

Pre-Cenozoic intra-plate magmatism along the Central Andes (17–34°S): Composition of the mantle at an active margin

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Abstract

Sr–Nd–Pb isotope ratios of alkaline mafic intra-plate magmatism constrain the isotopic compositions of the lithospheric mantle along what is now the eastern foreland or back arc of the Cenozoic Central Andes (17–34°S). Most small-volume basanite volcanic rocks and alkaline intrusive rocks of Cretaceous (and rare Miocene) age were derived from a depleted lithospheric mantle source with rather uniform initial $^{143}\text{Nd}/^{144}\text{Nd}$ (~ 0.5127 – 0.5128) and $^{87}\text{Sr}/^{86}\text{Sr}$ (~ 0.7032 – 0.7040). The initial $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are variable (18.5–19.7) at uniform $^{207}\text{Pb}/^{204}\text{Pb}$ ratios (15.60 ± 0.05). A variety of the Cretaceous depleted mantle source of the magmatic rocks shows elevated Sr isotope ratios up to 0.707 at constant high Nd isotope ratios. The variable Sr and Pb isotope ratios are probably due to radiogenic growth in a metasomatized lithospheric mantle, which represents the former sub-arc mantle beneath the early Palaeozoic active continental margin. Sr–Nd–Pb isotope signatures of a second mantle type reflected in the composition of Cretaceous (one late Palaeozoic age) intra-plate magmatic rocks ($^{143}\text{Nd}/^{144}\text{Nd} \sim 0.5123$, $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.704$, $^{206}\text{Pb}/^{204}\text{Pb} \sim 17.5$ – 18.5 , and $^{207}\text{Pb}/^{204}\text{Pb} \sim 15.45$ – 15.50) are similar to the isotopic composition of old sub-continental lithospheric mantle of the Brazilian Shield.

Published Nd and Sr isotopic compositions of Mesozoic to Cenozoic arc-related magmatic rocks (18–40°S) represent the composition of the convective sub-arc mantle in the Central Andes and are similar to those of the Cretaceous (and rare Miocene) intra-plate magmatic rocks. The dominant convective and lithospheric mantle type beneath this old continental margin is depleted mantle, which is compositionally different from average MORB-type depleted mantle. The old sub-continental lithospheric mantle did not contribute to Mesozoic to Cenozoic arc magmatism.

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1. Introduction

The western margin of South America (Fig. 1) was episodically active at least since latest Proterozoic/Early Palaeozoic (e.g., Coira et al., 1982; Ramos, 2000). This activity reworked Proterozoic lithosphere and formed

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the Phanerozoic western margin of the Proterozoic cratonic South America (Fig. 1; e.g., Damm et al., 1990; Pankhurst and Rapela, 1998; Cordani et al., 2000; Lucassen et al., 2001, 2004; Schwartz and Gromet, 2004). Orogens with thickened crust formed in Cenozoic and Early Palaeozoic times at this margin. Both orogens evolved during ongoing subduction and large volumes of hybrid magmatic rocks formed. These rocks represent mixtures of variable amounts of melts derived from sub-arc mantle sources and Proterozoic to Palaeozoic mainly felsic crust. The composition of the Early Palaeozoic (e.g., Damm et al., 1990; Pankhurst and Rapela, 1998; Coira et al., 1999; Pankhurst et al., 2000; Kleine et al., 2004) and Cenozoic to Recent magmatism (e.g., Harmon et al., 1984; Francis et al., 1989; Kay et al., 1994; Trumbull et al., 1999; Lindsay et al., 2001) and the Proterozoic to Palaeozoic crust (e.g., Damm et al., 1990; Pankhurst and Rapela, 1998; Lucassen et al., 2001, 2004) has been a subject of numerous studies. The composition of the sub-arc mantle source is, however, difficult to constrain. Uncontaminated mantle melts are rare in continental arcs, especially in compressional tectonic regimes with a thickened crust as in the modern Central Andes. The sub-arc mantle is the convective, asthenospheric mantle forming the mantle wedge, but substantial contributions from lithospheric mantle to the sub-arc mantle are likely during up to 200 km arc-normal shortening of the lithosphere since Cenozoic in the Central Andes (CVZ in Fig. 1b; e.g., Hindle et al., 2005 and references therein). This lithospheric mantle could comprise (i) Early Palaeozoic mantle, which formed or was modified from older mantle during the Early Palaeozoic orogeny at an active margin (references above), (ii) Proterozoic sub-continental lithospheric mantle (SCLM) belonging to the Brazilian shield, which is thought to be progressively underthrust beneath the Present Central Andes during shortening (e.g., McQuarrie et al., 2005 and references therein), and (iii) modifications of the mantle by magmatic activity during Mesozoic extension (e.g., Viramonte et al., 1999; Sempere et al., 2002). We investigated major and trace element and radiogenic isotope (Sr, Nd, and Pb) compositions of mafic, intra-plate magmatic rocks from a transitional area between the Phanerozoic western margin of South America and the Proterozoic Brazilian Shield (Fig. 1). The new data give insights into the composition of the lithospheric mantle ~17°–34°S and 64°–66°W in the late Mesozoic, before the onset of the Cenozoic shortening in the Central Andes. Further, Miocene alkaline rocks from two locations at the eastern side of the Andes (Fig. 1b) give information on the mantle after the onset of the Cenozoic orogeny. The

isotopic signatures of lithospheric mantle are discussed in the framework of published data from the Jurassic to Recent sub-arc mantle, mantle xenoliths representing late Cretaceous lithospheric mantle, and the composition of SCLM of the Brazilian shield.

