REVIEW PAPER

Active Andean volcanism: its geologic and tectonic setting

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ABSTRACT

The Andean volcanic arc includes over 200 potentially active Quaternary volcanoes, and at least 12 giant caldera/ ignimbrite systems, occurring in four separate segments referred to as the Northern, Central, Southern and Austral Volcanic Zones. Volcanism results from subduction of the Nazca and Antarctic oceanic plates below South America. Active volcanoes occur where the angle of subduction is relatively steep (>25°), and active arc segments are separated by regions below which subduction angle decreases and becomes relatively flat (<10°) at depths >100 km. Segments of low angle subduction formed beginning in the Miocene in association with subduction of buoyant oceanic plateaus and ridges, and current segmentation of subduction geometry and active Andean volcanic zones is clearly a transient feature related to Neogene tectonics. A genetic relation between subduction and volcanism is confirmed by geochemical studies indicating that generation of Andean magmas is initiated by dehydration and/or melting of subducting oceanic lithosphere and interaction of these slab-derived fluids/melts with the overlying mantle wedge. Continental crust is incorporated into Andean magmas by a combination of both subduction of crust into the subarc mantle and assimilation of crust into mantle-derived magmas. Variations in the rate of subduction erosion and subduction of continental crust significantly affect not only Andean magma chemistry, but also the along-strike intraplate mechanical coupling in the subduction zone and the dynamics of mountain building in the Andes. Crustal components are most significant in magmas erupted in the Central Volcanic Zone, where the crust is extremely thick (>70 km) and estimated rates of subduction erosion of the continental margin, possibly equivalent to as much as 4% of the volume of subducting oceanic crust, are also greatest due to the hyper-arid climate conditions and low sediment supply to the trench. Obvious hazards associated with Andean volcanoes include lava and pyroclastic flows, lahars, debris flows resulting from sector collapse, and tephra falls. More than 25,000 people have been killed by the >600 eruptions of these volcanoes catalogued since the year 1532, most of these by lahars generated during the eruption of Nevado del Ruiz, Colombia, in 1985. Despite the fact that >20 million people live within <100 km of an active Andean volcano, mostly in low-lying areas in the intermontane valleys of Colombia and Ecuador and the Central Valley of south-central Chile, only <25 of these volcanoes are continuously monitored for signs of activity.

Key words: South America, Andes, volcanoes, Magma genesis, Volcanic hazards.

RESUMEN

Volcanismo andino activo: marco geológico y tectónico. El arco volcánico andino incluye más de 200 estratovolcanes y, al menos, 12 sistemas de calderas gigantes potencialmente activos, dispuestos en cuatro segmentos separados de la cadena andina conocidos como Zonas Volcánicas Norte, Central, Sur y Austral, y cuya actividad es producto de la subducción de las placas oceanicas Nazca y Antártica bajo la placa sudamericana. Los cuatro segmentos con volcanismo activo ocurren en zonas donde el ángulo de subducción es relativamente inclinado (25°), y entre ellos existen regiones donde el ángulo de subducción es relativamente plano (<10°) a profundidades >100 km y el volcanismo está ausente. Las zonas de bajo ángulo de subducción habrían comenzado a formarse durante el Mioceno debido a la subducción de plateaus y dorsales oceánicas, indicando que la actual segmentación de la zona de subducción y el volcanismo andino es un rasgo transitorio relacionado a la actividad tectónica neógena. La relación genética entre subducción y volcanismo ha sido confirmada por estudios geoquímicos que indican que la actividad magmática se inicia por la deshidratación y/o fusión de la litosfera oceánica subductada y la interacción de los fluidos liberados con el manto astenosférico que la sobreyace. Componentes derivados de la corteza continental son también incorporados en los magmas andinos a través de la erosión por subducción del margen continental y/o asimilación de material cortical en los magmas derivados del manto. Las variaciones en la tasa de erosión por subducción y subducción de corteza continental afectan en forma significativa no sólo la química de los magmas andinos, sino también el acomplamiento mecánico de intraplaca en la zona de subducción y la dinámica orogénica a lo largo de los Andes. Componentes corticales son más significativos en los magmas extruidos en la Zona Volcánica Central donde la corteza es extremadamente gruesa (>70 km) y las tasas de erosión por subducción del margen continental alcanzan, posiblemente, a consumir un volumen de rocas equivalente hasta un 4% del volumen de la corteza oceánica subductada, son también muy elevadas debido a las condiciones climáticas hiperáridas y el bajo aporte de sedimentos a la fosa. Lavas, flujos piroclásticos, lahares, flujos de detritos producto de colapso sectorial de estratovolcanes, y la caída de tefra son algunos de los peligros y riesgos más importantes asociados al volcanismo andino. Desde el año 1532 más de 25.000 personas han muerto como consecuencia de >600 erupciones con registro histórico. La mayor parte de estas muertes ocurrió en 1985 durante la erupción de los Nevados del Ruiz en Colombia. A pesar de que más de 20 millones de personas viven a menos de 100 km de distancia de un volcán activo en los Andes, principalmente en los valles interandinos de Colombia y Ecuador y el Valle Central del centro-sur de Chile, en la actualidad, menos de 25 volcanes están siendo monitoreados para determinar los riesgos potenciales asociados a la actividad volcánica andina.

Palabras claves: Sud américa, Andes, Volcanes, Génesis de magma, Riesgos volcánicos

INTRODUCTION

Andean volcanoes are a natural laboratory for the study of volcanic hazard assessment and risk management, the generation of magmas in association with subduction of oceanic plates below continental lithosphere, and the geochemical evolution of continental crust. 'Volcanoes of the World' (Simkin and Siebert, 1994) lists 60 historically active and another 118 Recently (Holocene) active Andean volcanoes, for a total of 178 among 1,511 volcanoes worldwide. These occur in Colombia, Ecuador, Perú, Bolivia, Argentina and Chile, in a discontinuous segmented belt extending from Cerro Bravo volcano (5°N) in the north to Cook Island volcano (55°S) in the south (Fig. 1). However, this list does not include many 'potentially active' Quaternary volcanic centers and large silicic caldera/ ignimbrite systems, nor numerous small cones, domes or lava flows forming minor, often monogenetic centers, which occur both within and along the eastern flank of the Andes.

'Volcanoes of the World' catalogues 575 eruptions produced by the 60 historically active centers between the years 1532 (Cotopaxi, Ecuador) and 1993. The 2004 IAVCEI general assembly in Pucón, Chile, will take place at the foot of the Villarrica volcano, historically one of the most active volcanoes in the Andes, with 59 reported episodes of activity since 1558 (Moreno *et al.*, 1994a; Petit-Breuilh, 1994). Lahars produced by eruptions of Villarrica in 1948-1949, 1963-1964, and 1971-1972 together killed more than 75 people, and others are still missing. Villarrica has had 14 explosive post-glacial eruptions generating pyroclastic flows, two of which also produced calderas (Clavero and Moreno, 1994). Since the most recent eruption in 1984-85 (Fuentealba, 1984) the summit cone has contained an active lava lake (Witter *et al.*, 2004; Witter and Delmelle, 2004; Calder *et al.*, 2004), which was also there in 1976 when I climbed this volcano.

Numerous studies of individual volcanoes published in the 30 years since the 1974 IAVCEI general assembly in Santiago, Chile, have contributed a considerable amount of new information concerning Andean volcanism. Most significantly, field and remote sensing studies have identified a number of previously unrecognized, geologically young, potentially active giant ignimbrite centers and caldera systems, both in the Central Andes (de Silva, 1989a, b; de Silva and Francis, 1991), such as Cerro Panizos along the Bolivia-Argentina border (Ort, 1993; Ort *et al.*, 1996), Cerro Galán in Argentina (Francis and Baker, 1978; Sparks *et al.*, 1989; Francis *et al.*, 1989), Pastos Grandes in southwest Bolivia (Baker, 1981), and La Pacana caldera complex in northern Chile (Gardeweg and

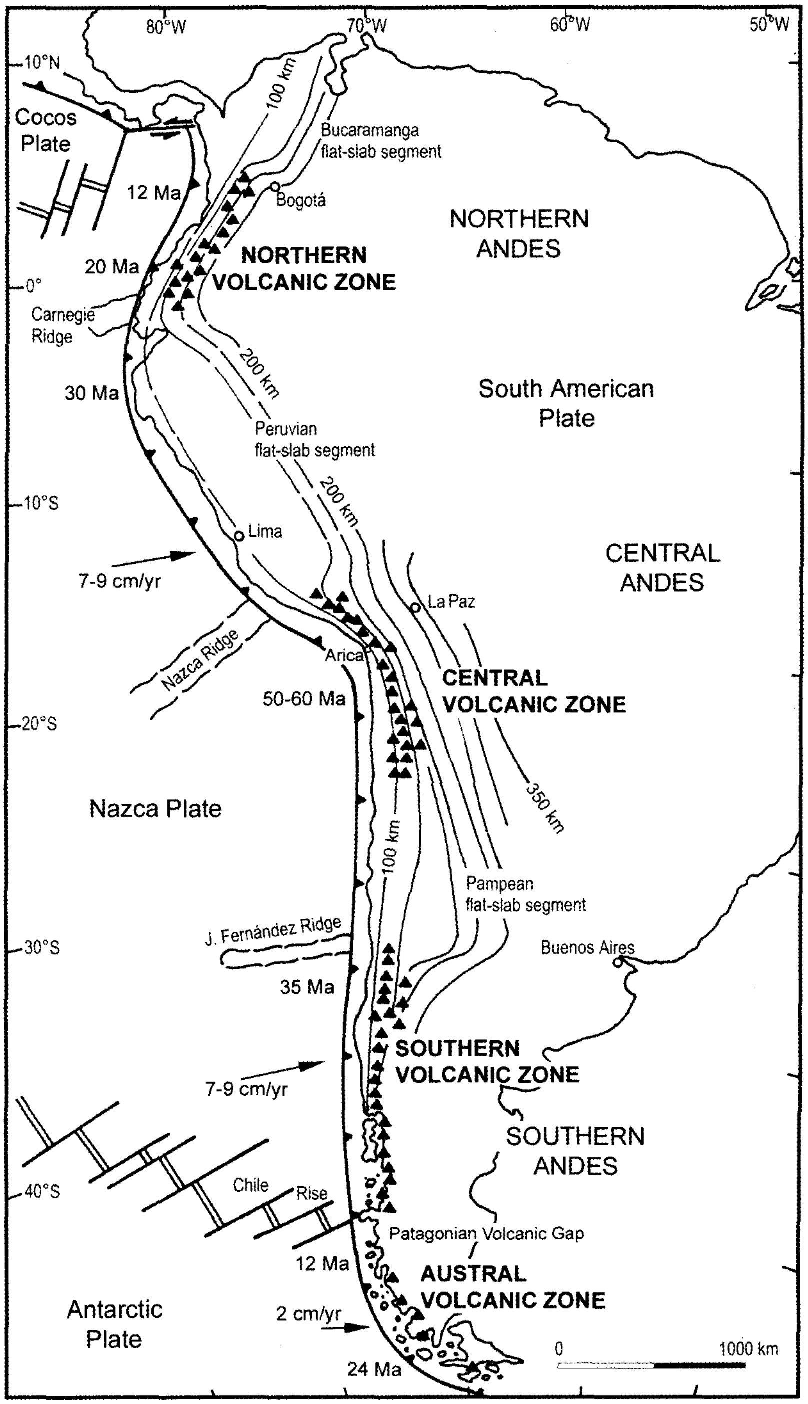

FIG. 1. Schematic map of South America and the Pacific oceanic plates, modified from Ramos and Alemán (2000), showing the four volcanically active segments in the Andes (Figs. 2-4, 7), subduction geometry as indicated by depth in kilometers to the Benioff zone (Engdahl *et al.*, 1995, 1998), oceanic ridges (Gutscher, 2002), ages of the oceanic plates close to the Perú-Chile trench, and convergence rates and directions along the length of the Andes (Norambuena *et al.*, 1998; Angermann *et al.*, 1999). Details concerning the morpho-structural segmentation of the Andes can be found in Kley *et al.* (1999), and basement ages *in* Ramos and Alemán (2000) and Loewy *et al.* (2004).

Ramírez, 1987; Lindsay *et al.*, 2001a), as well as in the southern Andes, along the border between Chile and Argentina, such as the Maipo (34°S; Stern *et al.*, 1984a), Calabozos (36°S; Hildreth *et al.*, 1984) and Caviahue (38°S; Muñoz and Stern, 1988) calderas. As noted by Hildreth *et al.* (1984), the 'long record of voluminous silicic pyroclastic activity' associated with these giant systems 'has important implications for the safety of several major cities near the mouths of Andean canyons.'

Hazards associated with Andean volcanism include pyroclastic and lava flows, lahars, debris flows generated by sector collapse, and tephra falls. Eruption of the Nevado del Ruiz volcano in November, 1985, produced lahars that resulted in the death of >23,000 people in Colombia (Williams, 1990), a clear indication that much more needs to be done concerning volcano hazard assessment and risk management in the Andes. Since this eruption, establishment of centers for hazard evaluation and risk management, field and remote sensing mapping, monitoring of individual centers, studies of distal tephra deposits, as well as more detailed historical research, have together significantly increased the base of information concerning the record of Andean volcanic activity. This special volume of the Revista Geológica de Chile, prepared for the 2004 IAVCEI general assembly, presents a number of such studies, as well as observations of eruptions in progress.

Not only has information concerning Andean volcanoes increased significantly in the last 30 years, but so also has the understanding of the geologic evolution of the Andes and the tectonic setting of Andean volcanism. This review paper presents a brief overview of the geologic and tectonic context of volcanic activity, magma genesis and volcanic hazards along the length of the Andes.

ANDEAN GEOLOGY AND TECTONICS

The Andean Cordillera forms a >7, 500 km long morphologically continuous mountain chain along the western margin of South America, from the Caribbean Coast in the north to Cape Horn in the south. Volcanism, however, occurs in four separate regions named the Northern (NVZ), Central (CVZ), Southern (SVZ) and Austral (AVZ) Volcanic Zones (Fig. 1). Each of these zones is further divided into a number of smaller volcanic arc segments. The divisions between larger zones, and between smaller segments, and differences in magmas erupted in each segment of each zone, reflect geologic and tectonic segmentation of the Andean Cordillera described briefly below.

The Andean Cordillera is segmented into regions with distinct pre-Andean basement ages, Mesozoic and Cenozoic geologic evolution, crustal thickness, structural trends, active tectonics and volcanism. A simple division into the Northern (12°N-5°S), Central (5-33°S), and Southern (33-56°S) Andes is useful for discussing the active volcanic belts (Fig. 1). The Northern Andes of Colombia and Ecuador trend northeast-southwest. The Central Andes include, to the north and south of the Arica bend, a northwest-southeast trending Peruvian segment called the Northern Central Andes, and the northern part of the north-south trending Andes in Chile and Argentina called the Southern Central Andes. The Northern Andes are characterized by mafic oceanic crust accreted in the Late Cretaceous and Cenozoic and pervasive strike-slip deformation (Gansser, 1973; Alemán and Ramos, 2000; Jaillard *et al.*, 2000), while the Central and Southern Andes are both built on Paleozoic or older ensialic crust and, at present, dominated by near orogen-orthogonal shortening (Mpodozis and Ramos, 1990; Ramos, 2000; Jaillard *et al.*, 2000; Hervé *et al.*, 2000). The Central and Southern Andes differ with respect to their pre-Andean basement ages, Late Oligocene to Recent geologic evolution, crustal thickness and possibly overall crustal composition (Lucassen *et al.*, 2001; Tassara and Yáñez, 2003). Also, climate (Clapperton, 1993; Hartley, 2003) and sediment supply to the trench (Kulm *et al.*, 1977; Thornburg and Kulm, 1987) varies significantly along-strike in the Andes, and this may affect rates of subduction erosion along the continental margin (Shreve and Cloos, 1986), the degree of hydration and consequently the shear stress in the subduction zone (Ruff, 1989; von Huene and Ranero, 2003), the dynamics of subduction and mountain building (Lamb and Davis, 2003), as well as Andean magma chemistry in the different segments of the Andes (Stern, 1991a).

