks of the Cordillera Occidental, southernmost Peru: Journal of Volcanology and Geothermal Research, v. 10, p. 157–173, doi: 10.1016/0377-0273(81)90060-3.

Trumbull, R.B., Wittenbrink, R., Hahne, K., Emmermann, R., Büsch, W., Gerstenberger, H., and Siebel, W., 1999, Evidence for late Miocene to Recent contamination of arc andesites by crustal melts in the Chilean Andes (25°–26°S) and its geodynamic implications: Journal of South American Earth Sciences, v. 12, p. 135–155, doi: 10.1016/S0895-9811(99)00011-5.

Trumbull, R., Riller, U., Oncken, O., Scheuber, E., Munier, K., and Hongn, F., 2006, The time-space distribution of Cenozoic volcanism in the south-central Andes: A new data compilation and some tectonic implications, *in* Oncken, O., et al., eds., The Andes: Active Subduction Orogeny: Berlin, Springer-Verlag, Frontiers in Earth Sciences, p. 29–43.

von Huene, R., and Scholl, D.W., 1991, Observations at convergent margins concerning sediment subduction, subduction erosion, and the growth of continental crust: Reviews of Geophysics, v. 29, p. 279–316, doi: 10.1029/91RG00969.

von Huene, R., Weinrebe, W., and Heeren, F., 1999, Subduction erosion along the North Chile margin: Geodynamics, v. 27, p. 345–358, doi: 10.1016/S0264-3707(98)00002-7.

Wasteneys, H.A.H.P., 1990, Epithermal Silver Mineralization with a Mid-Tertiary Diatreme: Santa Bárbara, Santa Lucía District, Puno, Peru [Ph.D. thesis]: Kingston, Ontario, Canada, Queen's University, 367 p.

Wörner, G., Harmon, R.S., Davison, J.D., Moorbath, S., Turner, T.L., McMillan, N., Nye, C., López-Escobar, L., and Moreno, H., 1988, The Nevados de Payachata volcanic region 18°S/69°W, N. Chile. I. Geological, geochemical and isotopic observations: Bulletin of Volcanology, v. 50, p. 287–303, doi: 10.1007/BF01073587.

Wörner, G., Moorbath, S., and Harmon, R.S., 1992, Andean Cenozoic volcanics reflect basement isotopic domains: Geology, v. 20, p. 1103–1106, doi: 10.1130/0091-7613(1992)020<1103:ACVCRB>2.3.CO;2.

Wörner, G., Moorbath, S., Horn, S., Entenmann, J., Harmon, R.S., Davidson, J.P., and López-Escóbar, L., 1994, Large geochemical variations along the Andean arc of northern Chile (17.5–22°S), *in* Reutter, K.J., Scheuber, E., and Wigger, P.J., eds., Tectonics of the Southern Central Andes: Structure and Evolution of an Active Continental Margin: Berlin, Springer-Verlag, p. 77–91.

Wörner, G., Hammerschmidt, K., Henjes-Kunst, F., Lezaun, J., and Wilke, H., 2000, Geochronology (^{40}Ar-^{39}Ar-, K-Ar-, and He-exposure-) ages of Cenozoic magmatic rocks from northern Chile (18°–22°S). Implications for magmatism and tectonic evolution of the Central Andes: Revista Geologica Chile, v. 27, p. 205–240.

Wörner, G., Uhlig, D., Kohler, I., and Seyfried, H., 2002, Evolution of the West Andean Escarpment at 18°S (N. Chile) during the last 25 Ma: Uplift, erosion and collapse through time: Tectonophysics, v. 345, p. 183–198, doi: 10.1016/S0040-1951(01)00212-8.

Yang, Y., Liu, M., and Stein, S., 2003, A 3-D geodynamic model of lateral crustal flow during Andean mountain building: Geophysical Research Letters, v. 30, p. 2093, doi: 10.1029/2003GL018308.

Yuan, X., Sobolev, S.V., and Kind, R., 2002, Moho topography in the Central Andes and its geodynamic implication: Earth and Planetary Science Letters, v. 199, p. 389–402, doi: 10.1016/S0012-821X(02)00589-7.

Manuscript Received 10 September 2008
Revised Manuscript Received 31 March 2009
Manuscript Accepted 3 April 2009

Printed in the USA