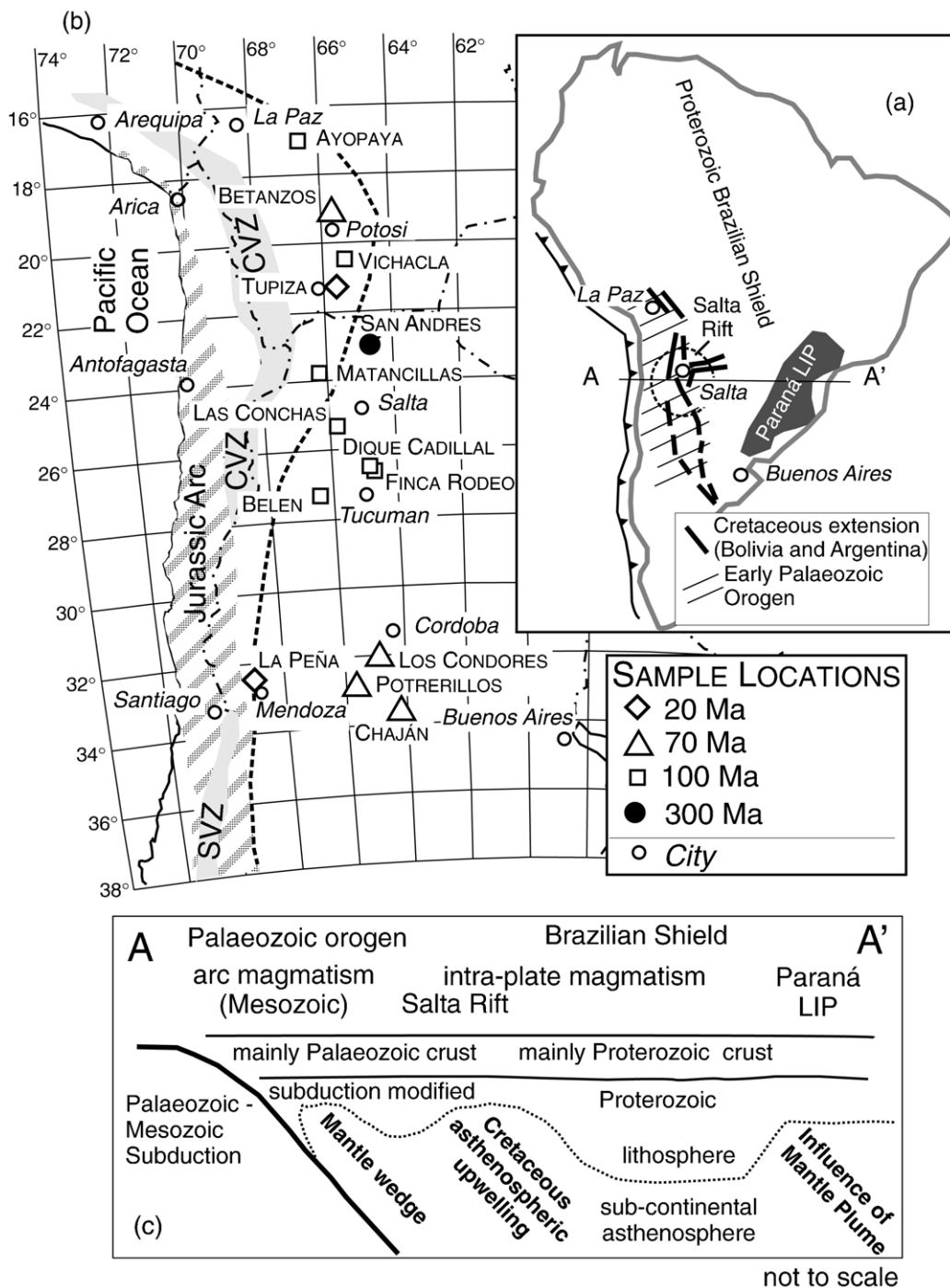
2. Geodynamic setting and geological framework of the intra-plate magmatism

Only magmatic rocks, which represent primary, uncontaminated, magmas, are good candidates for constraining the composition of their mantle source. Such magmas may be generated in arc, back arc and intra-plate settings (Fig. 1c). The South American continent experienced two geodynamic settings, the Early Palaeozoic and late Palaeozoic to Recent subduction in the west (Bahlburg and Hervé, 1997; above references) and Mesozoic continental break-up with the opening of the South Atlantic in the east (e.g., Hawkesworth et al., 1992; Storey, 1995). Mafic intra-plate magmatic rocks, which are the focus of this study (17°–34°S; Fig. 1), occupy an intermediate geographic position between the region influenced by the magmatic arcs in the west and plate separation in the east. The Jurassic and early Cretaceous magmatic arcs produced large volumes of juvenile magmatic rocks from the sub-arc mantle in (e.g., Rogers and Hawkesworth, 1989; Kramer et al., 2005; Lucassen et al., 2006). The composition of the Brazilian SCLM east of the study area is constrained by Cretaceous small-volume, alkaline, intra-plate magmatic rocks (e.g., Carlson et al., 1996; Gibson et al., 1996, 1999; Comin-Chiaromonte et al., 1997, 2005; Velázquez et al., 2006), which occur between the western borders of the early Cretaceous Paraná flood basalt province (e.g., Hawkesworth et al., 1992, 2000) and western Paraguay. The Paraná flood basalt province has been ascribed to mantle plume activity whereas the small-volume intra-plate magmatism was mainly triggered by recurrent lithospheric extension (above references).

The tectonic regime in the study area and much of the South American continent was dominated by extension to transpression from latest Palaeozoic to Early Tertiary (e.g., Ramos, 2000; Sempere et al., 2002). Extension to transpression also dominated the tectonics in the Mesozoic magmatic arcs during oblique subduction (e.g., Scheuber et al., 1994). The most prominent surface expression of extension in the study area is the Salta basin (e.g., Salfity and Marquillas, 1994; Marquillas et al., 2005) of the ‘Salta rift’ in northwest Argentina (Fig. 1a Galliski and Viramonte, 1988; Griens et al., 1991) with a maximum of ~5 km of continental sedimentation in depocentres from Early Cretaceous until Early Tertiary. Crustal thinning was not prominent

in general and, where it occurred, was compensated by continental sedimentation. A Cretaceous crustal thickness in the range of 30 to 40 km is indicated by the Moho depth from seismic data in undeformed crust east of the Cenozoic orogen (e.g., Yuan et al., 2002; Fromm et al., 2004). A crustal thickness of ~35 km has been

inferred in one of the sub-basins of the ‘Salta rift’ on the basis of thermo-barometry and isotopic dating on lower crust felsic and mafic granulite xenoliths in basanites (Las Conchas location; Fig. 1b; ages ~90–110 Ma; $P=1.0$ GPa; 900 °C; Lucassen et al., 1999). Spinel–ilherzolite xenoliths from the same location indicate high



temperatures of Cretaceous equilibration in the upper most mantle (900–1200 °C, $P < 1.4$ –1.7 GPa; Lucassen et al., 2005). The xenolith data indicate an unusually hot lithosphere. The thermal anomaly seems to be triggered by heat transfer within the mantle and the underlying process resulted in no or moderate thinning of the crust. As a consequence, the make-up of the lithosphere in the Cretaceous was still dominated by early Palaeozoic or Proterozoic rocks (Fig. 1).

The alkaline magmatism from the late Palaeozoic onwards (Fig. 1b) is of small volume (Viramonte et al., 1999) and most sub-volcanic rocks occur as dikes or sills on the meter to 10 m scale or small alkaline intrusions of a few 100 m (alkaline intrusions: Ayopaya, Vichacla, San Andres, La Peña, Fig. 1b). The intrusion level is in the uppermost crust. The small size of the occurrences hampers a systematic search for outcrops on satellite or aerial views and our study relies mainly on already described locations (Table 1). Key information on sample locations and references are summarized in Table 1; the approximate positions of the sample locations are in Fig. 1b.

3. Results

Sample preparation, analytical methods and data quality are described in the Appendix.

3.1. Age

Reliable isotopic dating is a major problem in the fine grained basic to ultrabasic rocks. Suitable minerals are commonly lacking and whole-rock dating could be biased by undetected alteration. We used the K–Ar method for dating of phlogopite or amphibole separates and, where hydrous phases are absent, whole-rock samples. New and published isotopic ages are summarized in Table 1. Details of the new K–Ar ages are in Table A1 of the electronic Appendix. The effects of regional or contact metamorphism are absent and all dikes and high-level intrusions have simple cooling

histories. A complex thermal evolution of the minerals can be excluded in fast cooled upper-crustal intrusions. Four age groups are discerned on the basis of new and published K–Ar ages (Fig. 1b). The majority of samples show ages between ~130 and 100 Ma (group referred as 100 Ma), or latest Cretaceous ages between ~80 and 70 Ma (group referred as 70 Ma). A late Palaeozoic alnöite (~300 Ma) is unique in our sample set, but is included here as such ages are common in extension-related magmatic rocks of Peru (compilation in Sempere et al., 2002). The late Palaeozoic and the Mesozoic ages are contemporaneous with magmatic activity in the arc. Alkaline volcanic rocks from the Tupiza area and the La Peña alkaline intrusion are Miocene in age (~20 Ma; Table 1) and fall within the age range of magmatic rocks from the active arc.