Each segment also has distinct arc-parallel morpho-structural units (Jordan *et al.*, 1983; Kley *et al.*, 1999). In the Central Andes, the western Cordillera Occidental and eastern Cordillera Oriental flank the high Altiplano and Puna plateaus. Volcanoes in the CVZ are largely restricted to the Cordillera Occidental. The Sierra Pampeanas ranges occur in the volcanically inactive segment that separates the CVZ from the SVZ (Fig. 1). In the Southern Andes, the Coastal Cordillera is separated by the Central Valley from the Main Andean Cordillera. The volcanoes of the SVZ and AVZ occur in the Main Cordillera. In the Northern Andes, a western Cordillera Occidental is separated by an intermontane valley from the eastern Cordillera Real (Ecuador) or Central (Colombia), with NVZ volcanoes occurring NVZ on both cordilleras as well as in the intermontane valley.

PRE-ANDEAN BASEMENT

The western part of the Northern Andes in Ecuador and Colombia are underlain by juvenile crust consisting of dominantly mafic igneous rocks accreted during the Late Cretaceous and Cenozoic development of the Andes (Alemán and Ramos, 2000; Jaillard *et al.*, 2000). However, everywhere else in the Andes, pre-Andean basement varies from Proterozoic to Early Mesozoic in age. Basically, Proterozoic crust, which was accreted or possibly rifted and re-accreted to the western margin of the Archean Amazonian cratons during the Paleozoic, underlies a large portion of the CVZ in the Central Andes, while younger Paleozoic to Early Mesozoic rocks underlie the eastern part of the NVZ, the southern part of the CVZ, the SVZ and AVZ.

Paleoproterozoic (2.0-1.8 Ga) metamorphic and igneous rocks of the Arequipa terrane in the Northern Central Andes of southern Perú, below the northern part of the CVZ, are the oldest basement below the Andean arc. Paleoproterozoic protoliths in this terrane may be either parautochthonous (perigondwanan) and related to the Brazilian craton (Tosdal, 1996), or allochthonous and part of the Grenvillian province of Laurentia (Dalziel, 1994, 1997; Wasteneys *et al.*, 1995; Loewy *et al.*, 2004). Metamorphic rocks with possible Paleoproterozic protolith ages have also been reported in the Southern Central Andes at both Cerro Uyarani on the Bolivian Altiplano and in the Antofalla basement exposed at Belen, Chile, implying that such rocks may underlie a large part of the CVZ (Wörner *et al.*, 2000a). Rocks with younger 'Grenvillian' (1.3-1.0 Ga) protolith ages also occur at Choja in northern Chile (Loewy *et al.*, 2004).

These proterozoic rocks were subsequently affected by the Neoproterozoic to Early Paleozoic (680-510 Ma) Pampean igneous and metamorphic event, first identified in the basement further south in the Sierras Pampeanas ranges in Argentina (Rapela *et al.*, 1998a,b), as well as younger Paleozoic (470-340 Ma) events (Rapela *et al.*, 1998a,b; Wörner *et al.*, 2000a; Lucassen *et al.*, 2000; Jaillard *et al.*, 2000; Ramos and Alemán, 2000). In the Northern Central Andes, north of the Arequipa terrane, and in the South Central Andes south of the Antofalla basement, accretion and amalgamation of the Oaxaquia, Famatina, Cuyania, and Chilenia allochthonous terranes took place during the Paleozoic (Ramos *et al.*, 1986; Pankhurst and Rapela, 1998; Ramos and Alemán, 2000; Astini and Dávila, 2002, 2004; Loewy *et al.*, 2004). The latter two terranes extend to the south into the northern part of the Southern Andes. Their southern boundary is the Paleozoic Patagonian terrane, which was obliquely accreted to South America in the Permian (Ramos, 1988; Ramos and Alemán, 2000; Ramos and Aguirre-Urreta, 2000; Pankhurst *et al.*, 2003).

Beginning in the Late Paleozoic, subduction below the western margin of Gondwanaland produced extensive subduction-related followed by extension-related magmatic activity, generating the Late Paleozoic to Early Mesozoic Choiyoi (Kay *et al.*, 1989; Mpodozis and Kay, 1990, 1992) and Chon Aike (Gust *et al.*, 1985; Riley and Knight, 2001) dominantly granitic and ryholite magmatic provinces prior to and in association with the opening of the South Atlantic. The Gondwanan orogeny is also characterized by an extensive Late Paleozoic low temperature-high pressure accretionary belt of mafic metavolcanics and pelitic schists, which outcrop in the Coastal Cordillera of central Chile (Kato, 1985; Hervé, 1988; Godoy and Kato, 1990; Kato and Godoy, 1995; Barra *et al.*, 1998; Pankhurst and Rapela, 1998; Martin *et al.*, 1999; Willner *et al.*, 2000; Hervé *et al.*, 2000, 2003; Bell and Suárez, 2000; Duhart *et al.*, 2001). To the south, somewhat younger Late Paleozoic to Early Mesozoic metamorphic and igneous rocks, such as the Triassic Chonos metamorphic complex (Hervé *et al.*, 1981, 1994; Hervé and Fanning, 2001) and the Madre del Dios terrane (Thompson and Hervé, 2002), were accreted in progressively younger stages to the southwestern margin of Gondwanaland, and these form the pre-Andean basement below the southern SVZ and AVZ in the Patagonian Andes.

In the Northern Andes, the Cordillera Occidental, upon which the westernmost volcanoes of the NVZ are located, consists of autochthonous mafic rocks of Cretaceous age accreted to the continental margin in the Late Cretaceous and Cenozoic (Feininger, 1987; Wallrabe-Adams, 1990; Alemán and Ramos, 2000; Jaillard *et al.*, 2000). To the east, Paleozoic and Mesozoic metamorphic and igneous rocks form the pre-Andean basement below the Cordillera Real in Ecuador and Cordillera Central in Colombia, on which more eastern volcanoes of the NVZ are located. Precambrian rocks outcrop further to the east in the Cordillera Oriental of Colombia, but not below the active arc. No Precambrian rocks outcrop in Ecuador, but Precambrian crust may underlie the eastern part of Cordillera Real and thus the eastern part of the NVZ in Ecuador (Feininger, 1987; Noble *et al.*, 1997; Barrágan *et al.*, 1998).

ANDEAN GEOLOGIC EVOLUTION

The Andean cycle began in the earliest Jurassic in association with the opening of the Southern Atlantic Ocean. Subduction-related magmatic activity had begun along the west coast of the Northern and Central Andes by at least 185 Ma (Pichowaik *et al.*, 1990), and the Andes evolved as a result of variations in the subduction process (James, 1970). The Mesozoic and Early Cenozoic evolution of the Andes is reviewed in Ramos and Alemán (2000), and other recent review papers provide details concerning the evolution of the Northern (Alemán and Ramos, 2000), Northern Central (Jaillard *et al.*, 2000), Southern Central (Ramos, 2000) and Southern Andes (Hervé *et al.*, 2000).

In the context of the active volcanic belts, the most significant tectonic and geological events in the evolution of the Andes have occurred since the Late Oligocene, after the breakup of the Farallon plate into the Cocos and Nazca plates at approximately 27±2 Ma (Cande and Leslie, 1986; Tebbens and Cande, 1997). This resulted in a change from oblique to more nearly orthogonal convergence between the Nazca and South American plates, as well as a greater than two-fold increase in convergence rates (Pardo-Casas and Molnar, 1987; Somoza, 1998), which together produced a more than three-fold increase in trench-normal convergence. This caused changes in subduction geometry which accelerated crustal shortening, thickening and uplift in the Northern Andes (Alemán and Ramos, 2000) and Northern Central Andes (Jordan *et al.*, 1983; Semperé *et al.*, 1990; James and Sacks, 1999; Jaillard *et al.*, 2000; Ramos, 2000), but resulted initially in extension and crustal thinning in the Southern Central and Southern Andes (Muñoz *et al.*, 2000; Jordan *et al.*, 2001). As a result of the increase in convergence rates, magmatic activity also increased nearly all along the Andean chain.

In the Northern Andes (Alemán and Ramos, 2000), the Late Oligocene change in plate convergence rates generated reactivation of strike-slip faults (Litherland and Aspden, 1992; Ferrari and Tibaldi, 1992), mountain building resulting from renewed folding and thrusting, the development of the Cauca depression (Colombia) and Valley Intermedia (Ecuador) between the Cordillera Occidental and Central/Real, as well as the beginning of an important cycle of magmatic activity (Barberi *et al.*, 1988; Lavenu *et al.*, 1992). In the Northern Central Andes a sequence of discrete compressional events, including, among others, the Aymara (28-26 Ma), Pehuenche (25-23 Ma), Quechua 1 (17-15 Ma), Quechua 2 (9-8 Ma), Quechua 3 (7-5 Ma) and Diaguita (3-5 Ma) events, caused crustal shortening and uplift (Allmendinger *et al.*, 1990, 1997; Jordan and Gardeweg, 1989; James and Sacks, 1999; Ramos *et al.*, 2002). Magmatic activity also was renewed, initially in the eastern Cordillera Oriental and subsequently broadening over all the area occupied by the current magmatic belt. Magmatic activity in the Central Andes has continued from the Late Oligocene to Recent times (Coira *et al.*, 1982, 1993; Petford and Atherton, 1985; Solar and Bonhomme, 1990; Hammerschmidt *et al.*, 1992; Peterson, 1999; James and Sacks, 1999; Kay *et al.*, 1999; Wörner *et al.*, 2000b; Victor *et al.*, 2004), except in the northernmost and southernmost Central Andes below where subduction angle decreased during the Miocene and magmatic activity terminated by the Late Miocene.

In the Southern Central Andes and the Southern Andes, Late Oligocene to Miocene extension generated a complex system of inter-connected fore-arc, intra-arc and back-arc basins extending from the eastern margin of the Coastal range into Argentina (Suárez and Emparan, 1995; Muñoz *et al.*, 2000; Jordan *et al.*, 2001; Charrier *et al.*, 2002). In the southern part of the fore-arc basin in south-central Chile, extension and subsidence was associated with Miocene marine sedimentation (Martínez and Pino, 1979; Cisternas and Frutos, 1994; Frutos and Cisternas, 1994). Late Oligocene to Miocene continental basins also formed within and to the east of the Andes in Argentina (Spalletti and Dalla-Salda, 1996; Morata *et al.*, in press). During the Late Oligocene and Miocene, magmatic activity occurred well west of the current position of the magmatic front, within the Coastal Cordillera at the northern end of the Southern Andes (33°S; Wall and Lara, 2001), and to as far west as the Pacific coast at latitude 42°S (Muñoz *et al.*, 2000), where magmatic activity also extended as far east as the Atlantic coast. Late Oligocene to Miocene magmatism and extension in south-central Chile may have resulted from the formation of an asthenospheric slab window as a result of the changes in subduction geometry resulting from the increase in trench normal convergence rates (Muñoz *et al.*, 2000). However, this was a transient event, and extension ended by the Early to Middle Miocene. The sediments and volcanic rocks deposited in the Late Oligocene to Early Miocene basins were deformed and uplifted in the Middle to Late Miocene when the magmatic arc returned to its current location in the Main Cordillera (Godoy *et al.*, 1999; Muñoz *et al.*, 2000; Jordan *et al.*, 2001).

Although many of the main features of the Andes were acquired during the Miocene, Quaternary neotectonic deformation made significant modifications of topography and also controls the location of active volcanoes, and thus the distinction among small arc segments within larger volcanic zones. Regional neotectonic zones are related to the dip of the subducted plate, the obliquity of convergence (Fig. 1; Dewey and Lamb, 1982: Jarrad, 1986), and the thickness and composition of the continental crust (Tassara and Yáñez, 2003). In the southern Southern Andes (39-47°S), oblique convergence of 22-30° northeast of the trench results in slip along the arc-parallel Liquiñe-Ofqui fault system (Cembrano *et al.*, 1996, 2000; Lavenu and Cembrano, 1999a,b; Arancibia *et al.*, 1999). This fault system controls the location of many of the larger volcanic centers as well as monogenetic cones in the southern SVZ (López-Escobar *et al.*, 1995a). In the Southern Central Andes (20-34°S), crustal shortening occurs with only minor arc-parallel slip (Dewey and Lamb, 1982). A transitional zone between these two areas occurs in the northern Southern Andes (39-34°S), where the SVZ is widest and includes extensional intra-arc basins with small monogenetic basaltic cones (Muñoz and Stern, 1988, 1989; Folguera *et al.*, 2002). In the Northern Central Andes of Perú, north of the Arica bend, obliquity is lower (20-24°), and the High Andes are affected by extension and uplift. This is among the areas of highest average altitude in the Andes, and here the Coropuna volcano (6,425 m; 15.3°S), the northernmost volcano in the CVZ, is only 240 km northeast of the trench, which has a depth of 6,865,m below sea level (Jaillard *et al.*, 2000). Further north in Ecuador and Colombia, obliquity of convergence again increases to 31-45° and significant dextral slip occurs both in the region of the forearc and volcanic arc (Dewey and Lamb, 1982). Other specific areas of neotectonic deformation are related to subduction of the Carnegie (Gutscher *et al.*, 1999a), Nazca (Macharé and Ortlieb, 1992; Hampel *et al.*, 2004a, b) and Juan Fernández ridges (von Huene *et al.*, 1997; Yáñez *et al.*, 2001, 2002), and Chile Rise (Lagabrielle *et al.*, 2000, 2004). For instance, Cerro Aconcagua (6,962 m; 32°S), the highest mountain in the Andes, overlies the locus of subduction of the Juan Fernández ridge (Fig. 1).

CRUSTAL THICKNESS

The crust below the CVZ in the Central Andes is >70 km thick below both the western Cordillera Occidental and the eastern Cordillera Oriental flanking the Altiplano and Puna plateaus, and 60-65 km thick below the Altiplano and Puna (James, 1971; Schmitz, 1994; Beck *et al.*, 1996; Dorbath *et *al.*, 1996; Schmitz *et al.*, 1999; Swenson *et al.*, 2000; ANCORP Working Group, 2003). This very thick crust underlies a region 700 km long and 200 km wide which has an average elevation >3,700 m, while many peaks in these two cordilleras reach >6,000 m. Crustal thickness decreases rapidly to the east of the Cordillera Oriental, to 43-47 km below the Subandean zone and 32-38 below the Chaco plain further to the east (Beck *et al.*, 1996). Crustal thickness also decreases gradually to the north, to 65 km at the northern end of CVZ (15°S) and to 40-45 below northern Perú (7°S). Further north, below the NVZ in Ecuador and Colombia the crust is 40-60 km thick below both the western Cordillera Occidental and eastern Cordillera Real and Central (Feininger and Seguin, 1983). Crustal thickness also decreases to the south of the Puna plateau to >55 km below both the Sierras Pampeanas ranges (Jordan *et al.*, 1983; Kay *et al.*, 1999) and the Main Cordillera at the northern end of the SVZ (33°S) in the Southern Andes, 40 km at 39°S (Lüth and Wigger, 2004) and ultimately 30-35 km south of 37°S below the central and southern SVZ and AVZ (Lowrie and Hey, 1981).