3.2. Geochemical characteristics of the magmatic rocks

3.2.1. Sample selection

The aim of this study is a characterization of mantle sources by the isotopic composition of mafic intra-plate magmatic rocks. Therefore, samples with obvious crustal contamination should be excluded beforehand. Further, effects of severe alteration, commonly caused by weathering, on the chemical composition cannot be completely removed by sample treatment and require sample selection. The loss on ignition (LOI, Tables 2a, and A2 in the Appendix) is used as a guide to the intensity of alteration. LOI is <2–3 wt.% in most samples, and traces of alteration in these rocks are absent or minor under the microscope. The samples with a LOI > 3 wt.% show moderate to considerable alteration of the primary minerals, including mainly serpentinization or rare carbonatization of olivine, hydration–oxidation of opaque and Fe–Mg minerals, and sericitization of feldspars. We restrict the discussion of mantle sources to samples with (i) LOI < 2.5 wt.% or Sr contents > ~900 ppm and (ii) MgO > 6 wt.%. Sr content and hence Sr isotope composition is sensitive to feldspar

Fig. 1. Major geological provinces of the South American Mesozoic to Recent lithosphere and sample locations. (a) Large scale overview of the tectonic situation; the investigated rocks come from an area transitional between the Mesozoic to Recent subduction zones in the west (already active in the early and late Palaeozoic) and the early Cretaceous Paraná large igneous province (LIP) (related to the opening of the South Atlantic and possible plume activity; Renne et al., 1993, 1996). A–A' indicates the position of the synoptic profile. (b) Sample locations of intra-plate magmatic rocks in the context of the distribution of Jurassic to lower Cretaceous arc-related rocks, the presently active Central Volcanic Zone (CVZ), and Southern Volcanic Zone (SVZ). In the North the sample locations are situated within the Cenozoic orogen (eastern extension is indicated by the broken line), in the South mostly east of the orogen. (c) Schematic W–E cross section through the South American lithosphere in the Cretaceous and relative spatial relations between different mantle and crustal domains. Late Mesozoic to Tertiary intra-plate magmatism occurs at various locations west of the Paraná LIP (e.g., Comin-Chiaramonti et al., in press). The Cretaceous Salta Rift marks the transition between the area influenced by the early Palaeozoic orogeny and arc and Proterozoic Brazilian Shield. Sources: (a) is modified from Schultz (2004); (c) is redrawn and modified from Schobbenhaus and Bellizzia (2001).

Table 1

Sample locations, key features of the samples, new and published isotopic ages and references to the local geology

Location [W longitude S latitude]	Mode of occurrence	Petrography	Isotopic age K–Ar data [Ma]	References
Ayopaya (1) 66°45', 17°00'	Alkaline intrusions, sub-volcanic rocks and dikes in Ordovician sedimentary rocks; map in ref. (1)	Lamprophyre, nepheline-syenite, calcio- and magnesio-carbonatite, and kimberlite	On biotite (1; three ages) ~100 on biotite (2) 98±3	(1) Schultz et al. (2004) (2) Kennan et al. (1995)
Betanzos Arollo Chorillos 65°25.69', 19°33.25'	Various dikes (<2 m, extension ~10–20 m) in Ordovician sedimentary rocks	Occasionally biotite bearing basanites; small lherzolite xenoliths	On biotite 82.0±3.3 81.5±3.0	This work
Vichacla (3) 65°33.02', 20°38.33'	Small alkaline complex (diameter ~100 m); coarse grained sub-volcanic rocks and dikes in Ordovician sedimentary rocks	Rocks of variable composition: pyroxene ≤40 vol% hornblende ≤40, biotite ≤20, plagioclase+nepheline ≤30; titanite, apatite ≤2 fresh material basalt dikes	On hornblende (1) 114.3±2.8 on biotite (1) 118.1±3.2; on biotite (2) 120±0.5	(1) This work (2) Tawackoli et al. (1999) (3) Santivañez et al. (1996)
NE Tupiza 65°43.17', 21°24.64'	Dacitic plug with autobreccia (diameter ~50 m) in Ordovician sedimentary rocks	Trachyandesite–dacite; fresh material	On hornblende 19.8±0.8 on biotite 21.2±0.8	This work
S of Tupiza; Road Tupiza–Villazon 65°41.57', 21°31.10'	Basaltic flows <10 m and dikes in Ordovician sedimentary rocks	Sub-alkaline basalt–trachyandesite, variable degrees of weathering, small altered lherzolite xenoliths	On biotite 19.2±0.8	This work
San Andres–Rio Seco/Oran 64°46.79', 23°06.19' nearby Los Alisos/Oran 64°46'; 23°02' (1, 2)	Boulders from the dry valley; sheet like small intrusion into Palaeozoic sedimentary rocks (Ordovician?)	Homogeneous alnöite (1, 2, 4) small scatter in mineral modes: olivine, pyroxene, biotite, melilite, titanite, some primary carbonate; fresh material	On biotite (3), Los Alisos 303±10; on biotite (4), San Andres 304±7; 300±7; 310±6; 304±10	(1) Barbieri et al. (1997) (2) Villar et al. (1997) (3) Mendez and Villar 1979) (4) this work
Matancillas near S.A. d.l.Cobres; 66°19.81', 23°55.06'	Dike ~3 m × 50 m in Ordovician sedimentary rocks	Biotite-rich phono-tephrite fresh material	On biotite 109±3 108±2	This work
Las Conchas (1, 2) Amblayo (north of Las Conchas) 65°50.37', 25°26.53'	Numerous stratigraphically controlled dikes and small sills in Cretaceous rift-related sedimentary rocks; map in ref. (2)	Basanite, lherzolite xenoliths, lower crust xenoliths (Sm–Nd age ca.100 Ma (3) fresh material		(1) Viramonte et al. (1999) (2) Lucassen et al. (2002a) (3) Lucassen et al. (1999)
Dique Cadillal near Tucuman, 65°11.25', 26°37.01'	Trachyte intrusion and basanite dike in Palaeozoic sedimentary rocks	Trachyte, basanite; basanite contains small lherzolite xenoliths; fresh material	On whole rock (1) 100–130	(1) Bossi and Wampler (1969)
Finca del Rodeo 64°57.48', 26°39.34'	Dikes in weakly metamorphosed Palaeozoic basement	Basanite, some alteration of olivine; contains small lherzolite xenoliths	On whole-rock 95±3 (96±3 repeat) 98±3 (99±3 repeat)	This work
Belén, Puerta de San José (1) 66°59.52', 27°32.76'	Pyroclastic flows over Cretaceous (?) Redbeds and Ordovician granite; ~30 m basanite flows	Basanite, variable degree of alteration of olivine; small lherzolite xenoliths	On whole rock ~130 Ma	(1) Rossello et al. (1999)
Los Condores (1, 2) 64°30', 32°10'– 32°20'	Various cogenetic centres on/in Palaeozoic sedimentary rocks map in ref. (2)	Basalt–trachy-basalt	Whole rock (cited by 1) 115–130	(1) Lagorio (1998) (2) Lucassen et al. (2002a)
Chaján (1, 2) 64°00', 33°40'	Lava- and intercalated pyroclastic flows in/on Palaeozoic sedimentary and granitoid rocks map in ref. (2)	Basanite; fresh material; lherzolite xenoliths with garnet pseudomorphs (1)	On biotite (2) ~67 Ma	(1) Escayola et al. (1998) (2) Lucassen et al. (2002a)

Table 1 (continued)

Location [W longitude S latitude]	Mode of occurrence	Petrography	Isotopic age K–Ar data [Ma]	References
La Peña (2, 3) 69°00', 32°42'	Small alkaline intrusive complex in Palaeozoic metasedimentary rocks; maps in (2, 3)	Malignite (nepheline + plagioclase ≤ 60 vol%–garnet ≤ 20 , hornblende + pyroxene in variable proportions ≤ 20 , titanite, opaque minerals ≤ 5); clinopyroxenite (with biotite, hornblende), nepheline-syenite fresh material	On biotite from pyroxenite (1) 18.8 ± 0.5 on amphibole (1) 17.9 ± 0.5 18.5 ± 0.5 , 19.5 ± 0.6 on amphibole (Ar–Ar) from malignite (3) 18.7 ± 0.5	(1) This work (2) Villar and Zappettini (2000) (3) Zappettini et al. (2005)
Potrerrillos 65° 50', 32° 50'	Small plug intruding granitoid batholith (2)	Transition trachy-basalt–basanite; fresh material	On whole rock (1) 78.7 ± 2.6	(1) This work (2) Lopez and Sola (1981)

alteration and the general trend during basalt weathering is lowering of Ca and Sr (e.g., Patino et al., 2003). Sr content bound to possibly undetected carbonate is removed by leaching of the samples and, therefore, high Sr content (900 ppm is more than three times the value of bulk crust; Taylor and McLennan, 1995) is considered to be a primary feature. We further include alnöites (location: San Andres) and carbonatites (location: Ayopaya; Schultz et al., 2004) in the data set, which show $\text{LOI} \gg 2.5$ wt.% due to primary carbonate. Samples from the Miocene La Peña alkaline intrusion show systematic compositional variation due to magmatic differentiation (Figs. 2a and 4; see below), but little crustal influence in the radiogenic isotopes (Fig. 3b; see below) and no alteration. They are therefore included in our sample set. The selected samples are shown in Tables 2a, 2b and 2c. The complete sample set is in Table A2 of the electronic Appendix and includes previously studied samples of the Las Conchas (Lucassen et al., 2002a) and Ayopaya locations (Schultz et al., 2004; Pb isotope ratios of these samples were analysed in the course of this study) and the excluded samples.

The effect of sample selection is illustrated by the TAS classification diagram (Fig. 2) and Sr–Nd isotope composition (Fig. 3). The TAS classification diagram including all analysed samples of silicate rocks shows a broad compositional range from foidite to dacite (Fig. 2a). The filtering reduces the compositional variation of the silicate rocks to the foidite, basanite, basalt, and trachy-basalt fields (Fig. 2b) and removes most samples suspected of crustal contamination and/or alteration according to their Sr–Nd isotope signatures (Fig. 3a,b; Table 2c). The excluded samples are mainly from the Tupiza and Los Condorez locations. Some Tupiza rocks show severe alteration and the Sr and Nd isotopic composition of other samples indicate mixing with isotope composition of the ambient continental

crust and only the most primitive sample meets the selection criteria (Fig. 3a). The Los Condorez samples have been discussed previously and found to be weakly contaminated by lower crust (Lucassen et al., 2002a). Their least contaminated samples represent a similar source as the rocks from Chaján. We show the respectively least contaminated Tupiza and Los Condorez samples in the isotope and trace element diagrams.

3.2.2. Major and trace elements

All selected samples are alkaline in the TAS classification diagram (Fig. 2b) despite their different ages. Volcanic and sub-volcanic rocks (70 Ma: Betanzos, Chaján, Potrerillos; 100 Ma: Matancillas, Las Conchas, Belén, Dique Cadillal, Finca Rodeo; Table 1; Fig. 1b) plot in the basanite and trachy-basalt fields except one sample of a phono-tephrite. The Cretaceous alkaline intrusions (100 Ma: Ayopaya and Vichacla) comprise similar or more alkaline compositions compared with the volcanic rocks. The Ayopaya location is unique in the sampling area with the occurrence of Ca- and Mg-carbonatites, which were described in detail previously (Schultz, 2004; Schultz et al., 2004). The 300 Ma old San Andres samples are alnöite (foidite in the TAS nomenclature, Fig. 2b). Sodium contents accounts for the bulk alkalinity in most samples (molar $\text{K}_2\text{O}/\text{Na}_2\text{O} < 0.5$ to 0.2, except the Chaján samples with ~ 0.9 –0.6) and only the samples from Matancillas and San Andres are strongly potassic. Systematic variations of major element contents with MgO are small in all groups (not shown). The compositional trends in the Miocene La Peña alkaline intrusion indicate magmatic differentiation of single parent magma (Fig. 2b) and the pyroxenite are the cumulates of the more evolved rocks.

Primitive mantle normalized trace element patterns of the 100 Ma and 70 Ma sample groups show enrichment of incompatible elements, but Rb and K are strongly

Table 2a

XRF data; major element oxides in wt.%, trace elements in ppm; <outside the calibrated range; b.d.l. = not detected

Location	Ayopaya	Vichacla				Matancillas		Las Conchas	Belén			
Sample	140	220900-24	220900-26	220900-32	220900-33	02-70	02-71	170900-5	131000-1	02-93	02-94	02-97
Rock type	Ca-carbonatite	Nepheline gabbroid	Nepheline gabbroid	Nepheline gabbroid	Nepheline gabbroid	Phonotephrite	Phonotephrite	Basanite	Basanite	Basanite	Basanite	Basanite
Age (Ma)	100	120	120	120	120	100	100	100	130	130	130	130
SiO ₂	0.3	44.6	44.2	44.7	43.3	50.3	47.1	45.3	45.3	44.7	44.9	44.5
Al ₂ O ₃	0.1	13.0	14.2	13.9	13.1	19.0	17.3	10.6	14.4	14.6	14.6	14.7
Fe ₂ O ₃	3.4	13.0	12.6	12.8	13.0	7.5	7.4	14.7	13.8	14.1	14.1	13.4
MnO	0.92	0.19	0.21	0.21	0.19	0.14	0.14	0.19	0.26	0.20	0.20	0.17
MgO	0.71	10.0	7.8	8.6	9.4	6.0	6.8	10.6	6.3	7.8	7.6	7.6
CaO	48.3	8.9	8.8	8.8	9.2	5.0	9.7	10.4	8.6	8.1	8.2	8.9
Na ₂ O	0.1	3.7	4.9	3.6	4.2	1.6	2.2	4.3	4.0	3.5	3.4	4.1
K ₂ O	0.03	2.7	3.4	2.8	3.3	7.9	6.7	0.8	2.3	1.8	1.9	1.5
TiO ₂	0.04	3.4	3.3	3.9	3.6	1.9	1.9	2.2	4.0	4.1	4.1	4.1
P ₂ O ₅	1.3	0.58	0.70	0.73	0.68	0.74	0.65	0.92	1.2	1.0	1.1	1.0
L.O.I	41.0	0.8	2.2	1.6	1.2	3.0	6.0	2.4	4.9	3.3	3.7	3.4
Sr	14500	1360	1850	1260	1280	1160	893	788	1120	1140	1100	1020
Ba	8750	939	1100	1230	1080	1910	1550	776	820	913	1240	790
Cr	<3	249	153	189	224	110	138	342	99	97	102	151
Nb	2025	90	137	130	101	152	125	92	96	85	85	83
Ni	<3	206	121	142	173	51	67	236	87	59	60	80
V	32	286	236	300	277	172	171	226	253	224	222	224
Zr	439	374	521	496	393	166	149	253	387	322	320	322