The very thick crust below the Central Andes may be due to crustal shortening (Isacks, 1988; Beck *et al.*, 1996; Allmendinger *et al.*, 1997; Kley *et al.*, 1999), magmatic underplating (Tosdal *et al.*, 1984; Schmitz *et al.*, 1997), a combination of the two, with more crustal shortening in the eastern Cordillera Oriental and more magmatic underplating and lithospheric hydration below the western Cordillera Occidental (James, 1971; James and Sacks, 1999; Giese *et al.*, 1999; Victor *et al.*, 2004), or lithospheric delamination (Kay *et al.*, 1994, 1999). Crustal thickening and uplift in the northern Central Andes began during the Eocene (the 'Incaic' tectonic event) as a result of an episode of very rapid oblique convergence (Pardo-Casa and Molnar, 1987), and continued episodically to generate the thick crust below the Central Andes (Mpodozis *et al.*, 1995; Kay *et al.*, 1999; Jaillard *et al.*, 2000). The timing and style of deformation, uplift and magmatism varied from one region of the Central Andes to another (James and Sacks, 1999; Kay *et al.*, 1999). While the Altiplano (James and Sacks, 1999) and northern Puna (Kay *et al.*, 1999) plateaus began to evolve as subduction angle increased after the Eocene event, the Main Andean Cordillera of the Pampean flat-slab segment began to be uplifted as subduction angle decreased (Jordan *et al.*, 1983; Cornejo *et al.*, 1993; Kay *et al.*, 1999; Kay and Mpodozis, 2002; Ramos *et al.*, 2002).

CLIMATE

Two important along-strike variables in the Andes are precipitation and temperature (Clapperton, 1993). These control snow lines on volcanoes, and therefore lahar potential, as well as affecting erosion rates and sediment supply to the trench (Kulm *et al.*, 1977). In the Central Andes of southern Perú and northern Chile, hyper-arid desert conditions along the west coast cause sediment supply to the trench to be very low (Hartley *et al.*, 2000; Hartley, 2003), and the Perú-Chile trench is essentially devoid of sediment (Thornberg and Kulm, 1987). In the northern Andes of Colombia and Ecuador, and the southern Andes of southern Chile, high precipitation rates result in high sediment supply to the trench, which in the southern Andes is enhanced by glacial erosion. In these areas the morphology of the Perú-Chile trench is masked by sedimentary infill. Sediment supply to the trench influences rates of subduction erosion along the continental margin, which increases when sediment supply is low (Shreve and Cloos, 1986; von Huene and Scholl, 1991; Bangs and Cande, 1997; von Huene and Ranero, 2003; Lamb and Davis, 2003), such as at the latitudes of the Central Andes. Rates of subduction erosion and subduction of continental crust in turn affect the degree of hydration (Ranero and Sallarés, 2004) and consequently the shear stress in the subduction zone (Ruff, 1989; von Huene and Ranero, 2003), the dynamics of subduction and mountain building (Lamb and Davis, 1993), as well as Andean magma chemistry in the different segments of the Andes (Stern, 1991a).

OCEANIC PLATES AND SUBDUCTION GEOMETRY

Active volcanoes in the NVZ, CVZ and SVZ result from subduction of the Nazca plate, and those in the AVZ from subduction of the Antarctic plate (Fig. 1). These two plates are separated by the Chile Rise, an active spreading ridge the eastward extension of which has progressively been subducted below the southernmost part of the continent during the late Cenozoic (Cande and Leslie, 1986; Hervé *et al.*, 2000). Ages of subducting Nazca plate range from Recent just north of the Chile Rise at the southern end of the SVZ, to a maximum of 50-60 Ma below the central portion of the CVZ, where the trench also reaches a maximum depth of 8055m below sea level at 23°S, and decrease again to 10-20 Ma below the NVZ north of the Carnegie ridge (Fig. 1). The age of the subducting Antarctic plate increases southward from <12 Ma just south of the Chile Rise to 24 Ma below the southernmost part of the SVZ (Cande and Leslie, 1986; Polonia *et al.*, 1999). Estimated convergence rates between the Nazca and South American plates range from 7-9 cm/yr (Pardo-Casas and Molnar, 1987; Somoza, 1998; Norambuena *et al.*, 1998; Angermann *et al.*, 1999), while that of the Antarctic plate is only 2 cm/yr. Convergence directions between the Nazca and South American plates vary between 20-45° from orthogonal to the trench (Dewey and Lamb, 1982), while convergence between the Antarctic and South American plates varies from close to orthogonal just south of the Chile Rise to highly oblique further south where both the coast and trench bend to the east (Fig. 1),

The dip of the subducting Nazca plate to a depth of 90-100 km is a very uniform ~25° everywhere below the west coast of South America. However, the angle of subduction at greater depths is distinctly different, decreasing to significantly lower in three flat-slab segments with subhorizontal subduction (Fig. 1; Barazangi and Isacks, 1976, 1979; Bevis and Isacks, 1984; Cahill and Isacks, 1992; van der Hilst and Mann, 1994; Engdahl *et al.*, 1995, 1998; Pardo *et al.*, 2002). These three segments, which lack recent volcanic activity, are the Bacaramanga flat-slab segment below northern Colombia north of 5°N, the Peruvian flat-slab segment between the Gulf of Guayaquil (5°S) and Arequipa (14°S), and the Pampean flat-slab segment below central Chile and Argentina (27-33°S). In these regions the slab at depths greater than 100 km dips at a low angle of 5-10° several hundred kilometers to the east, and then subduction angle again increases.

Zone of high-Q, interpreted as asthenosphere, occur below the volcanic arc in the segments with subduction angle remaining relatively high (25-30°) at depths greater than 90-100 km (Barazangi and Isacks, 1976, 1979). The absence of any well developed zones of high-Q, or recent volcanic activity, in the three flat-slab segments suggests that in these segments subducting oceanic plate is occluded to the base of the overriding continental plate without intervening asthenosphere. Shallow level intra-plate seismicity above the flat-slab segments is 3-5 times higher than in the adjacent areas of steeper subduction (Gutscher, 2002), and these segments are characterized by crustal shortening, thickening and uplift, which has produced the Cordillera Blanca in northern Perú and the Sierras Pampeanas ranges east of central Chile in Argentina (Ramos *et al.*, 2002). Based both on the patterns of distribution of volcanic activity and deformation through time in the region of the flat-slab segments, it is clear that they developed only after the Late Oligocene change in plate convergence directions (Kay *et al.*, 1999; James and Sacks, 1999).

Transitions between areas of low angle and steeper subduction are either gradual or associated with sharp bends, but not breaks, in the subducting plate (Bevis and Isacks, 1984; Cahill and Isacks, 1992; Engdahl *et al.*, 1998). Sharp bends in the subduction angle occur at the locus of subduction of the Nazca and Juan Fernández ridges (Fig. 1), and subduction of these, as well as the Carnegie ridge and the Inca plateau, may play a role in the segmented geometry of plate subduction (Pilger, 1981, 1984; von Huene *et al.*, 1997; Gutscher *et al.*, 1999a,b, 2000; Yáñez *et al.*, 2001, 2002; Gutscher, 2002). However, decreasing subduction angle below the southern portion of the CVZ and the northern boundary of the Pampean flat-slab segment is gradual and may result from bending of the slab as a consequence of a greater amount of crustal shortening below the CVZ (Isacks, 1988; Tinker *et al.*, 1996). Subduction of oceanic transform faults might also play a role in controlling the location and compositions of volcanoes in the active arc (Herron, 1981; Hall and Wood, 1985; López-Escobar *et al.*, 1995a), and the division between small segments of arc within the four larger volcanic zones.

SUBDUCTION EROSION

A notable aspect of the western margin of South America is the paucity of Mesozoic and Cenozoic arc-trench gap accretionary complexes, and the occurrence of pre-Andean Paleozoic and Early Mesozoic exotic to parautochthonous terranes along the coast (Rutland, 1971; Kulm *et al.*, 1977; Schweller and Kulm, 1978). This is interpreted to result from subduction erosion, or rasping off of the continental margin by a combination of both continental slope retreat and erosion of the underside of the upper plate (Rutland, 1971; Scholl *et al.*, 1980; Ziegler *et al.*, 1981; Stern, 1991a; von Huene and Lallemand, 1990; von Huene and Scholl, 1991, 1993; Lallemand *et al.*, 1992a,b; Lallemand, 1995; von Huene *et al.*, 1999, 2004; von Huene and Ranero, 2003, Kay *et al.*, 1999, in press). Other geologic evidence for subduction erosion along the west coast of South America, particularly west of the Central Andes where extent of subduction erosion has been greatest, include the ~250 km eastward migration of the Andean Jurassic to Cenozoic magmatic belts in the Central Andes (Ziegler *et al.*, 1981; Peterson, 1999), the northwest strike and almost complete disappearance north of 27°S of the Late Paleozoic Gondwanan subduction accretionary complexes that form the Coastal Cordillera in central Chile (Stern, 1991a; Stern and Mpodozis, 1991) and shortages in the amount of crustal shortening that can be accounted for by crustal area balance calculations (Allmendinger *et al.*, 1990; Schmitz, 1994; Kay *et al.*, 2004). Also, the Upper Triassic Cifuncho Formation, which occurs along the coast of the Southern Central Andes in northern Chile at 25.5°S, consists of terrigenous conglomerates and sandstones derived from a Paleozoic source once located to the west off the current coast, but now is nowhere to be found (Suárez and Bell, 1994).

The suggested mechanism for slope retreat is continental margin collapse towards the trench caused by extensional faults, generating debris that are pushed into the horst and graben structures on the subducting oceanic plate, which resemble the teeth of a chain saw (Hilde, 1983; von Heune *et al.*, 2004). Such structures and processes have been well documented along the continental margin of southern Perú (Lallemand *et al.*, 1992a, b; Hampel *et al.*, 2004a, b) and northern Chile (Delouis *et al.*, 1996; von Huene *et al.*, 1999; Hartley *et al.*, 2000; von Huene and Ranero, 2003; González *et al.*, 2003).

Rates of subduction erosion depend on factors such as subduction angle, sediment supply to the trench, and subduction of buoyant features such as ridges (Sholl *et al.*, 1980; Shreve and Cloos, 1986; Bangs and Cande, 1997; Lamb and Davis, 2003). Both low angle subduction and low sediment supply to the trench, features of the plate boundary west of the Central Andes, are factors that enhance the potential for increased rates of subduction erosion. Here the Perú-Chile trench receives almost no sediment from the few mountain glacier-fed drainage systems that cross the Atacama desert (Thornberg and Kulm, 1987; Hartley, 2003). Long-term average rates of subduction erosion west of the Central Andes may be as high as 40 km^3 of crust *per* million years *per* kilometer of trench, in order to have removed an estimated 250 km of 40 km thick crust in the last 25 million years of near orthogonal convergence (Zeigler *et al.*, 1981; Lallemand, 1995). This is equivalent to 4% of the volume of subducted oceanic crust in the same time period, which is well within the carrying capacity of underthrusting grabens entering the subduction zone below northern Chile (von Huene and Ranero, 2003). This subducted crust may have an important influence on the extent to which the subducted slab is hydrated (Ranero and Sallarès, 2004), the dynamics of the interaction between the subducting oceanic and overlying continental plate (Lamb and Davis, 2003), and on the chemistry of Andean magmas (Stern, 1991a). Subduction of buoyant features such as ridges also enhances rates of tectonic erosion, as documented for the Nazca (von Huene, Suess *et al.*, 1988; Hampel *et al.*, 2004a, b) and Juan Fernández ridges (von Huene *et al.*, 1997; Yáñez *et al.*, 2001, 2002), as well as the Chile Rise (Cande and Leslie, 1986; Bourgois *et al.*, 1996, 2000; Polonia *et al.*, 1999).

VOLCANIC ARC SEGMENTATION

NVZ (5°N-2°S)

The NVZ (Fig. 2) consists of 19 volcanoes in Colombia (Méndez Fajury, 1989) and 55 in Ecuador (Hall, 1977; Barberi *et al.*, 1988). Among the largest, more recently active and/or better known of these are Nevado del Ruiz (5,390 m; Williams, 1990), Huila (5,750 m), Galeras (4,482 m; Stix *et al.*, 1997) and Cumbal (4,764 m) in Colombia, and Mojanda (4,263 m; Robin *et al.*, 1997), Guagua Pichincha (4,784 m), Cayambe (5,790 m; Samaniego *et al.*, 1998), Reventador (3560m; Hall *et al.*, 2004), Antisana (5,753 m; Barrágan *et al.*, 1998), Cotopaxi (5,911 m; Barberi *et al.*, 1995; Mothes *et al.*, 1998), Chimborazo (6,300 m; Kilian *et al.*, 1995), Tungurahua (5,020 m) and Sangey (5,230 m) in Ecuador.

The volcanoes of the NVZ are distributed in north-south belts along two mountain chains, the western Cordillera Occidental and the eastern Cordillera Central (Colombia) or Cordillera Real (Ecuador), separated by an intermontane valley called the Cauca-Patria depression (Colombia) or Valle Intermedio (Ecuador). In Colombia the volcanoes are divided into three separate groups termed the north, central and south segments, with a distance of >150 km separating the north and central segments. Volcanoes in the northern and central segments are located in the Cordillera Central east of the central depression, while those in the southern segment also occur within the Cauca-Patria depression and in the Cordillera Occidental. In Ecuador four separate sub-parallel north-south belts are recognized. The western belt of volcanoes along the volcanic front forms a nearly continuous chain in the Cordillera Occidental. These large volcanic centers, and fields of small cones and flows in the Valle Intermedio, connect to the north into the southern Colombian segment. Those further to the east in the Cordillera Real are more widely and irregularly separated and occur in the same tectonic position, east of the central depression, as the central and northern segments in Colombia. Two large calderas, Chacana and Chalupas, occur in this belt. Three other alkaline volcanic centers also occur 50 km east of the Cordillera Real and ~150 km east of the volcanic front, in what may be considered the back-arc region.

Below the NVZ, 12-20 Ma Nazca oceanic lithosphere is being subducted at an angle of 31-45° northeast of orthogonal with the trench, at 7 cm/yr and with a dip of 25-30° (Gutscher *et al.*, 1999a). The Perú-Chile trench west of the NVZ is filled with sediment. The northern end of this segment corresponds with a break in subduction angle, which decreases to the north below the Bucaramanga flat-slab segment (Pennington, 1981; Hall and Wood, 1985; van der Hilst and Mann, 1994). The southern part of this segment occurs above the region of subduction of the Carnegie Ridge, where subduction angle may also be somewhat lower (Barberi *et al.*, 1988; Gutscher *et al.*, 1999a). The southern end of this arc segment correspond to the landward extensions of the northeast-southwest trending Alvarado and Sarmiento Fracture Zones which bound the Gulf of Guayaquil (Hall and Wood, 1985).