^aLucassen et al. (2002a).

Table 2b

ICP-MS trace element data in ppm

Location	Ayopaya	Vichacla				Matancillas		Las Conchas	Belén			
Sample	140	220900-24	220900-26	220900-32	220900-33	02-70	02-71	170900-5	131000-1	02-93	02-94	02-97
Rb		27	36	23	29	266	201	3	47	47	55	24
Sr	13440	1010	1450	959	883	1120	891	795	1120	1150	1130	1010
Y	185.7	21.4	28.7	32.9	23.5	22.0	20.7	20.1	16.0	31.8	32.2	29.4
La	911	54.6	99.9	83.7	76.2	73.4	64.8	29.5	17.7	68.8	67.7	68.1
Ce	1600	105.4	183.9	158.6	145.4	121.1	108.0	62.4	40.8	137.3	137.1	133.0
Pr	160.4	11.8	19.5	17.6	15.4	12.1	11.1	7.6	5.2	16.1	15.5	15.3
Nd	538.4	44.9	70.1	65.9	56.8	41.4	37.7	31.4	23.1	62.5	60.7	59.0
Sm	86.0	8.3	12.0	12.1	9.6	6.5	6.2	6.5	5.3	11.7	11.5	10.8
Eu	26.8	2.5	3.7	3.6	3.0	2.0	1.8	2.1	2.0	3.6	3.6	3.4
Gd	69.4	7.1	9.7	10.4	7.9	5.4	5.4	5.7	4.8	10.0	10.0	9.1
Tb	9.41	0.96	1.29	1.40	1.04	0.77	0.69	0.83	0.68	1.29	1.32	1.26
Dy	44.63	5.05	6.88	7.66	5.54	4.45	4.20	4.65	3.83	7.12	7.05	6.59
Ho	7.34	0.88	1.19	1.32	0.95	0.83	0.77	0.86	0.66	1.25	1.19	1.12
Er	17.77	2.24	2.87	3.38	2.40	2.35	2.26	2.20	1.82	3.21	3.16	3.01
Tm	2.31	0.27	0.36	0.42	0.31	0.32	0.32	0.29	0.23	0.40	0.40	0.40
Yb	14.19	1.71	2.11	2.50	1.79	2.15	1.93	1.98	1.56	2.48	2.38	2.40
Lu	1.90	0.23	0.29	0.34	0.24	0.32	0.30	0.29	0.23	0.36	0.35	0.35
Hf	0.15	7.86	10.19	10.46	8.19	4.30	4.06	5.58	9.45	9.43	9.26	9.40
Pb	22.16	7.16	8.25	43.05	8.62	3.65	4.44	5.58	5.10	6.88	6.92	7.98
Th	13.15	7.15	11.84	10.19	7.84	11.29	9.22	5.00	6.05	7.90	7.85	7.88
U	3.48	2.52	3.78	3.51	2.57	2.51	2.10	1.47	2.42	2.07	2.04	2.24

^aLucassen et al. (2002a).

Cadillal	Finca Rodeo				Los Condores ^a	Betanzos					
081000-5	02-98	02-74	02-75	02-78	13398-15	230900-36	230900-43	230900-46	02-25	02-28	02-29
Basanite	Basanite	Basanite	Basanite	Basanite	Trachy-basalt	Basanite	Basanite	Basanite	Basanite	Basanite	Basanite
100	130	100	100	100	110	80	80	80	80	80	80
45.2	44.2	41.9	41.3	41.4	46.2	45.0	43.0	42.4	43.0	43.1	43.6
14.8	14.6	13.6	13.6	13.5	11.6	13.9	13.3	14.5	14.0	13.4	14.3
10.7	13.7	12.4	12.5	12.5	11.4	12.5	13.2	14.0	13.8	13.4	13.5
0.19	0.24	0.19	0.18	0.19	0.16	0.28	0.22	0.21	0.19	0.21	0.19
10.4	8.2	10.9	11.0	11.0	11.1	8.2	9.0	7.2	7.5	9.6	7.2
10.3	9.0	12.3	12.4	12.3	10.1	11.1	10.4	10.8	10.5	10.4	10.2
4.4	3.5	4.1	4.3	4.4	3.7	3.3	4.0	4.7	4.4	3.6	4.6
0.8	1.4	0.6	0.8	0.8	1.4	1.4	1.9	1.4	1.9	1.5	1.7
2.2	4.1	3.2	3.2	3.2	3.3	3.3	3.8	3.9	3.9	3.8	3.9
0.94	1.0	0.79	0.79	0.76	0.91	0.88	1.2	0.95	0.84	1.1	0.83
3.4	3.7	3.2	2.4	2.4	4.6	3.2	1.6	1.2	1.2	2.9	1.6
894	1200	965	1300	976	1480	1190	1350	1170	990	1180	1000
755	801	974	958	915	1190	1060	1020	959	861	890	934
324	161	232	239	219	366	274	236	71	98	286	96
94	86	71	71	71	82	143	144	127	103	121	110
228	90	146	147	144	288	150	231	75	56	204	59
215	245	256	261	261	227	275	290	372	319	276	316
259	309	196	181	194	413	411	404	354	275	319	284