Despite the rather uniform convergence rate, age and subduction angle of the Nazca oceanic plate, the distance from the trench to the arc, the depth of the subducted oceanic lithosphere below the arc, and the age and thickness of the continental crust vary significantly below different parts of the NVZ. The arc front in the northern segment of Colombia occurs up to >380 km east of the trench and these volcanoes occur 140-160 km above the subducted slab. The distance from the trench to the arc front decreases southward to ~300 km in the southern part of the NVZ in Ecuador, and here arc front volcanoes are only 80-100 km above the slab. In conjunction with these changes, the arc varies from a single chain of volcanoes in the northern segment in Colombia to a broad belt between 80-120 km wide in Ecuador, where alkaline back-arc centers also extend another 50 km to the east. Crustal thickness ranges from ~40 to possibly >55 km, with thinner crust below the western Cordillera Oriental and thicker crust below the intermontane valley and Cordillera Central/Real to the east (Feininger and Seguin, 1983). Basement ages and lithologies also vary very significantly from relatively young Cretaceous accretionary mafic crust below the western part of the arc to older Paleozoic

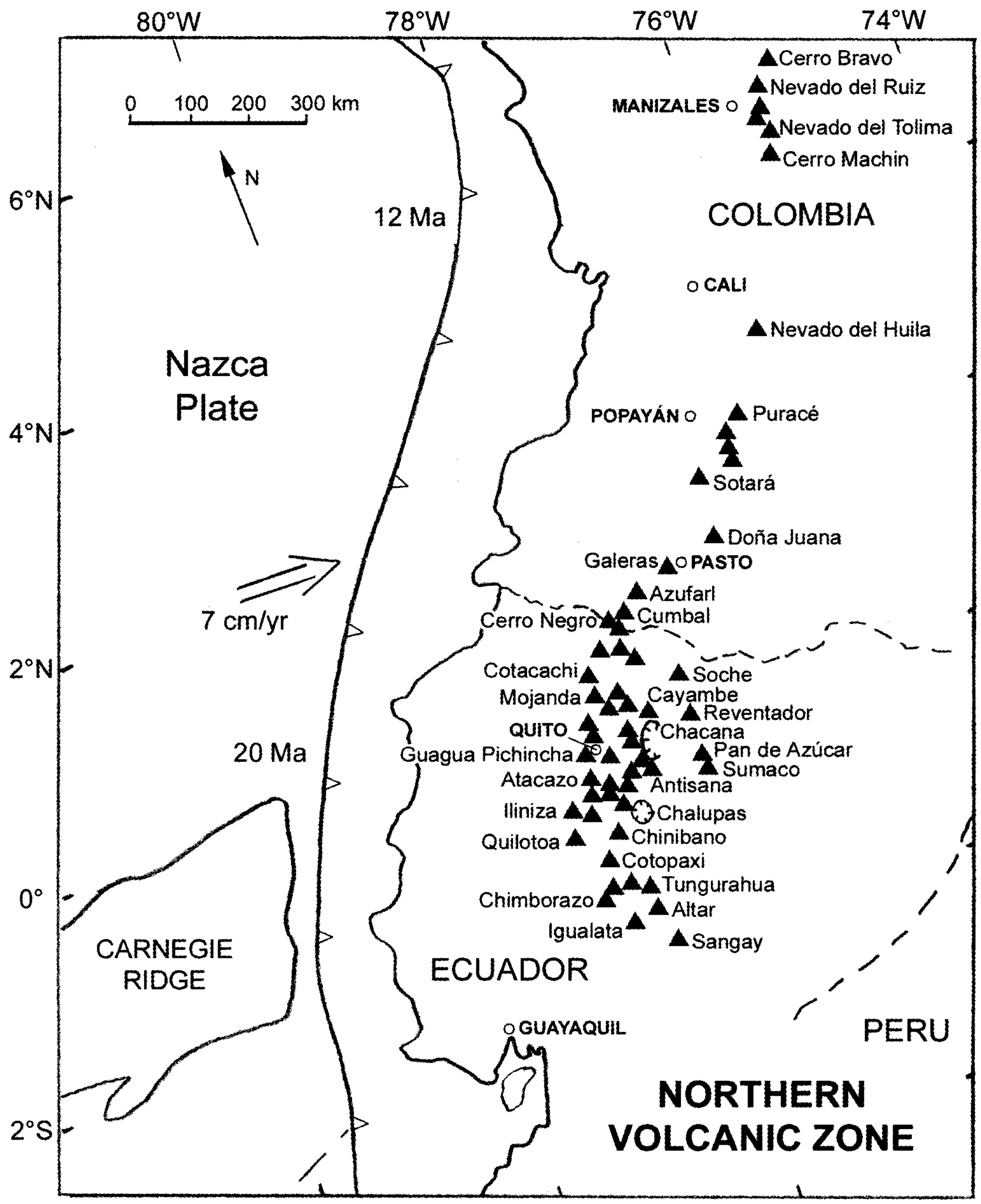

FIG. 2. Schematic map of the Nothern Volcanic Zone showing the location of some of the better known volcanoes mentioned in the text (Hall, 1977; Barberi *et al.*, 1988; Méndez Fajury, 1989). Details of oceanic plate structures and motions can be found in Gutscher *et al.* (1999a), and of the geology and morpho-structural segmentation of the Northern Andes *in* Hall and Wood (1985) and Alemán and Ramos (2000).

metamorphic rocks and Mesozoic granitoids below the eastern part of the arc (Feininger, 1987), east of the Dolores-Guayaquil Megashear (Alemán and Ramos, 2000). Further to the east, granulites of the Guyana shield may underlie the back-arc volcanic centers in Ecuador (Feininger, 1987; Barrágan *et al.*, 1998), although here the crust is thinner than below the Cordillera Real (Feininger and Seguin, 1983).

PERUVIAN FLAT-SLAB SEGMENT (5-14°S)

This flat-slab segment extends from just south of the region of subduction of the eastward extension of the Carnegie Ridge below southern Ecuador to the locus of subduction of the Nazca ridge below central Perú (Fig. 1). The northern boundary of this flat-slab segment corresponds to the boundary between the northeast-southwest trending Northern Andes and the northwest-southeast Northern Central Andes. This boundary occurs across major structures within the continental crust, such as the Dolores-Guayaquil megashear in Ecuador and the Huancabamba deflection in Perú, which separate Cretaceous mafic accretionary crust in the north from older Paleozoic ensialic crust intruded by the Peruvían coastal batholith to the south. At the southern end of this flat segment, the Abancay or Pisco deflection separates Paleozoic accretionary crust to the north from older Proterozoic rocks in the Arequipa massif to the south, and marks the northern limit of the Altiplano in Perú. The transition from low angle to steeper angle subduction across this boundary is associated with an abrupt bend, but not a tear, in the subducted Nazca plate (Hasagawa and Sacks, 1981; Engdahl *et al.*, 1995, 1998). Magmatic activity occurred across this entire segment prior to ~10 Ma, but decreased in intensity, progressively from north to south in conjunction with the southward migration of the locus of subduction of the Nazca ridge (Hampel, 2002), and ultimately terminated between 5 and 2 Ma (Petford and Atherton, 1995; Peterson, 1999).

CVZ (14-27°S)

The CVZ includes 44 active volcanic edifices, as well as more than 18 active minor centers and/or fields and at least six potentially active Quaternary large silicic ignimbrite center and/or caldera systems (de Silva, 1989a, b; de Silva and Francis, 1991), located in the highly elevated (>4,000 m) region of southern Perú, northern Chile, southwestern Bolivia and northwestern Argentina (Fig. 3). Among the more active and better known volcanic centers in the CVZ are Coropuna-Solimana (6,377 m; Vatin-Pérignon *et al.*, 1992), Sabancaya (5,967 m; Gerbe and Thouret, 2004), Chachani (6,057 m), Misti (5,822 m; Thouret *et al.*, 2001) and Huaynaputina (4,200 m; Thouret *et al.*, 1999, 2002; Adams *et al.*, 2001) in Perú, Tata Sabaya in Bolivia (de Silva *et al.*, 1993), and Parinacota (6,348 m; Wörner *et al.*, 1988, 1992a; Clavero *et al.*, 2002, 2004a), Taapaca (5,850m; Clavero *et al.*, 2004b), Nevados de Payachata (Wörner *et al.*, 1988), San Pedro (6,145 m; Francis *et al.*, 1974; O'Callaghan and Francis, 1986), Láscar (5,550 m; Matthews *et al.*, 1994, 1997, 1999; Gardeweg *et al.*, 1998), Socompa (6,051 m; Francis *et al.*, 1985; van Wyk de Vries *et al.*, 2001), Llullaillaco (6,739 m; Gardeweg *et al.*, 1984; Richards and Villeneuve, 2004), Lastarria (Naranjo, 1985, 1992), Ollagüe (5,863 m; Feeley *et al.*, 1993; Feeley and Davidson, 1994), and Ojos del Salado (6,887 m; Baker *et al.*, 1987), the world's highest volcano, in Chile. Giant potentially active Quaternary silicic ignimbrite centers and caldera systems include Cerro Panizos (Ort, 1993; Ort *et al.*, 1996), Pastos Grandes (Baker, 1981), Cerro Guacha (Francis and Baker, 1978), and La Pacana (Gardeweg and Ramírez, 1987; Lindsay *et al.*, 2001a, b; Schmitt *et al.*, 2001, 2002) within the Altiplano-Puna Volcanic Complex (de Silva, 1989a,b), as well as the Los Frailes ignimbrite plateau in Bolivia (Schneider, 1987) and Cerro Galán (Francis and Baker, 1978; Francis *et al.*, 1989; Sparks *et al.*, 1989) and Incapillo (Kay and Mpodozis, 2000) caldera complexes in Argentina. Almost all the large ignimbrite eruptions from these centers are >1 Ma (de Silva, 1989a, b; de Silva and Francis, 1991).

Below the CVZ the <60 Ma Nazca plate lithosphere is being subducted at 7-9 cm/yr, in a direction that varies from 20 to 24° southeast of orthogonal with the trench below the northern CVZ in southern Perú, to 27° northeast of orthogonal with the trench in the southern CVZ of northern Chile (Fig. 3). The subducting slab descends at a dip of 25° to a depth of >400 km (Dorbath *et al.*, 1996). The volcanic front is located in the western Cordillera Occidental, 120 km above the subducted slab and 240-300 km east of the trench, which reaches a maximum depth of 8,055 m below sea level at 23°S. Crustal thickness below the CVZ reach >70 km, and basement ages range from as old as ~2000 Ma below the northern part of the segment in Perú and northernmost Chile and Bolivia, to late Precambrian and Paleozoic below the southern part of this segment in northern Chile and Argentina. The northern end of the CVZ coincides with the locus of subduction of the Nazca Ridge and the Abancay

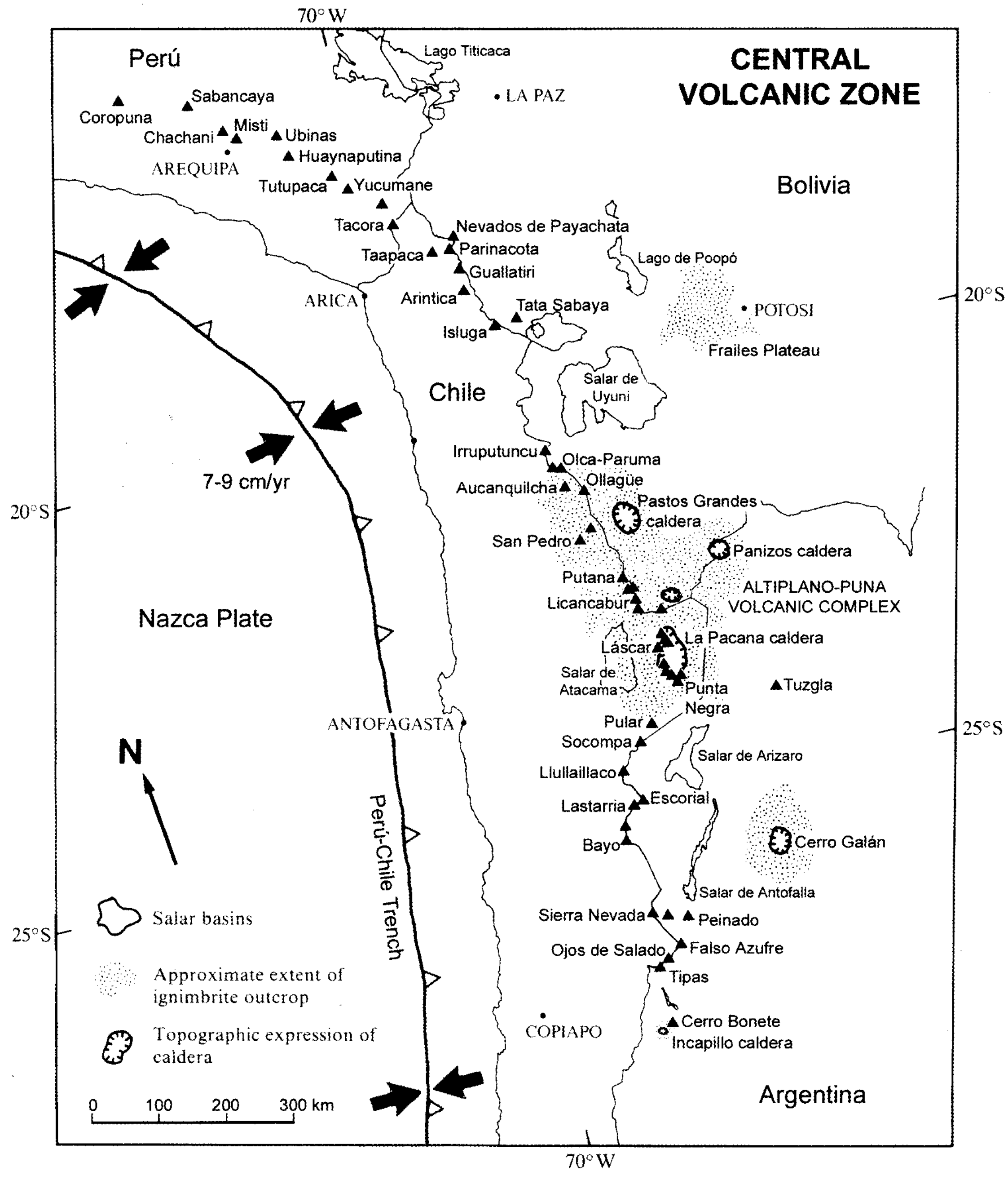

FIG. 3. Schematic map of the Central Volcanic Zone, modified from de Silva and Francis (1991), showing the location of some of the better known volcanoes and caldera systems mentioned in the text. Minor centers are not shown. Further details concerning other aspects of the volcanic arc can be found *in* González-Ferrán (1994), and of the morpho-structural segmentation of the Central Andes *in* Isacks (1988).

Deflection, while the southern boundary coincides with a seismic discontinuity at 27°S (González-Ferrán *et al.*, 1985) superimposed on a gradual decrease in subduction angle to the south (Cahill and Isacks, 1992).

The centers along the volcanic front form a nearly continuous chain within the Cordillera Occidental, although >100 km gap in this chain occurs between the Isluga (19°S) and Irruputuncu (21°S) volcanoes (Fig. 3). The minor centers and giant caldera systems are located both in the Cordillera Occidental and to the east in the Cordillera Oriental, which borders the eastern edge of the Altiplano and Puna in Bolivia and Argentina. The CVZ has been divided into two segments by de Silva and Francis (1991), one oriented northwest-southeast in southern Perú and the other oriented north-south in northern Chile, and into nine separate segments by Wood *et al.* (1987) based on separation of volcanic centers and orientation of centers along and behind the volcanic front. The active volcanoes in the CVZ overlie volcanic rocks of Late Oligocene to Quaternary age, including large ignimbrite sheets, stratovolcanoes and caldera systems (Coira *et al.*, 1982, 1993; de Silva, 1989a, b; Mpodozis *et al.*, 1995, 1996; Wörner *et al.*, 2000b; Lindsay *et al.*, 2001a; Schmitt *et al.*, 2002; Caffe *et al.*, 2002). Many of these older volcanic centers are extremely well preserved because of the hyper-arid conditions in the region. Based on the concentrations of the older centers, de Silva and Francis (1991) suggest that onset of volcanism in the CVZ was earlier in the south and more recent in the north.

PAMPEAN FLAT SLAB SEGMENT (27-33°S)

The subduction angle below the CVZ in central Chile decreases gradually to the south and no break in the subducted plate occurs between the southern boundary of this segment and the less steeply dipping plate below the Pampean flat-slab segment (Bevis and Isacks, 1984; Cahill and Isacks, 1992). In contrast, the southern boundary of this flat-slab segment, which occurs at the locus of subduction of the Juan Fernández Ridge, is a sharp bend, but not a break, in the subducting slab, with steeper subduction angle below the northern portion of the SVZ (Bevis and Isacks, 1984; Cahill and Isacks, 1992). High intra-plate seismicity along this boundary result from the subduction of the eastern prolongation of the Juan Fernández ridge (Yáñez *et al.*, 2001, 2002; Pardo *et al.*, 2002) The locus of subduction of this ridge has migrated from north to south below the Pampean flat-slab segment during the late Cenozoic (Pilger, 1981, 1984; Yáñez *et al.*, 2001, 2002), which may in part be the cause of the low subduction angle in this region. Late Oligocene to Late Miocene widening and ultimately eastward migration of the arc in response to decreasing subduction angle has been documented in considerable detail (Cornejo *et al.*, 1993; Mpodozis *et al.*, 1995; Kay and Abruzzi, 1996; Kay *et al.*, 1999; Kay and Mpodozis, 2002; Ramos *et al.*, 2002). The locus of Late Miocene volcanic activity in this segment was concentrated in the Pocho volcanic belt in Argentina, ~500 km east of the Early Miocene volcanic arc (Kay and Gordillo, 1994), although minor igneous activity may have persisted locally in the High Cordillera until the Late Pliocene (Bissig *et al.*, 2002). Shallowing of the subducted slab was accompanied by crustal deformation and thickening, thinning and hydration of the lithosphere, and substantial loss of the asthenosphere wedge, which occurred in a sequence of tectonic episodes at 18 to 16 Ma, ~10 Ma, 7 to 5 Ma and ~2 Ma (Kay *et al.*, 1999; Kay and Mpodozis, 2002; Ramos *et al.*, 2002).

SVZ (33-46°S)

The SVZ includes, at least, 60 historically and potentially active volcanic edifices in Chile and Argentina, as well as three giant silicic caldera systems and numerous minor eruptive centers (Fig. 4). Among the larger, more active and/or better known volcanic centers in the SVZ are Tupungato (6,650 m; Hildreth and Moorbath, 1988), the northernmost and highestest center in SVZ, San José-Marmolejo (5,830 m; López-Escobar *et al.*, 1985), Maipo stratovolcano (5,290 m)-Diamante caldera complex (Fig. 5; Stern *et al.*, 1984a), Planchón (4,090 m; Tormey *et al.*, 1995; Naranjo and Haller, 2002), Calabozos caldera (Hildreth *et al.*, 1984), Cerro Azul-Quizapu (3,788 m; Hildreth and Drake, 1992), Laguna del Maule (Frey *et al.*, 1984), Tatara-San Pedro (3,621 m; Davidson *et al.*, 1987, 1988; Dungan *et al.*, 2001), Tromen (>4,000 m; Llambías *et al.*, 1982), Payún-Matrú (Llambías, 1966), Longaví (3,242 m; Sellés *et al.*, 2004), Chillán (3,212 m; Dixon *et al.*, 1999; Naranjo and Lara, 2004), Antuco (2,987 m; López-Escobar *et al.*, 1981), Lonquimay (2,822 m; Moreno and Gardeweg, 1989; Naranjo *et al.*, 1992), Copahue stratovolcano (2,952 m; Varekamp *et al.*, 2001; Naranjo and Polanco, 2004) and Caviahue caldera complex (Fig. 6; Muñoz and Stern, 1988; Folguera and Ramos, 2000), Sollipulli (Naranjo *et al.* 1993a; Gilbert *et al.*, 1996), Llaima (3,124 m; Naranjo and Moreno, 1991), Villarrica (2,840 m; Clavero and Moreno, 1994; Petit-Breuilh, 1994; Hickey-Vargas *et al.*, 1989), Lanín (3,778 m; Lara *et al.*, 2004a), Puyehue (2,240 m; Gerlach *et al.*, 1988), Cordón Caulle (2,236 m; Lara *et al.*, in press), Tronador (3,565 m; Mella *et al.*, in press), Osorno (2,661 m; López-Escober *et al.*, 1992; Petit-Breuilh, 1999), Calbuco (2,015 m; Hickey-Vargas *et al.*, 1995; Lopéz-Escobar *et al.*, 1995b), Michinmahuida (2,404 m; Naranjo and Stern, 2004) and, finally, Hudson (Naranjo *et al.*, 1993b; Naranjo and Stern, 1998), the southernmost volcano in the SVZ. In contrast to the CVZ, the calderas in the SVZ have all formed in the last <1.1 Ma.

The northern end of the SVZ coincides with the locus of subduction of the Juan Fernández Ridge and the southern end with the Chile Rise. All along the SVZ, the 0-45 Ma Nazca plate is being subducted below the continent at 7-9 cm/yr in a direction 22-30° northeast of orthogonal with the trench. Subduction angle increases from ~20° at the northern end of the SVZ to >25° further to the south. As a consequence, the distance from the trench to the volcanic front decreases from >290 km in the north to <270 km in the south, and the depth to the subducted slab below the front decreases from 120 to 90 km. Crustal thickness also decreases southwards from >50 km below the northern end of the SVZ to approximately 30-35 km below the southern end. Pre-Andean basement ages range from Paleozoic to early Mesozoic (Munizaga *et al.*, 1988; Nelson *et al.*, 1999).

The SVZ has been divided into the northern (NSVZ, 33-34.5°S; Stern *et al.*, 1984b; Hildreth and Moorbath, 1988), transitional (TSVZ, 34.5-37°S; Tormey *et al.* 1991; Dungan *et al.*, 2001), central (CSVZ, 37-41.5°S; Hickey-Vargas *et al.*, 1984, 1986, 1989; Lopéz-Escobar *et al.*, 1995a) and southern (SSVZ, 41.5-46°S; López-Escobar *et al.*, 1993; Naranjo and Stern, 2004) segments (Fig. 4). A prominent tectonic feature in the CSVZ and SSVZ is the 1000 km long, ~N10°E tending Liquiñe-Ofqui fault zone (López-Escobar *et al.*, 1995a; Cembrano *et al.*, 1996, 2000; Lavenu and Cembrano, 1999a,b), which, along with northeast-southwest and northwest-southeast lineaments, control the location of some of the larger stratovolcanoes and hundreds of small monogenetic Holocene minor eruptive centers in the CSVZ and SSVZ.

The Northern SVZ consists of only three volcanic complexes - Tupungato-Tupungatito, Marmolejo-San José, and Maipo forming a short, narrow north-south chain located on the continental divide along the Chile-Argentina border 290 km east of the trench. At the latitude of the NSVZ the volcanic front migrated eastward over 40 km to its current position during the Pliocene in conjunction with a decrease of the subduction angle below the northern margin of the SVZ as a consequence of the subduction of the Juan Fernández ridge (Stern, 1989; Stern and Skewes, 1995).

The Transitional SVC includes volcanoes in both Chile and Argentina in an up to >200 km wide belt in which arc stratovolcanoes occur on uplifted pre-volcanic basement blocks separated by inter-arc extensional basins containing numerous mono-genetic basaltic cones and lava flows in a complex transition between subalkaline arc and alkaline back-arc volcanism (Muñoz and Stern, 1988, 1989; Bermúdez and Delpino, 1989; Folguera *et al.*, 2002). The volcanic front in the TSVZ is located west of the continental divide in Chile, 270-280 km east of the trench, and trends more northeast-southwest than the NVSZ. The southern boundary of the TSVZ may correspond to the subducted eastward extension of the Mocha Fracture Zone on the Nazca plate (Herron, 1981; López-Escobar *et al.*, 1995a). North of this fracture zone relatively older >35 Ma oceanic crust is being subducted. The TSVZ volcanoes overlie Miocene and Pliocene volcanic rocks deposited on Paleozoic and Mesozoic basement, and the Holocene volcanoes formed on or near Pliocene and Pleistocene centers (Drake, 1976) such as Cerro Campanario (35.9°S; Hildreth *et al.*, 1998).

The northern portion of the Central SVZ is also a broad arc, up to 120 km wide, with intra-arc basins and arc volcanoes in both Chile and Argentina. However, south of 39°S, the latitude of the landward project of the Valdivia Fracture Zone (Herron, 1981), south of which younger <18 Ma oceanic lithosphere is being subducted, the arc in the CSVZ narrows into a chain approximately 80 km wide without intra-arc

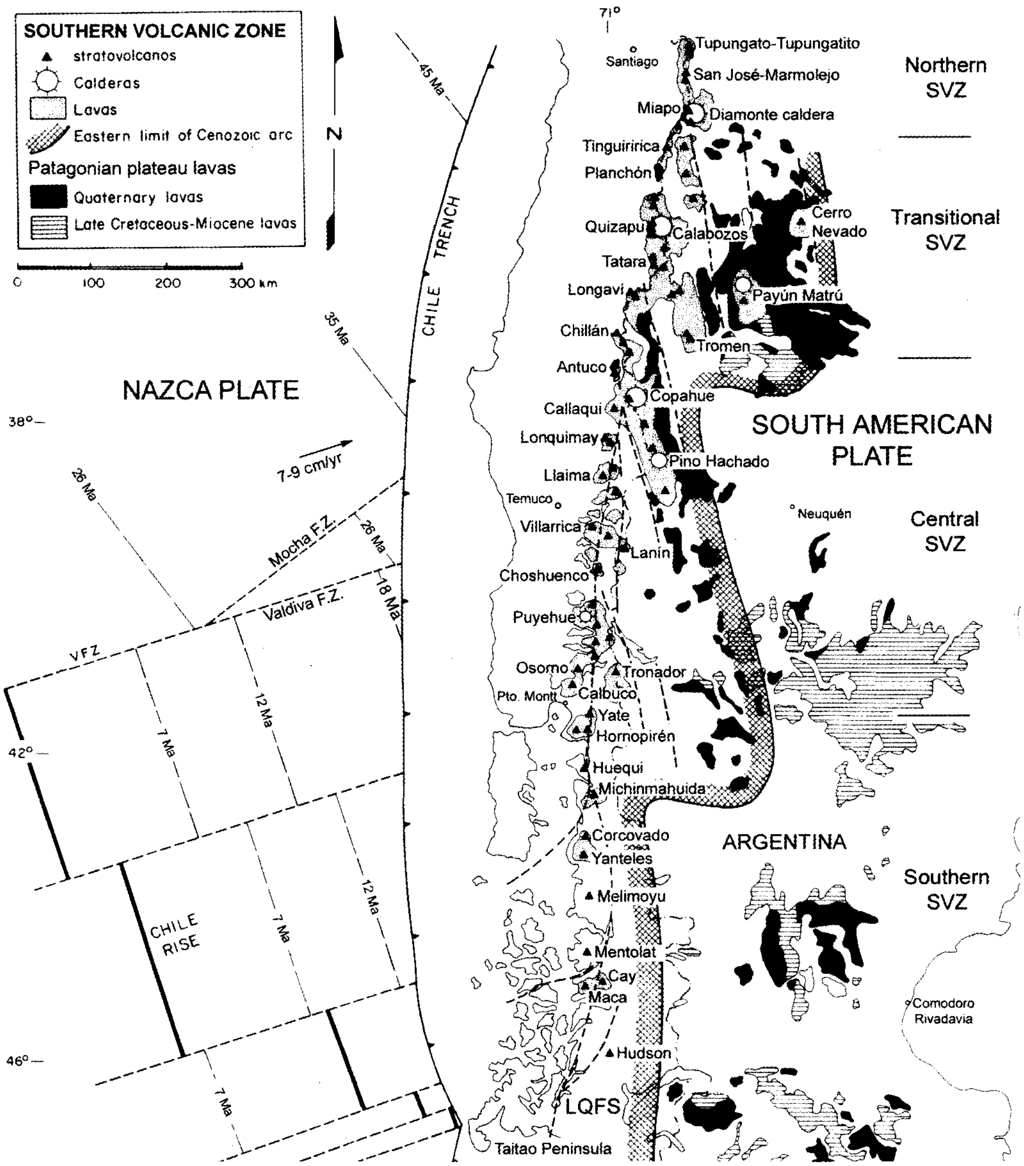

FIG. 4. Schematic map of the Southern Volcanic Zone, modified from Stern *et al.* (1990), showing the location of some of the better known volcanoes and caldera systems mentioned in the text. Note the significantly wider arc in the Transitional SVZ segment north of 37°S (Muñoz and Stern, 1988, 1989; Bermúdez and Delpino, 1989). Dark cross-hatched line indicates the eastern limit of Cenozoic arc volcanism, which divides the area of transitional (to the west) from cratonic (to the east) alkali olivine back-arc basalts (Fig. 8) of the Patagonian plateau lavas. The approximately north-south dashed line labeled LQFS is the Liquiñe-Ofqui fault system. Further details of the volcanoes and morpho-structural segmentation of the SVZ can be found *in* González-Ferrán (1994) and López-Escobar *et al.* (1995a).

basins. The volcanic front in the CSVZ occurs in Chile west of the continental divide, and in the southern part of the CSVZ along the boundary of the Central Valley and the western edge of the Main Cordillera. The volcanic front of the CSVZ either migrated to the west to its current position in the Late Pleistocene (Stern, 1989), or the amplitude of volcanic activity along the current front increased significantly at this time (Lara *et al.*, 2001). The eastern part of the arc, east of the current volcanic

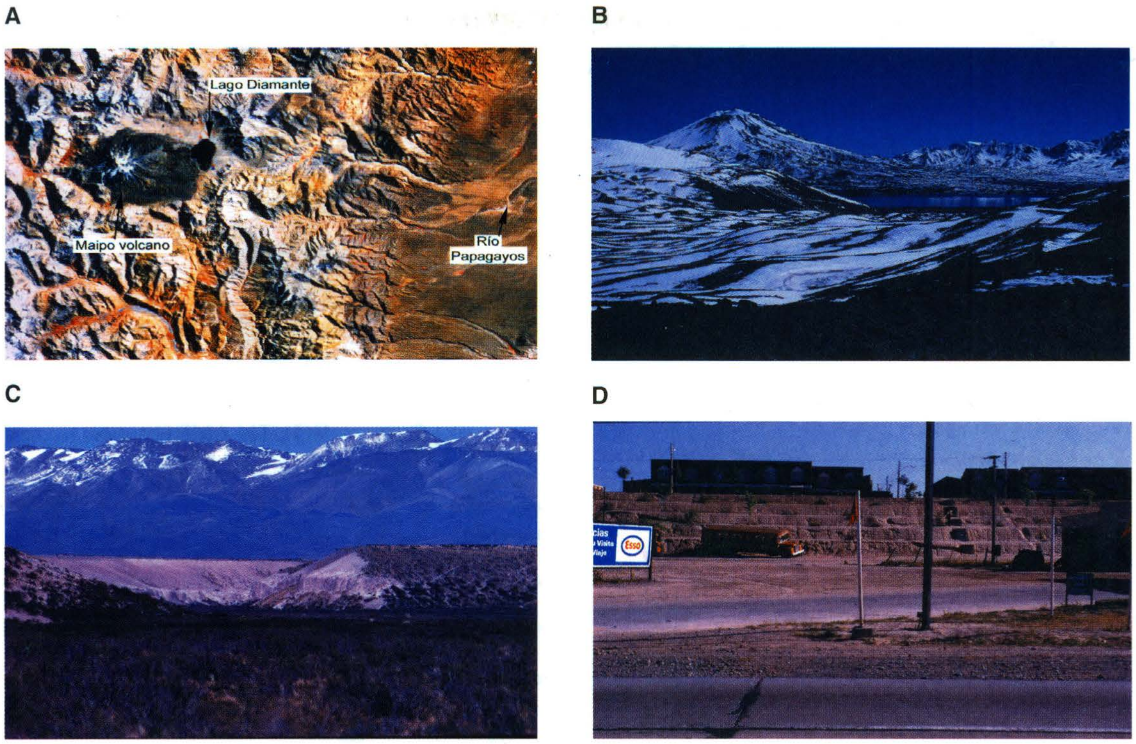

FIG. 5. A. Satellite photo of the Maipo volcano and Diamante lake inside the elliptical 15 by 20 km Diamante caldera. B. The Maipo volcano and Diamante lake inside the Diamante caldera wall. C. Partially welded rhyolite tuffs, derived from the Pleistocene eruption of the Diamante caldera, which outcrop along the Papagayos river in Argentina. D. The 30 m thick Tuff of Pudahuel, also derived from the Pleistocene eruption of the Diamante caldera, outcropping along the road to the international airport west of Santiago, Chile. The Diamante caldera formed at 450 ± 60 ka, and the volume of rhyolite tuff erupted in this event is estimated as approximately 450 km^3 (Stern *et al.*, 1984a).