	Cadillal	Finca Rodeo			Los Condores ^a	Betanzos					
02-98	081000 -5	02-74	02-75	02-78	13398-15	230900-36	230900-43	230900-46	02-25	02-28	02-29
32	7	13	13	18	15	8	22	15	19	22	14
1180	801	970	1280	989	1410	888	1330	1140	992	1120	990
28.9	21.0	26.7	26.3	27.0	29.3	26.5	23.5	23.2	29.1	29.5	28.4
66.2	28.6	64.2	62.1	63.8	86.0	83.3	44.4	36.1	76.1	92.8	78.5
128.8	60.8	117.6	114.3	119.9	166.8	159.5	93.5	80.6	145.9	172.7	149.6
14.4	7.5	13.1	12.9	13.4	19.7	17.4	11.3	10.1	16.7	19.2	16.8
57.0	31.6	50.1	49.7	51.0	70.8	64.7	45.8	42.2	63.8	71.6	64.1
10.4	6.4	9.2	9.2	9.7	12.5	10.9	8.8	8.5	11.5	12.2	11.4
3.3	2.1	3.0	2.9	3.0	3.7	3.4	3.1	3.1	3.4	3.7	3.4
9.1	6.0	8.2	8.1	8.4	10.4	8.9	7.4	7.5	9.5	9.7	9.7
1.19	0.84	1.10	1.08	1.08	1.31	1.16	1.03	1.01	1.22	1.29	1.26
6.42	4.71	5.91	5.82	5.88	6.51	6.30	5.56	5.62	6.72	6.92	6.63
1.13	0.85	1.02	1.00	1.03	1.11	1.11	0.98	0.94	1.17	1.21	1.17
2.91	2.29	2.58	2.58	2.65	2.72	2.86	2.49	2.49	3.02	3.03	2.96
0.37	0.30	0.33	0.32	0.32	0.34	0.35	0.34	0.33	0.35	0.38	0.37
2.28	1.96	1.93	1.88	1.91	1.98	2.18	2.11	1.98	2.22	2.31	2.22
0.33	0.30	0.27	0.26	0.26	0.29	0.32	0.29	0.30	0.32	0.33	0.31
9.30	6.29	6.49	6.50	6.74	8.51	8.19	9.83	8.93	8.48	9.22	6.95
7.06	5.52	7.10	6.73	7.34	7.68	5.07	4.85	3.78	7.11	7.73	4.27
7.56	5.64	7.84	7.65	7.86	8.63	10.04	10.06	6.54	9.99	11.31	9.42
2.06	1.43	1.53	1.49	1.73	1.26	3.06	3.18	1.70	2.56	3.25	2.60

Table 2a (continued)

Location	Chaján			Potrerillos				South of Tupiza	La Pena	
Sample	121000-17	121000-19	121000-21	03-13b	03-13a	03-15	03-16	02-58	03-2a	03-3
Rock type	Basanite	Basanite	Basanite	Trachy-basalt	Trachy-basalt	Trachy-basalt	Trachy-basalt	Basalt	Pyroxenite	Pyroxenite
Age (Ma)	70	70	70	80	80	80	80	20	20	20
SiO ₂	41.3	42.2	40.7	48.4	48.5	48.2	48.3	48.4	35.3	42.1
Al ₂ O ₃	8.8	9.8	9.5	11.8	11.8	11.9	11.9	17.4	6.2	12.1
Fe ₂ O ₃	12.9	11.9	11.8	10.1	10.0	10.1	10.0	10.9	22.5	14.0
MnO	0.19	0.17	0.18	0.13	0.13	0.13	0.13	0.23	0.24	0.23
MgO	14.6	14.1	14.5	9.6	9.5	9.7	9.5	7.3	10.6	5.1
CaO	12.1	12.5	13.1	9.0	8.9	9.1	9.0	9.0	19.4	13.0
Na ₂ O	2.8	2.0	2.7	4.0	4.2	4.3	4.2	3.1	0.0	1.6
K ₂ O	2.8	2.7	2.5	2.7	2.8	2.3	2.6	1.4	1.0	4.0
TiO ₂	3.8	3.4	3.3	2.9	2.9	2.9	2.9	1.8	2.3	1.7
P ₂ O ₅	0.84	1.38	1.62	1.33	1.33	1.37	1.33	0.40	2.50	1.25
L.O.I	1.9	3.8	2.4	2.6	2.5	3.1	2.7	2.4	0.0	1.3
Sr	2360	2230	2450	1920	1950	1940	1950	407	560	1680
Ba	1340	1230	1370	1110	1100	1070	1090	632	475	876
Cr	401	421	428	264	278	279	270	199	<20	<20
Nb	184	169	191	89	91	91	92	41	6	14
Ni	371	356	398	207	209	211	204	43	<20	<20
V	220	186	171	125	114	123	115	199	586	420
Zr	543	519	566	259	262	261	260	161	98	172

depleted with respect to neighbouring elements (Fig. 4a,c; Chaján samples show moderate Rb and K depletion). Nb contents are high and slightly enriched with respect to Th and U. A more or less pronounced Pb trough is common.

Zr, Hf, Ti, and Y are slightly depleted with respect to neighbouring elements of similar compatibility at some locations. The potassic samples from Matancillas show high K and Rb contents (Tables 2a, 2b). The REE (rare

Table 2b (continued)

Location	Chaján			Potrerillos				South of Tupiza	La Pena	
Sample	121000-17	121000-19	121000-21	03-13b	03-13a	03-15	03-16	02-58	03-2a	03-3
Rb	58	47	47	45	58	41	35	32	37	89
Sr	2970	2840	3100	1950	1900	1910	1760	415	578	1700
Y	27.6	24.7	26.3	24.6	24.7	24.7	23.1	27.1	26.5	29.1
La	119	109	123	71.5	71.3	72.3	68.7	33.7	30.4	42.8
Ce	238	220	250	150.4	150.6	150.0	144.0	67.8	75.2	92.9
Pr	27.8	26.0	29.4	18.8	18.9	19.0	18.3	7.8	11.0	12.5
Nd	108	101	114	76.7	75.0	77.5	74.9	30.8	52.7	54.7
Sm	17.8	16.0	18.0	13.8	13.5	14.0	13.5	6.1	12.2	12.2
Eu	5.13	4.73	5.26	4.1	4.1	4.1	4.0	2.0	3.7	3.6
Gd	12.4	11.5	12.7	10.4	10.8	11.1	10.2	6.1	11.7	10.7
Tb	1.49	1.36	1.50	1.28	1.25	1.27	1.21	0.91	1.37	1.36
Dy	7.00	6.19	6.89	6.12	6.29	6.10	5.97	5.27	6.50	6.86
Ho	1.11	1.01	1.07	0.97	0.95	0.96	0.94	1.05	1.04	1.16
Er	2.50	2.29	2.45	2.20	2.27	2.26	2.24	2.99	2.46	2.83
Tm	0.29	0.26	0.27	0.26	0.26	0.26	0.26	0.39	0.27	0.36
Yb	1.65	1.55	1.59	1.44	1.53	1.55	1.47	2.63	1.57	2.10
Lu	0.25	0.24	0.25	0.20	0.21	0.21	0.19	0.38	0.22	0.31
Hf	9.28	8.93	9.58	7.38	7.23	8.81	7.28	5.15	5.67	5.52
Pb	5.24	5.16	5.11	7.38	7.82	11.16	7.39	25.32	6.03	8.04
Th	8.95	7.54	8.86	7.80	7.98	7.62	6.47	4.70	1.59	4.77
U	2.41	1.93	2.29	2.18	2.41	2.39	2.43	1.12	0.38	1.52