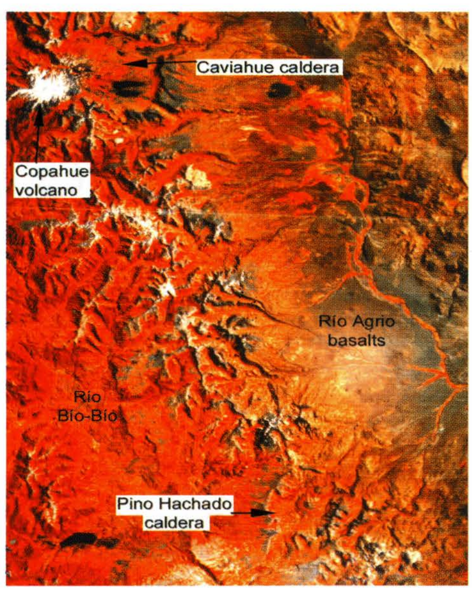

FIG. 6. Satellite photograph of the Copahue volcano inside the 1.1 Ma Caviahue caldera located at the northwest end of an uplifted basement block on which also occurs the Pino Hachado caldera (Muñoz and Stern, 1988, 1989; Folguera and Ramos, 2000; Naranjo and Polanco, 2004).

front, was active in the Pliocene and Early Pleistocene (Muñoz and Stern, 1988, 1989; Mella *et al.*, en prensa), and overlies Late Miocene volcanic rocks.

The Southern SVZ consists of 13 volcanic centers, all located in Chile west of the continental divide and <270 km east of the trench (López-Escobar *et al.*, 1993; D'Orazio *et al.*, 2003; Naranjo and Stern, 2004). SSVZ volcanoes overlie deeply eroded Paleozoic metamorphic rocks intruded by Mesozoic and Cenozoic plutons, and in some cases Miocene and Pliocene volcanic rocks. Michimauhida and Hudson volcano (Orihashi *et al.*, 2004), and possibly Yate, and Hualaihué as well, were active in the Late Pleistocene. Michinmauhida and Hudson volcanoes are the only ones with historic eruptions (Naranjo and Stern, 1998, 2004).

PATAGONIAN VOLCANIC GAP (46-49°S)

A gap in active volcanism occurs between 46-49°S in the region south of the Chile Rise-Trench triple junction where the southeast extension of the Chile Rise has been subducted during the last ~8 million years (Fig. 7; Herron *et al.*, 1981; Cande and Leslie, 1986; Cande *et al.*, 1987; Behrmann *et al.*, 1994; Tebbens and Candie, 1997; Tebbens *et al.*, 1997; Murdie *et al.*, 2000; Lagabrielle *et al.*, 2004). No Benioff zone of seismic activity occurs below this volcanic gap. The last manifestation of subduction-initiated magmatic activity in this region are 12 Ma adakitic andesites of the Cerro Pampa volcanic complex, which occur in Argentina well east of the north-south strike of the current arc (Fig. 7; Kay *et al.*, 1993). As a consequence of ridge subduction, arc volcanism terminated and an asthenospheric slab-window resulted in the development of an extensive back-arc alkaline basalt field, the southernmost part of the Patagonian plateau lavas (Fig. 7; Ramos and Kay, 1992; Kay *et al.*, 1993; Gorring *et al.*, 1997, 2003; D'Orazio *et al.*, 2000, 2001; Lagabrielle *et al.*, 2004; Espinoza *et al.*, in press). Near-trench magmatic activity close to the Pacific coast on the Taitao peninsula reflects the effects of the subduction of segments of the Chile Rise during the Pliocene (Nelson *et al.*, 1993; Forsythe and Nelson, 1985; Mpodozis *et al.*, 1985; Forsythe *et al.*, 1986; Le Moigne *et al.*, 1996; Lagabrielle *et al.*, 2000, 2004; Guivel *et al.*, 2003).

AVZ (49-55°S)

The AVZ consists of five stratovolcanoes and a small complex of Holocene domes and flows on Cook Island, the southernmost volcanic center in the Andes (Fig. 7; Stern *et al.*, 1976, 1984c; Suárez *et al.*, 1985; Puig *et al.*, 1984; Harambour, 1988; Martinic, 1988; Heusser *et al.*, 1990; Stern and Kilian, 1996). The AVZ was first identified as an independent segment of Andean volcanoes in 1976 (Stern *et al.*, 1976, 1984c), while Cook Island volcano (55°S), the southernmost in the Andean chain, was not discovered until 1978 (Puig *et al.*, 1984; Martinic, 1988), and Reclus volcano (51°S), previously mistakenly identified as Cerro Mano del Diablo, was accurately located only in 1987 (Harambour, 1988). The three northernmost of these volcanoes (Lautaro, Viedma and Aguilera), which are located within or along the flanks of the Southern Patagonian ice cap, have similar chemistry and have been grouped together as the Northern AVZ volcanoes (NAVZ; Stern *et al.*, 1984c; Stern and Kilian, 1996). Cook Island center is located south of the Magallanes Fault Zone (Polonia *et al.*, 1999) and therefore on the Scotia plate (Forsyth, 1975; Giner-Robles *et al.*, 2003).

The southeast extension of the Chile Rise was subducted below this region between 8-14 Ma. This volcanic zone results from the subduction of 12-24 Ma oceanic lithosphere of the Antarctic plate below South America at a velocity of 2 cm/yr. There is no Benioff zone of seismic activity associated with the slow subduction of this young oceanic plate. Crustal thickness below the AVZ is <35 km, and pre-Andean basement rocks consist of Late Paleozoic to Early Mesozoic sedimentary and metamorphic rocks of the Eastern Andean metamorphic complex and Madre de Dios terrane (Mpodozis and Forsythe, 1983; Ramos, 2000; Hervé *et al.*, 2000, 2003; Thompson and Hervé, 2002). The volcanoes of the AVZ erupted onto glacially eroded pre-Andean basement and Mesozoic Andean rocks, and there is no information concerning pre-Holocene activity in this belt. However, peridotite xenoliths from 1-3 Ma Cerro del Fraile alkali basalts, located 25 km east of the AVZ (Fig. 7), are metasomatized by adakitic silicate melts (Kilian and Stern, 2002). This suggests that, following a hiatus in volcanic activity associated with subduction of the Chile Rise below

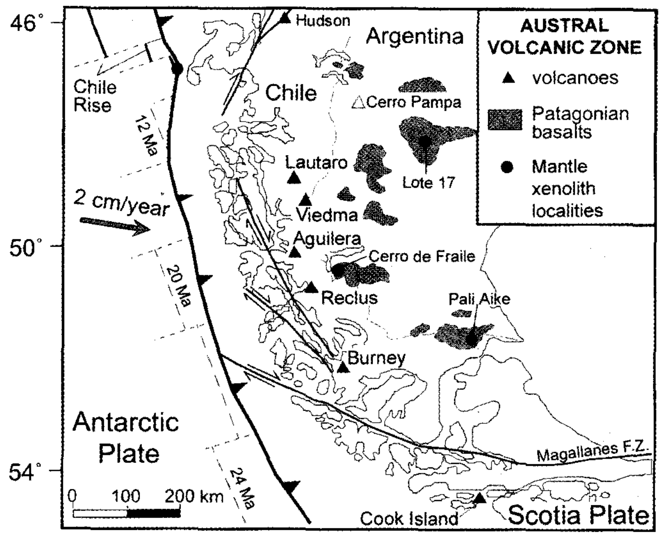

FIG. 7. Map of the Austral Volcanic Zone, modified from Stern and Kilian (1996) and Kilian and Stern (2002), showing the location of the six volcanic centers that make up this volcanic zone, and outcrops of the Patagonian plateau basalts to the east. Details concerning the age and motion of the Antarctic plate can be found in Cande and Leslie (1986), Tebbens and Cande (1997) and Lagabrielle *et al.* (2000, 2004).

this region beginning after 14 Ma (Cande and Leslie, 1986; D'Orazio *et al.*, 2000, 2001; Lagabrielle *et al.*, 2000, 2004), subduction-related magmatic activity may have reestablished itself below the AVZ by the Pliocene.

ANDEAN MAGMA GENESIS

GENERAL

Thorpe (1984) concluded that Andean magmatism was initiated by the dehydration of the subducted oceanic lithosphere resulting in the addition of subducted components into and melting of the overlying mantle wedge. In this sense the initial subcrustal stages of generation of Andean magmas are similar to magma generation processes in oceanic convergent plate boundary island arcs. In fact, tholeiitic and high-Al basalts from the volcanic front of the Central SVZ of the Andes, where the crust is only ~30 km thick, have isotope and trace-element ratios similar to oceanic island arc basalts, and they have not obviously assimilated continental crust (Hickey-Vargas *et al.*, 1984, 1986, 1989).

However, differences in isotopic composition of volcanic rocks from the Northern SVZ and CVZ of the Andes compared with those erupted in the Central SVZ and oceanic island arcs indicate the participation of continental crust and/or subcontinental mantle lithosphere in the formation and evolution of the NSVZ and CVZ Andean magmas. This may occur during interaction of magmas derived in the subarc asthenospheric mantle wedge with continental lithosphere (Rogers and Hawkesworth, 1989; Stern *et al.*, 1990; Stern and Kilian, 1996; Hickey *et al.*, 2002), by intra-crustal assimilation (AFC or MASH processes; James, 1984; Hildreth and Moorbath, 1988; Davidson *et al.*, 1991), and/or by source region contamination of subarc mantle by subducted continental components (Stern *et al.*, 1984b; Stern, 1988, 1991a; Stern and Skewes, 1995; Kay *et al.*, in press). Relative roles of each processes has been debated in each different Andean volcanic zone, as described briefly below.

SVZ

CENTRAL AND SOUTHERN SVZ

In the Central and Southern SVZ, where the crust is relatively thin (<30 km), tholeiitic and high-Al basalts and basaltic andesites are the dominant rocks type erupted from both stratovolcanoes and many minor eruptive centers (López-Escobar *et al.*, 1977, 1993, 1995a; Hickey-Vargas *et al.*, 1984, 1986, 1989; Gerlach *et al.*, 1988; Futa and Stern, 1988), although andesites, dacites and rhyolites do occur. Sr, Nd, Pb and O isotopic data for CSVZ basalts preclude any significant assimilation of continental crust. Detailed studies of CSVZ basalts therefore provide information on the genesis and chemical composition of mafic mantle-derived Andean magmas that have not interacted extensively with continental crust, and thus a 'baseline' for evaluating the extent of crustal interaction in other segments of the arc. These studies have concluded that CSVZ basalts form by melting of the subarc mantle contaminated by fluids derived from the dehydration of the subducted oceanic lithosphere, including sediments, based on: 1) Be isotope data which imply a subducted sediment component in CSVZ basalts (Morris *et al.*, 1990; Sigmarsson *et al.*, 1990; Hickey *et al.*, 2002); 2) excess ^{226}Ra over ^{230}Th and ^{238}U over ^{230}Th which both also imply addition of slab-derived fluids to the mantle magma source (Sigmarsson *et al.*, 2002); 3) Pb isotopic data that suggest Pb is derived from a mixture of mantle and subducted Nazca plate sediment (Barreiro, 1984; Macfarlane, 1999); 4) high ratios, compared to oceanic island basalts (OIB), of large-ion-lithophile-elements (LILE) to rare-earth-elements (REE; for example Ba/La; Fig. 8a), REE to high-field-strength-elements (HFSE; for example La/Nb; Fig. 8b), and very high ratios of LILE to HFSE (for example Ba/Nb; Fig. 8c), which reflect the fact that LILE are more soluble than REE, and REE more soluble than HFSE, and therefore relatively enriched in slab-derived fluids (Hickey-Vargas *et al.*, 1984, 1986, 1989); and 5) Sr (Fig. 8c), Nd, and $\delta^{18}O$ isotopic values indicating only a small difference between CSVZ and OIB basalts, which implies that the mass of components derived from the subducted slab must be small relative to the mass of the mantle source of these basalts (Hickey-Vargas *et al.*, 1984, 1986, 1989; Stern *et al.*, 1990).

Decreasing Ba/La, La/Nb and Ba/Nb (Fig. 8), in magmas erupted progressively east of the volcanic front in the CSVZ suggest decreasing input of slab-derived fluids into the mantle source of volcanoes behind the volcanic front as a result of progressive dehydration of the down-going slab (Futa and Stern, 1988). Decreasing Ba/La and La/Nb across the CSVZ arc are associated with increasing light-REE (La; Fig. 8a) content and light-REE to heavy-REE ratios, both of which are interpreted as a measure of the degree of partial melting of the mantle. Thus, as the input of slab-derived fluids into the subarc mantle decreases, so does the percent of mantle partial melting (Hickey-Vargas *et al.*, 1986, 1989; Muñoz and Stern, 1989; López-Escobar *et al.*, 1995a). Further to the east in the back-arc region, alkali basalts derived by relatively low degrees of partial mantle melting exhibit even lower Ba/La, La/Nb and Ba/Nb, similar to OIB (Fig. 8), and therefore little or no evidence for input of slab-derived components (Skewes and Stern, 1979; Stern *et al.*, 1990; Kay *et al.*, 1993; Gorring *et al.*, 1997; Espinoza *et al.*, in press). Also, samples of the South American continental mantle lithosphere, which occur as peridotite xenoliths in back-arc alkali basalts, present evidence of metasomatism by slab-derived fluids only when their host basalts were erupted in areas of Cenozoic arc volcanism, such as at Cerro del Fraile, Argentina (Fig. 7; Kilian and Stern, 2002). However, xenoliths in alkali basalts erupted further to the east, in regions with no evidence of Cenozoic arc volcanism such as Pali Aike, Chile, and Lota 17, Argentina (Fig. 7), have been metasomatized by fluids without any slab-derived isotopic or trace-element signature (Stern *et al.*, 1986, 1989, 1999; Gorring and Kay, 2001).

Andesites, dacites and rhyolites in CSVZ and SSVZ volcanoes have the same isotopic composition as basalts and basaltic andesites, indicating that they formed either by crystal-liquid fractionation without assimilation (Gerlach *et al.*, 1988), or assimilated young, isotopically similar crust such as Miocene plutonic rocks (McMillian *et al.*, 1989).

NORTHERN SVZ

A large rhyolite pyroclastic flow was erupted in the Pleistocene from the Diamante caldera (Fig. 5; Stern *et al.*, 1984a), but the volcanic rocks forming the three stratovolcanoes in the Northern SVZ,

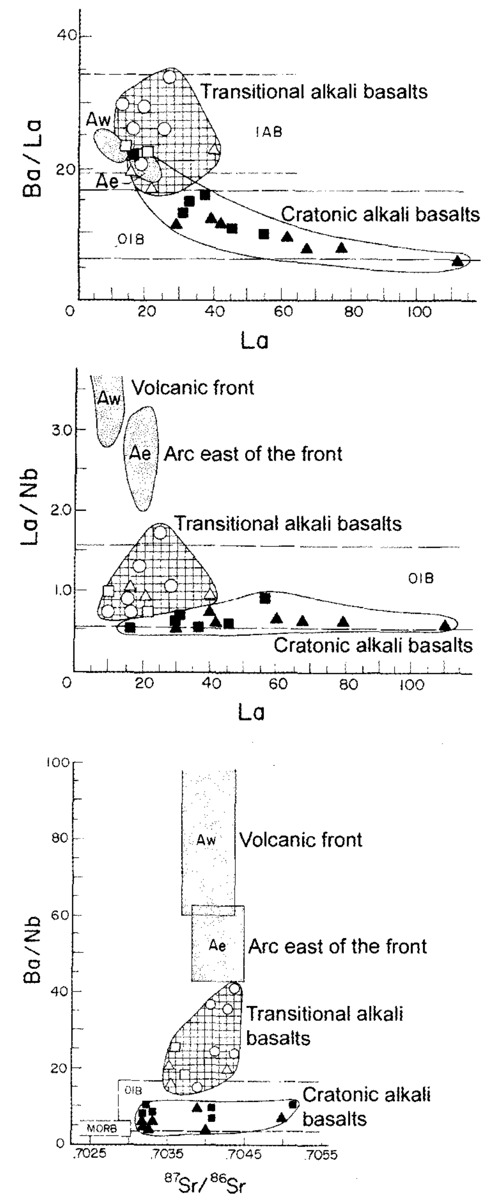

FIG. 8. A. Ba/La *versus* La content; B. La/Nb *versus* La content; and C. Ba/Nb *versus* $^{87}Sr/^{86}Sr$, for samples from along the volcanic front (Aw) of the Central SVZ, CSVZ centers east of the front (Ae), and back-arc alkali basalts in areas of Cenozoic arc volcanism (transitional alkali basalts, open circles and triangles; Stern *et al.*, 1990) and cratonic regions that have not experienced Cenozoic arc volcanism (cratonic alkali basalts; solid squares and triangle; Stern *et al.*, 1990). Figure 8 illustrates the higher Ba/La, La/Nb and Ba/Nb ratios, and La contents, from west to east across the CSVZ arc and into the back-arc region, interpreted to result from decreasing influx of slab-derived fluids into the subarc asthenospheric mantle wedge, and, as a consequence, decreasing degree of partial mantle melting across the arc.

below which the crust is >45 km thick, are dominantly andesites, along with basaltic andesites and dacites. These have higher $^{87}Sr/^{86}Sr$ (Fig. 9) and lower $^{143}Nd/^{144}Nd$ ratios, than either basalts or rocks of similar SiO_2 content in the CSVZ, indicating incorporation of continental crust in the magmas erupted from these three volcanoes (Stern *et al.*, 1984b; López-Escobar *et al.*, 1985; Futa and Stern, 1988; Hildreth and Moorbath, 1988; Stern, 1988, 1989, 1991a). How this crust is incorporated in these rocks is still under debate. Hildreth and Moorbath (1988) outlined a model of Mixing, Assimilation, Storage and Homogenization (MASH) within intra-crustal magma chambers. Stern *et al.* (1984b) and Stern (1988, 1991a), noted that Sr and Nd isotopic ratios were independent of SiO_2 in the range basaltic andesites through dacites, and suggested that isotopic differences between the NSVZ and CSVZ were caused by a northward increase in mantle source region contamination by subducted continental components, due to either 1) a smaller volume of mantle wedge associated with the northward decrease in subduction angle and increase in crustal thickness below the NSVZ; and/or 2) increased subduction erosion caused by the southward migration of the locus of subduction of the Juan Fernández ridge. Stern (1991a) proposed that a small increase of from 1% to 2% subducted components would be sufficient to cause the isotopic differences observed in the NSVZ compared to the CSVZ (Fig. 9), while intra-crustal assimilation of a mass of crust equal to the mass of mantle-derived basalt would be required to generate the same isotopic change. NSVZ volcanic rocks also have higher La/Yb and lower Yb than CSVZ volcanic rocks, suggesting garnet in the source, which could be either in the mantle (López-Escobar *et al.*, 1977; Stern *et al.*, 1984b; Stern, 1991a) or deep crust (Hildreth and Moorbath, 1988).

Stern and Skewes (1995), Nyström *et al.* (2003) and Kay *et al.* (in press) demonstrated that the isotopic difference between the NSVZ and CSVZ developed during the Late Miocene and Pliocene, prior to the migration of the volcanic front to its current position in the Main Cordillera above relatively thick >45 km crust (Fig. 10). At the latitude of the Maipo volcano (34°S), progressive temporal changes between the Early Miocene and Pliocene occurred in the isotopic composition of mantle-derived olivine basalts, implying changes in the

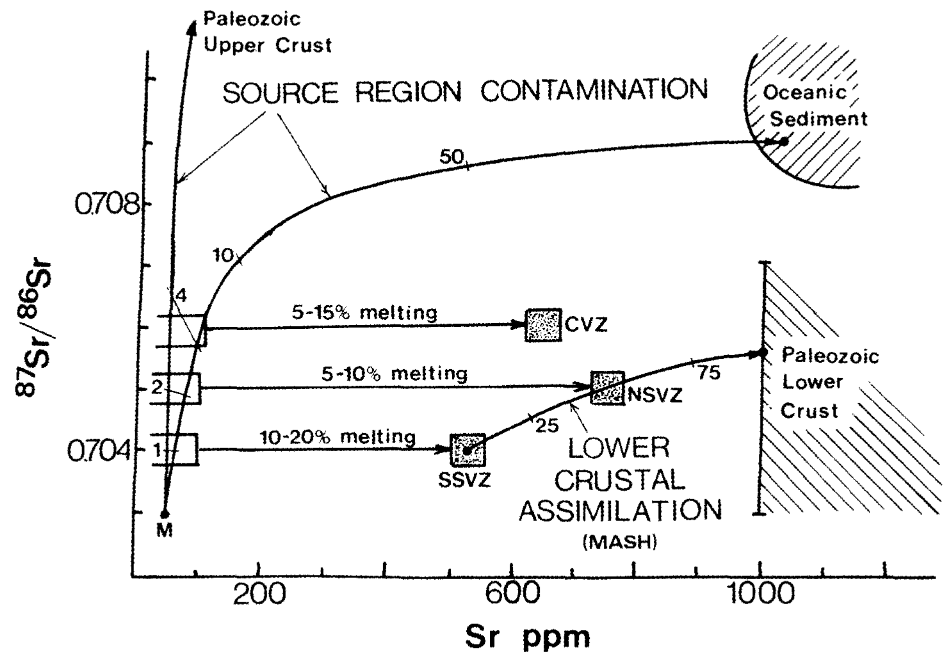

FIG. 9. $^{87}Sr/^{86}Sr$ *versus* Sr content for models of source region contamination compared to crustal assimilation for generation of basalts and basaltic andesites from the Central and Southern SVZ (SSVZ), Northern SVZ (NSVZ) and the CVZ (Stern, 1991a). In order to generate an NSVZ basaltic andesite from a CSVZ basalt requires assimilation of an equal mass of high Sr gabbroic lower crust. Alternatively, an additional 1% of subducted crustal components can produce a mantle with the required higher $^{87}Sr/^{86}Sr$. Increased source region contamination of the subarc mantle below the Northern SVZ may result from either 1) increased subduction erosion of the continental margin due to the subduction of the Juan Fernández ridge, and/or 2) the smaller volume of mantle below the NSVZ due to the lower subduction angle below this region (Stern *et al.*, 1984b; Stern, 1988, 1991a).

isotopic composition of the mantle source region (Stern and Skewes, 1995). Similar progressive changes in isotopic composition occurred between the Early and Late Miocene further north at latitude 32°S, and between the Early Miocene and Pliocene at latitude 33°S (Fig. 10). The diachronous southward migration of these changes closely follows the southward migration of the locus of subduction of the Juan Fernández ridge. This suggests that an important part of these changes resulted from increased contamination by crustal components of the mantle source region below the arc, caused by either increased subduction erosion or decreasing subduction angle and mantle volume in association with ridge subduction. No isotopic changes occurred in conjunction with the 40 km eastward migration of the front in the Pliocene (Fig. 10), and therefore this change is not correlated with crustal thickness.

TRANSITIONAL SVC

Volcanic centers in the Transitional SVZ have erupted magma type ranging from basalts to rhyolites, although andesites and dacites dominate. An ultra-detailed study of the Tatara-San Pedro complex in the TSVZ indicates that open system processes involving mixing newly arrived magma batches with evolved magmas in conduit-reservoir systems played a more significant role than differentiation in large long-lived reservoirs (Davidson *et al.*, 1987, 1988; Ferguson *et al.*, 1992; Feeley and Dungan, 1996; Singer *et al.*, 1997; Feeley *et al.*, 1998; Nelson *et al.*, 1999; Dungan *et al.*, 2001; Costa and Singer, 2002; Costa *et al.*, 2002; Dungan and Davidson, 2004). Dungan *et al.* (2001) suggest that these processes may account for a large part of the trace element variability observed northward along the arc from the CSVZ to

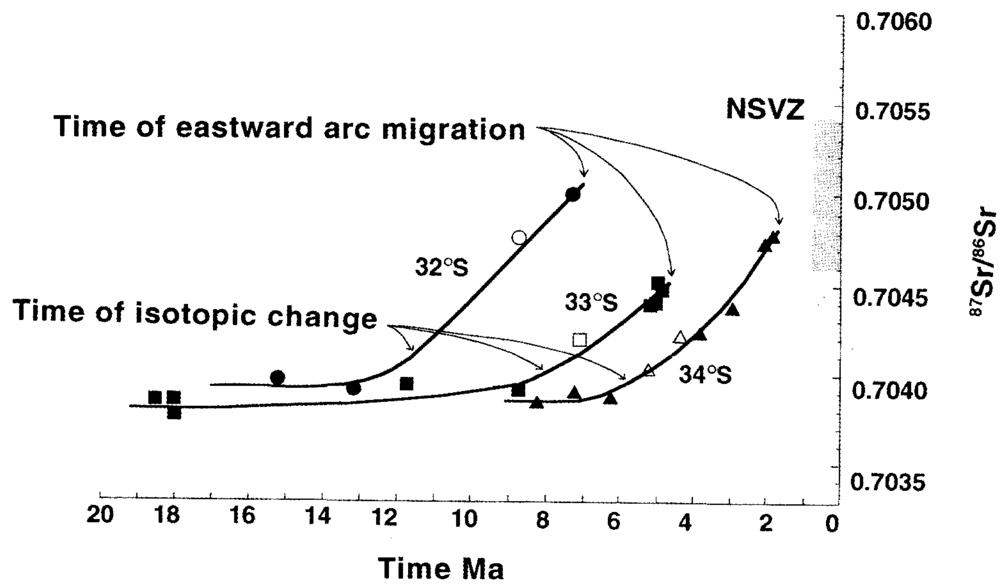

FIG. 10. $^{87}Sr/^{86}Sr$ *versus* time, between the Miocene and Recent, for igneous rocks from central Chile. The figure illustrates the diachronous southward migration of temporal isotopic changes in these magmas, which closely follows the southward migration of the locus of subduction of the Juan Fernández ridge. These temporal changes in isotopic composition have been interpreted as resulting from increased rates of subduction erosion and source region contamination of the subarc mantle below the NSVZ as a result of the ridge subduction (Stern and Skewes, 1995). Note that the 40 km eastward migration of the Northern SVZ arc in the Pliocene to its current locality, overlying the thick crust below the crest of the Main Cordillera between Chile and Argentina (Fig. 3), did not result in any change in the isotopic composition of the most mafic magmas.

the NSVZ. However, the Tatara-San Pedro complex does not exhibit the isotopic variability observed between the CSVZ and NSVZ (Davidson *et al.*, 1988; Dungan and Davidson, 2004), and thus these studies do not address directly the problem of the regional along-strike variations in the quantity of crustal components in SVZ magmas. Tormey *et al.* (1991, 1995) suggest that spatial variations in trace element ratios of basalts northward from the CSVZ through the TSVZ are regional trends caused by a combination of both decreasing degree of mantle partial melting and an increase in the influx of slab-derived fluids as the volume of the subarc mantle decreases northwards.

Large volumes of rhyolites were erupted in the TSVZ from the Pleistocene to Recent Calabozos caldera (Hildreth *et al.*, 1984; Grunder, 1987; Grunder and Mahood, 1988), Puelche volcanic field (Hildreth *et al.*, 1990), Laguna del Maule volcanic center (Frey *et al.*, 1984) and Domuyo volcano (Llambias *et al.*, 1978). Isotopic data indicate that the generation of these rhyolites involved significant contributions from crustal partial melts. As noted by Hildreth *et al.* (1999), these TSVZ centers are all located over the fold-and-thrust belt developed along the eastern edge of the Main Cordillera (Folguera *et al.*, 2002), although this region of the arc is also characterized by the occurrence of intra-arc extensional basins (Muñoz and Stern, 1988).

CVZ

Andesites, dacites and rhyolites are the dominant rock type erupted in the CVZ, although basaltic andesites and occasional basalts also occur. CVZ magmas have elevated $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ (James, 1982, 1984), and lower $^{143}Nd/^{144}Nd$ (Wörner *et al.*, 1988; Davidson *et al.*, 1990) than SVZ magmas, as well as distinctly different Pb isotopic compositions (Barreiro, 1984; Barreiro and Clark, 1984; Wörner *et al.*, 1992b; Aitcheson *et al.*, 1995). These data imply that a greater amount of continental crust has been incorporated in CVZ than in SVZ magmas (Davidson *et al.*, 1991; Francis and Hawkesworth, 1994; Lindsay *et al.*, 2001b; Schmitt *et al.*, 2001, 2002), as might be expected in this region of both extremely thick crust and high rates of subduction erosion.

Correlated variations of Sr, Nd and O isotopes with increasing SiO_2 content (Francis *et al.*, 1984; Feeley and Davidson, 1994), and regional correlations of Pb isotopic compositions and Andean basement age (Wörner *et al.*, 1992b, 1994; Aitcheson *et al.*, 1995), indicate that an important part of the process of crustal contamination of CVZ magmas occurred by intra-crustal assimilation combined with crystallization of these magmas, and/or crustal anatexis (Hawkesworth *et al.*, 1982; Schneider, 1987; Davidson and de Silva, 1995). Geophysical studies interpret crustal structure as having been affected by partial melting (Schmitz *et al.*, 1997). Temporal variations in CVZ magmas indicate that the extent of crustal contamination of CVZ magmas has increased through time, from prior to the Miocene to Recent, as the crust has thickened to >70 km (Rogers and Hawkesworth, 1989; Miller and Harris, 1989; Trumbull *et al.*, 1999; Lucassen *et al.*, 2001). However, arid and hyperarid climate conditions developed in the Andes beginning in the Miocene, and this may have decreased rates of sediment supply to the trench and thus increased the rate of subduction erosion of the continental margin. The rate of subduction erosion of the continental margin in the Central Andes is higher than in either the Northern or Southern Andes. However, the extent to which the base-line composition of magmas generated in the subarc mantle below the CVZ has been affected by source region contamination has not been evaluated because of the lack of basalts in this segment of the Andean arc, which may in itself be an indication of the extent of subarc source region contamination below the CVZ.

NVZ

Andesites and dacites are the dominant rock type erupted in the NVZ, although basaltic andesites and rhyolites also occur. Basalts are relatively scarce. Rocks erupted from volcanoes along the volcanic front in the western Cordillera Oriental of Ecuador and southern Colombia are medium-K calc-alkaline pyroxene basaltic andesites, (± hornblende) andesites, and hornblende (± biotite) dacites (Kilian *et al.*, 1995; Barrágan *et al.*, 1998). In the eastern Cordillera Real, or Central (Colombia), both medium and high-K calc-alkaline basaltic andesites, andesites and dacites occur (Vatin-Pérignon *et al.*, 1990; Calvache and William, 1997; Barrágan *et al.*, 1998). To the east in Ecuador, alkaline to highly alkaline feldspathoid and sodalite bearing tephrites, basanites and phonolites occur (Barrágan *et al.*, 1998).

Despite the variations in crustal age and thickness across the arc, Sr, Nd and Pb isotopic compositions of NVZ volcanic rocks vary very little either regionally or with respect to rock type, and have values similar to basalts erupted in the Central SVZ (James and Murcia, 1984; Kilian *et al.*, 1995; Barrágan *et al.*, 1998). James and Murcia (1984) concluded that small systematic variations in Sr, Nd and O isotopic composition among basaltic andesites, andesites and dacites from Galeras and Nevado del Ruiz volcano, as well as the relatively larger range of whole-rock $\delta^{18}O$ values (6.5-7.8 *per* mil), indicate that formation of these rocks involved as much as 20% intra-crustal assimilation, and that subducted crustal components also must have been incorporated in the subarc mantle-derived primary magmas before they rose into the crust. However, significant differences may occur between measured whole-rock and mineral $\delta^{18}O$ values due to post-crystallization alteration, even in rocks that microscopically appear fresh (Magaritz *et al.*, 1978), and thus the extent of crustal contamination based on whole-rock $\delta^{18}O$ values might be over-estimated. More recent studies at Nevado del Ruiz (Vatin-Pérignon *et al.*, 1990) and Galeras (Calvache and William, 1997), as well as at Chimborazo (Kilian *et al.*, 1995) and Atacazo (Barrágan *et al.*, 1998), the latter two occurring on the mafic accretionary crust of the Cordillera Oriental in Ecuador, conclude that the trace-element and isotopic compositions of these rocks are inconsistent with >10% crustal assimilation of the appropriate crustal lithologies underlying each volcano. Vanek *et al.* (1994), and more recently Barrágan *et al.* (1998), also suggested that across-the-arc differences in trace-element ratios, for rocks that are isotopically similar with each other, are inconsistent with assimilation of the very variable crustal lithologies that underly these volcanoes. They concluded that these trace-element ratio variations result from variations in the extent of slab-dehydration, mantle-source-region contamination and the degree of mantle melting, and that volcanic front magmas formed by a relatively high degree of partial melting (~15%) of the subarc mantle initiated by a large input of slab-derived components, while east of the front, the extent of slab-dehydration and mantle partial melting both decrease.