							San Andres			
03-4	03-5	03-10	03-6	03-7	03-9	03-12	02-1	02-4	02-5	02-6
Pyroxenite	Pyroxenite	Pyroxenite	Malignite	Malignite	Syenite	Syenite	Alnöite	Alnöite	Alnöite	Alnöite
20	20	20	20	20	20	20	300	300	300	300
41.3	38.4	36.6	50.6	49.5	55.6	58.9	37.5	39.5	37.0	37.6
16.5	13.1	7.8	19.7	16.8	18.4	19.9	5.0	5.6	4.6	5.4
15.4	16.6	22.1	7.7	10.6	7.4	4.3	13.6	13.7	14.0	12.9
0.32	0.33	0.34	0.32	0.26	0.17	0.18	0.16	0.17	0.17	0.17
5.5	6.3	7.7	1.3	2.9	1.8	0.7	22.2	22.1	23.0	22.8
15.0	15.6	20.3	7.9	9.6	6.6	4.8	13.4	11.5	13.8	13.5
2.1	1.7	0.5	3.7	3.1	4.9	5.6	0.5	0.0	0.0	0.0
0.8	1.1	0.2	7.3	5.2	4.0	5.0	1.9	2.3	1.8	2.4
1.8	1.9	2.5	1.1	1.4	0.9	0.6	4.1	4.1	4.1	4.1
1.18	1.46	1.85	0.29	0.59	0.34	0.12	1.6	1.0	1.6	1.1
0.7	0.7	0.4	2.6	2.0	1.0	1.9	4.5	4.1	3.5	4.4
2150	2000	779	2630	2130	1840	1550	640	622	691	911
529	466	313	1120	1160	1380	1080	771	1030	623	1010
22	21	27	26	21	28	<20	782	816	776	765
7	8	8	21	16	13	12	139	127	135	135
b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	<20	b.d.l.	494	542	527	523
304	420	734	236	301	122	60	174	195	201	203
85	165	194	207	159	180	278	241	248	289	267

earth elements) patterns are similar for the different locations and the La_N/Yb_N ratios range between ~ 15 and 30 except in samples from Chaján, which reach up to 70 (Fig. 5). The carbonatite samples from Ayopaya show extreme trace element contents and variable patterns as

common in carbonatite rocks (Fig. 4b; details in Schultz et al., 2004).

The trace element composition (Fig. 4d) and age (~ 300 Ma) of the San Andres alnöite are very similar to that of the larger sample set from the nearby Los Alisos

						San Andres			
03-5	03-10	03-6	03-7	03-9	03-12	02-1	02-4	02-5	02-6
19	7	194	117	76	119	56	76	52	79
1980	786	2610	2090	1810	1540	638	626	702	915
42.2	42.3	38.8	34.8	30.5	23.9	27.9	23.8	28.9	23.2
61.9	59.6	71.3	52.1	44.2	51.0	148.1	129.4	156.5	125.5
139.6	136.8	147.0	118.4	96.4	97.1	301.9	276.8	334.0	250.5
19.1	18.8	18.3	15.7	12.5	11.4	35.3	33.1	38.5	28.8
84.0	81.0	74.3	64.7	52.1	41.2	139.5	126.8	149.3	108.6
18.0	18.1	14.8	13.7	10.6	7.7	23.3	21.6	25.1	17.9
5.5	5.5	4.5	4.2	3.2	2.3	6.4	5.7	6.9	4.9
15.7	16.9	12.6	11.6	9.1	6.4	17.1	15.3	17.8	13.1
1.97	2.04	1.67	1.51	1.18	0.87	1.91	1.65	2.07	1.48
9.83	10.24	8.41	7.94	6.54	4.81	8.46	7.45	9.09	6.63
1.69	1.63	1.47	1.35	1.19	0.87	1.21	1.07	1.33	0.96
4.10	4.02	3.81	3.62	3.16	2.53	2.63	2.38	2.82	2.27
0.52	0.48	0.50	0.48	0.40	0.34	0.27	0.24	0.28	0.24
3.01	2.79	3.16	2.94	2.56	2.19	1.42	1.31	1.53	1.29
0.44	0.41	0.47	0.40	0.36	0.33	0.20	0.18	0.21	0.19
5.66	5.87	6.94	6.90	4.90	5.73	8.44	6.81	8.08	6.98
1.92	3.25	19.06	9.00	5.13	18.20	4.76	3.96	2.57	4.83
3.23	4.54	8.90	5.13	5.24	7.24	19.96	17.59	19.64	16.30
0.88	0.75	3.38	1.78	1.58	2.36	3.10	2.71	3.16	2.80

Table 2c

Nd, Pb, and Sr isotope data; (t) indicates ratios corrected for in-situ decay

Location	Ayopaya	Vichacla				Matancillas		Las Conchas	Belén	
Sample	I40	220900-24	220900-26	220900-32	220900-33	02-70	02-71	170900-5	131000-1	02-93
Age (Ma)	100	110	110	110	110	100	100	100	130	130
$^{87}\text{Sr}/^{86}\text{Sr}$	0.703291	0.705192	0.705147	0.704579	0.704092	0.707424	0.707118	0.703617	0.703601	0.703799
$\pm 2\sigma$ (m)	0.000008	0.000010	0.000010	0.000008	0.000008	0.000007	0.000007	0.000008	0.000007	0.000008
$^{87}\text{Rb}/^{86}\text{Sr}$		0.076	0.073	0.070	0.093	0.682	0.651	0.015	0.121	0.117
$^{87}\text{Sr}/^{86}\text{Sr}(t)$	0.70329	0.70508	0.70504	0.70448	0.70396	0.70645	0.70619	0.70360	0.70343	0.70363
$^{143}\text{Nd}/^{144}\text{Nd}$	0.512794	0.512507	0.512522	0.512301	0.512486	0.512730	0.512756	0.512830	0.512775	0.512753
$\pm 2\sigma$ (m)	0.000006	0.000008	0.000009	0.000007	0.000008	0.000006	0.000006	0.000008	0.000008	0.000006
$^{147}\text{Sm}/^{144}\text{Nd}$	0.0967	0.1119	0.1038	0.1110	0.1026	0.0953	0.0994	0.1241	0.1380	0.1128
$^{143}\text{Nd}/^{144}\text{Nd}(t)$	0.51273	0.51243	0.51245	0.51223	0.51242	0.51267	0.51269	0.51275	0.51268	0.51268
$^{206}\text{Pb}/^{204}\text{Pb}$	carbonate	18.72	18.76	18.45	18.64	19.36	19.11	18.89	19.76	19.58
$^{207}\text{Pb}/^{204}\text{Pb}$		15.54	15.54	15.60	15.53	15.60	15.60	15.57	15.60	15.58
$^{208}\text{Pb}/^{204}\text{Pb}$		38.84	38.96	38.59	38.81	39.59	39.29	38.73	39.28	39.47
$^{238}\text{U}/^{204}\text{Pb}$		25.9	33.8	6.0	21.9	51.6	35.2	19.4	35.6	22.6
$^{232}\text{Th}/^{204}\text{Pb}$		74.0	106.6	17.4	67.3	234.0	155.8	66.5	89.7	86.8
$^{206}\text{Pb}/^{204}\text{Pb}(t)$	19.20	18.32	18.23	18.35	18.30	18.56	18.56	18.58	19.21	19.23
$^{207}\text{Pb}/^{204}\text{Pb}(t)$	15.64	15.52	15.51	15.60	15.51	15.56	15.58	15.56	15.57	15.56
$^{208}\text{Pb}/^{204}\text{Pb}(t)$	39.15	38.47	38.43	38.50	38.48	38.43	38.52	38.40	38.83	39.04

^aLucassen et al. (2002a).

alnöite (Barbieri et al., 1997; Villar et al., 1997) and both most likely represent the same magmatic pulse. The REE element pattern of the alnöite is steep with La_N/Yb_N of ~ 70 (Fig. 5).