AVZ

The six volcanic centers in the AVZ have erupted adakitic hornblende andesites and dacites (Stern *et al.*, 1984c; Stern and Kilian, 1996). Neither basalts or basaltic andesites, nor rhyolites, occur in this segment of the Andes. Compared to other Andean arc magmas, Cook Island high-Mg adakitic andesites have very unusual Sr, Nd, Pb and O isotopic compositions, similar to that of MORB, and may have formed by partial melting of young (<24 Ma) obliquely subducted oceanic crust, followed by only very limited interaction with the overlying mantle wedge and continental crust. Adakitic andesites and dacites erupted from the other volcanoes in the AVZ also formed from slab melts which have interacted to a greater extent with the overlying continental mantle lithosphere and crust (Stern *et al.*, 1984c; Stern and Kilian, 1996; Sigmarrson *et al.*, 1998). The extent of this interaction increases progressively towards the north as subduction angle becomes more orthogonal.

EXOTIC MAGMAS

As well as basalts, andesites and rhyolites, other more unusual igneous rocks types occur associated with Andean volcanoes. In the CVZ of northern Chile these include iron-oxide rich (magnetite ± apatite) lava flows at El Laco (Henríquez and Martin, 1978; Nyström and Henríquez, 1994; Henríquez *et al.*, 2003), formed from Fe-rich magmas possibly separated from silicate magmas by liquid immiscibility. Sulfur flows on the flanks of Lastarria and Bayo volcanoes (Naranjo, 1985, 1988) are interpreted to have formed by remobilization of hydrothermal sulfur. Anhydrite-bearing lavas have been erupted at Láscar volcano (Matthews *et al.*, 1994), and anhydrite-rich hypabyssal intrusive rocks of Pliocene age occur at El Teniente (Funk *et al.*, 2004), the world's largest Cu deposit in central Chile. Andean Cu deposits in central Chile also contain extremely large magmatic-hydrothermal breccia pipes, such as the giant Braden diatreme at El Teniente (>1 km diameter at the surface and >3 km known vertical extension; Skewes *et al.*, 2002), and the Cu-rich Sur-Sur (Vargas *et al.*, 1999) and Donoso (Skewes *et al.*, 2003) tourmaline breccia pipes at Rio Blanco-Los Bronces, another giant Cu deposit in Central Chile.

VOLCANIC HAZARD ASSESSMENT AND RISK MANAGEMENT

NVZ

On November 13, 1985, lahars generated by melting of summit snow and ice during a relatively small eruption of the Nevado del Ruiz volcano in Colombia produced the deadliest volcanic event recorded in the Andes, in which >23,000 people were killed (Williams, 1990). The eruption of Nevado del Ruiz made the covers of Time and Newsweek magazines, attracted worldwide attention, and motivated a more systematic effort to evaluate hazards and monitor active volcanoes not only in Colombia, but elsewhere in South America as well. Currently, the Colombian Institute of Geology and Mining -INGEOMINAS- operates three volcanic observatories, one in each of the three segments of the Colombian NVZ. The Observatorio Vulcanológico y Sismológico located in Manizales monitors Nevado del Ruiz and other volcanic centers in the northern segment, an observatory located in Pasto monitors Galeras and other volcanoes in the southern segment, and one located in Popayán monitors volcanoes in the central segment (Fig. 2). In part as a consequence of the increased interest in Andean volcanism generated by the Nevado del Ruiz eruption, the Galeras volcano, one of the historically most active volcanoes in Colombia, which had begun a new cycle of activity in 1988, was selected as a Decade Volcano for the United Nations International Decade for Natural Disaster Reduction, the only volcano so selected in South America. The Galeras volcano erupted on January 14, 1993, while volcanologists attending a Decade Volcano workshop in Pasto were on a field trip to the volcanoes crater. This produced far fewer deaths than the eruption of Nevado del Ruiz, but six of those killed were volcanologists (Baxter and Gresham, 1997). Two special volumes of the Journal of Volcanologic and Geothermal Research presented the results of intense interdisciplinary research conducted at the Nevado del Ruiz and Galeras volcanoes (Williams, 1990; Stix *et al.*, 1997), and provide a model for volcanologic studies and hazard assessment, as well as information on risk management.

In Ecuador, 30 volcanic centers are considered dangerous (Alvarado *et al.*, 1999). Hazards involve pyroclastic flows, lahars and tephra falls. A rhyolite explosive eruption of Cotopaxi at 4,500 BP caused melting of the summit glacier, generating the largest known lahar in the NVZ, with a volume of 3.8 km^3, similar to that of the Osceola mudflow generated by Mt Rainier (Mothes *et al.*, 1998). This Chilles Valley lahar flowed north through the inter-Andean valley, an area currently inhabited by >3 million people, ultimately reaching the Pacific Ocean 326 km to the northwest of the volcano. Cotopaxi has had >30 historic eruptions and Barberi *et al.* (1995) calculate that explosive eruptions of Cotopaxi have produced lahars once every 117 years over the last 2,000 years, the last one 127 years ago in 1877! Prehistoric lahars from Guagua Pichincha have reached the area of the capital city of Quito, located at the foot of this volcano, and the 1660 eruption of this volcano deposited >30 cm of ash in Quito. These are 2 of 12 volcanoes with permanent seismic monitoring stations installed by the Instituto Geofísico de la Escuela Politécnica Nacional in Quito, which maintains a website www.epn.edu.ec providing updated information on volcanic activity in Ecuador. The November 2002 eruption of the Reventador volcano, which occurred without any precursor activity prior to entering into eruption, is described by Hall *et al.* (2004).

CVZ

The largest historic explosive eruption in the Andes was generated by the Huaynaputina volcano, located just 75 km northeast of the city of Arequipa in southern Perú (Fig. 3), between February 19 and March 6, 1600 (de Silva and Zielinski, 1998; Thouret *et al.*, 1999, 2002; Adams *et al.*, 2001). This VEI = 6 magnitude explosive event produced 10.2-13.1 km^3 bulk volume, or 6-8.8 km^3 dense rock equivalent, of dacitic tephra and pumice. Tephra falls distributed to the northwest, and pyroclastic flows, killed a relatively large number (~1500) of people given the population in the area at that time compared to today's, buried seven villages, and dammed the Tambo canyon, creating two temporary lakes that eventually generated debris flows that swept down the 120 km long valley to the Pacific ocean. Arequipa, currently with a population >900,000, is located just 17 km southeast of the summit of Mitsi volcano (5,820 m; Thouret *et al.*, 2001), which has produced 20 tephra falls as a result of Vulcanian and sub-Plinian eruptions in the last 50,000 years, the last in 1,455 ± 15 AD. Although the other volcanoes in the CVZ of Perú are located further from major population centers, the Instituto Geológico Minero y Metalúrgico (INGEMMET) directs a program of geologic risks, which includes the goal of compiling an inventory and mapping the volcanoes of Perú. Remote sensing studies also provide a basis for evaluating hazards of Peruvian volcanoes (de Silva and Francis, 1990, 1991).

The CVZ volcanoes of northern Chile, southwest Bolivia and northwest Argentina, like those in Perú, also occur relatively remote from large population centers, and they are not continuously monitored. Láscar volcano in northern Chile, recently the most active volcano in the CVZ, entered into a cycle of activity beginning in 1984 (Matthews *et al.*, 1997; Gardeweg *et al.*, 1998), with strong eruptions in 1986, 1990, 1993, and again in 2000. This activity generated significant SO_2 emissions, as well as ash plumes that produced tephra fall far to the southeast in Argentina. Fieldwork and remote sensing studies (de Silva and Francis, 1991) have documented a number of large debris flows from Chilean CVZ volcanoes, such as Parinacota (Francis and Wells, 1988; Clavero *et al.*, 2002), Taapaca (Clavero *et al.*, 2004), Socompa (Francis *et al.*, 1985; van Wyk de Vries *et al.*, 2001), Lastarria (Naranjo and Francis, 1987), and Llullaillaco (Richards and Villeneuve, 2004). At least six large, long-lived caldera systems and ignimbrite centers with documented Quaternary activity have been identified in the southern CVZ (de Silva, 1989a,b; de Silva and Francis, 1991). These represent the extrusive portions of batholith-size magma chambers (Chmeilowski *et al.*, 1999; Haberlan and Rietbrock, 2001). Their potential for future eruptions of similar magnitude is uncertain.

SVZ

The SVZ occurs at the latitudes where >70% of the population of Chile live, and contains some of the most active volcanoes in the Andes. In 1932, Quizapu volcano erupted >9.5 km^3 of tephra, the largest explosive eruption of any Andean volcano in the 20th century (Hildreth and Drake, 1992), and it erupted an approximately equal volume of magma in 1846. In 1991, the Hudson volcano, the southernmost in the SVZ (Fig. 4), erupted 4-7 km^3 of tephra, the second largest eruption in the Andes in the XX century (Naranjo *et al.*, 1993b; Scasso *et al.*, 1994). Hudson was only first recognized as a volcano in 1971 (Fuenzalida and Espinoza, 1974), after a minor eruption melted the glacier inside the 9 km wide caldera, producing a catastrophic lahar which flowed from the volcano 40 km east to the coast (Best, 1992). Tephrochronology studies indicate that in 6,720 BP Hudson produced the largest Holocene eruption in the southern Andes (>18 km^3 of tephra; Fig. 11; Stern, 1991b; Naranjo and Stern, 1998), as well as another very large explosive eruption in 3,600 BP. The andesite tephra ejected by the 1991 eruption of Hudson covered more than 150,000 km^2 in Chile and Argentina, and resulted in unconfirmed reports of the death of over 500,000 sheep. A much smaller eruption of the Lonquimay volcano in 1988, which produced only 0.12 km^3 of andesite tephra and lava during a 13 month eruption, resulted in the death of 14,000 animals due to the high fluorine content of the gas generated by the eruption (Moreno and Gardeweg, 1989).

Numerous other SVZ volcanoes have produced Holocene explosive activity of significant magnitude, including Planchón (Naranjo and Haller, 1993), Villarrica (Clavero and Moreno, 1994), Michimahuida and Melimoyu (Naranjo and Stern, 2004). Some, like the Hudson, have generated calderas during their evolution, including Sollipulli (Gilbert *et al.*, 1996; Naranjo *et al.*, 1993a), Llaima (Naranjo and Moreno, 1991), Villarrica (Clavero and Moreno, 1994) and Puyehue (Gerlach *et al.*, 1988). Naranjo *et al.* (2001) and Naranjo and Stern (1998, 2004), have documented a total 62 small, medium and large explosive Holocene eruptions from 48 SVZ volcanoes.

Perhaps more significant are the three giant calderas Diamante (Fig. 5; Stern *et al.*, 1984a), Calabozos (Hildreth *et al.*, 1984) and Coviahue (Fig. 6; Muñoz and Stern, 1988), which have produced large volume pyroclastic flows in the last 1.1 Ma. The 450 ± 60 ka eruption of the Diamante caldera generated the 450 km^3 Pudahuel rhyolite tuff, which flowed through and covered to a depth of 30 meters the northern part of the central valley in Chile, including a large part of the area now occupied by Santiago (Fig. 5; Stern *et al.*, 1984a).

Villarrica and Llaima volcanos have had over 50 reported episodes each of activity between 1640 and 1999 (Fuentealba, 1984; Naranjo and Moreno, 1991; Clavero and Moreno, 1994; Moreno *et al.*, 1994a,b; Petit-Breuilh, 1994; Petit-Breuilh and Lobato, 1994), and Villarrica has maintained an active summit lava lake since its last eruption in 1985 (Witter *et al.*, 2004; Witter and Delmelle, 2004; Calder *et al.*, 2004). In 1996 the Servicio Nacional de Geología y Minería (SERNAGEOMIN) created the Observatorio Volcanológico de Andes del Sur (OVDAS), located in Temuco, to monitor these two volcanoes. Other SVZ volcanoes that are currently monitored include Lonquimay, Choshuenco, Osorno and Calbuco, and hazard maps have been completed for Planchón-Peteroa (Naranjo *et al.*, 1999), Villarrica, Osorno, Calbuco, Copahue, Callaqui, Tolguaca and Lonquimay. OVDAS provides updated information on these and other volcanoes in Chile at the www.sernageomin.cl website.

The three highest volcanic complexes in the

FIG. 11. Photograph of a 20 cm thick tephra layer in Tierra del Fuego (54°S) derived from the 6,720 BP eruption of the Southern SVZ Hudson volcano, which was the largest (>18 km^3) Holocene eruption in the southern Andes (46°S; Fig, 7; Stern, 1991b; Naranjo and Stern, 1998). This olive green andesitic tephra layer, which was deposited >1,000 km south of its source volcano, has been affected by soft sediment deformation as well as a small thrust fault.

SVZ, the snow and ice covered Tupungato-Tupungatito, San Jose-Marmolejo and Maipo volcanoes, located at the northern end of the SVZ, all drain into tributaries of the Maipo River, which runs through the southern edge of Santiago, home to >6,000,000 people. However, these centers are not continuously monitored.

AVZ

The volcanoes of the AVZ are remote from any large population center and these volcanoes are not monitored for activity. The only certain historic eruption in the AVZ was the eruption of the Lautaro volcano in 1959-60 (Martinic, 1988). Tephrochronology studies indicate only 4 large explosive eruptions during the Holocene (Stern, 1990, 1992, 2002; Kilian *et al.*, 2003). The thickest Holocene tephra fall deposits in Magallanes and Tierra del Fuego were those generated by the 6720 BP eruption of Hudson in the SVZ (Fig. 11; Stern, 1991b; Naranjo and Stern, 1998).

CONCLUSIONS

Active Andean volcanoes are the most recent manifestation of over 185 million years of magmatic activity related to oceanic-continent collision along the west coast of South America. Subduction geometry, which is affected by subduction of buoyant oceanic plateaus and ridges, controls the presence or absence of volcanic activity in different segments, and geochemical studies confirm the close genetic link between the subduction process and magma genesis below the Andes. Other factors related to subduction, such as the age of the oceanic plate being subducted, the degree of obliquity of subduction direction relative to the continental margin, and the extent of sedimentary infill of the trench, which affects the rates of subduction erosion along the continental margin, all have an important impact on the details of processes of magma generation and the distribution of volcanic centers in each of the four volcanically active zones of the Andes. Pre-Andean basement ages, Andean structural evolution, crustal thickness and Neogene tectonic activity also influence the genesis and fine structure of the Andean volcanic arc.

What is most clear from the many detailed studies of individual Andean volcanoes published in the last 30 years is that andesites are not the only volcanic rock type erupted in the Andes, and that there is no unique 'Andean-type' volcanic arc. The diversity of Andean volcanic rocks reflects a diversity of magma genesis processes that need to be evaluated in the context of both the along and across-strike variations in the geologic and tectonic characteristics of the continental South American plate, the oceanic plates it is colliding with, and the details of the collision process. To the extent that volcanic hazards reflect differences in eruptive styles and types of magmas erupted from different Andean volcanoes, and not just the quantity of snow and ice present on a volcano, a diversity of approaches and potential outcomes need to be considered in the process of volcanic hazard assessment and risk management. There is still plenty of work to be done.

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