The high Nb content (Table 2a, Table A2 in the Appendix), the absence of a Nb trough, and the patterns of the more compatible elements and REE (Fig. 4c) of the most primitive Tupiza sample (20 Ma) are similar to those of the older rocks, but the element contents are somewhat lower in the Tupiza sample and a prominent positive Pb spike occurs. The composition of the Miocene La Peña samples is likely controlled by magmatic differentiation as seen in the evolution of the trace element patterns of pyroxenite, which forms the most primitive cumulate, malignite, and syenite (Fig. 4c) and the correlation of major and trace elements (not shown). The low Nb contents and deep Nb troughs of all La Peña samples resemble the pattern of average late Cenozoic andesite from the magmatic arc (Fig. 4c).

3.3. Radiogenic isotopes

The effects of age correction for radioactive in-situ decay are small for most samples because Rb/Sr and Sm/Nd ratios are low (Table 2c; Table A2 in the Appendix). This enables the comparison of the Sr and Nd isotope compositions of samples of different ages of crystallization. Sr and Nd initial isotope ratios define different compositional arrays (Fig. 3b). Many samples of the

100 Ma group (Ayopaya, Las Conchas, Belén, Cadillal, and Finca Rodeo) fall into a small field of the depleted mantle quadrant ($^{143}\text{Nd}/^{144}\text{Nd} \sim 0.5127\text{--}0.5128$; $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.7032\text{--}0.7040$), which is similar or slightly more radiogenic in its Sr isotope composition than HIMU mantle (HIMU = high μ ; $\mu = ^{238}\text{U}/^{204}\text{Pb}$; e.g., Zindler and Hart, 1986; Hofmann, 1997). Samples from the 70 Ma (Betanzos) and 20 Ma (La Peña) groups also plot within this field. A few samples from Las Conchas and Ayopaya (100 Ma), from Matancillas (100 Ma), and the Tupiza sample (20 Ma) have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of up to ~ 0.707 and radiogenic Nd isotope signatures ($\sim 0.5127\text{--}0.5128$). Chaján and Potrerillos (70 Ma), Vichacla and Los Condorez (100 Ma), and San Andres (300 Ma) samples all plot in the field of enriched mantle composition. Their compositions are similar to EM1 (EM = enriched mantle; Zindler and Hart, 1986) and the composition of SCLM from the Brazilian Shield (Fig. 3b).

The distinction into a depleted and enriched source by the Nd (and Sr isotope ratios for many samples) ratios is largely consistent with the distinction of the respective samples in the $^{207}\text{Pb}/^{204}\text{Pb}\text{--}^{206}\text{Pb}/^{204}\text{Pb}$ compositional field (Fig. 6a). The samples with a depleted mantle source and the samples with radiogenic Sr and Nd isotope composition show similar $^{207}\text{Pb}/^{204}\text{Pb}$ ratios (15.55–15.65, less spread at single locations), but a considerable range of $^{206}\text{Pb}/^{204}\text{Pb}$ ratios ($\sim 18.5\text{--}19.5$). HIMU mantle compositions are absent. A clear correlation between

			Cadillal	Finca Rodeo			Los Condores ^a	Betanzos		
02-94	02-97	02-98	081000-5	02-74	02-75	02-78	13398 -15	230900 -36	230900-43	230900-46
130	130	130	100	100	100	100	110	80	80	80
0.704039	0.704158	0.705162	0.703694	0.704032	0.704698	0.703986	0.705546	0.703375	0.703447	0.703205
0.000008	0.000007	0.000006	0.000010	0.000007	0.000007	0.000008	0.000004	0.000008	0.000008	0.000011
0.141	0.068	0.078	0.024	0.039	0.030	0.051	0.035	0.027	0.049	0.037
0.70384	0.70406	0.70505	0.70366	0.70398	0.70466	0.70391		0.70334	0.70338	0.70315
0.512759	0.512769	0.512771	0.512830	0.512818	0.512806	0.512800	0.512273	0.512813	0.512822	0.512832
0.000005	0.000006	0.000007	0.000007	0.000006	0.000006	0.000007	0.000006	0.000009	0.000007	0.000006
0.1144	0.1107	0.1105	0.1222	0.1116	0.1120	0.1145	0.1092	0.1023	0.1161	0.1220
0.51268	0.51270	0.51270	0.51275	0.51275	0.51273	0.51273	0.51220	0.51275	0.51275	0.51275
19.60	19.71	19.70	18.88	19.37	19.40	19.41	18.22	19.90	20.10	20.07
15.59	15.59	15.59	15.57	15.60	15.58	15.59	15.51	15.63	15.61	15.61
39.49	39.54	39.56	38.71	39.44	39.42	39.40	38.93	39.71	39.76	39.71
22.1	21.1	22.0	19.1	16.1	16.6	17.7	12.0	45.7	49.8	34.1
85.8	74.8	81.3	75.8	83.3	85.6	80.7	82.8	151.0	158.6	132.2
19.26	19.38	19.35	18.58	19.12	19.14	19.14	18.03	19.18	19.32	19.53
15.57	15.57	15.58	15.56	15.59	15.57	15.58	15.50	15.59	15.57	15.59
39.07	39.17	39.16	38.33	39.03	38.99	39.00	38.52	38.96	38.97	39.06

$^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios only is seen among the La Peña samples. The Pb isotope ratios of these samples (including $^{208}\text{Pb}/^{204}\text{Pb}$ ratios) also decrease with increasing Sr (Fig. 6b) and decreasing Nd isotope ratios (not shown). The $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of most samples from a depleted mantle source and radiogenic Sr composition are strongly correlated with $^{206}\text{Pb}/^{204}\text{Pb}$, parallel to NHRL (Fig. 6b).

Uranogenic Pb isotope ratios of the samples from an enriched mantle source plot in or near the large Brazilian SCLM field. Their $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ ratios vary between 17.5 and 18.5 and 15.45 to 15.50, with somewhat higher $^{207}\text{Pb}/^{204}\text{Pb}$ ratios in the San Andres samples. Samples from an enriched source form two groups according to their $^{208}\text{Pb}/^{204}\text{Pb}$ ratios. One group shows $^{208}\text{Pb}/^{204}\text{Pb}$ ratios significantly higher than the NHRL values at given $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (similar to EM1; Vichacla, San Andres, Los Condores), the $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of the other group (Chaján, Potrerillos) are similar to NHRL. Three Ayopaya carbonatite samples show very high $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of ~ 41 to 42.7 (not shown; Table A2 in the Appendix).

4. Discussion

The interpretation of data from mantle-derived magmatic rocks from continental settings in terms of mantle composition must consider (i) the role of crustal contamination versus primary mantle composition and

(ii) the types of mantle involved, which could be lithospheric or asthenospheric.

4.1. Crustal contamination or primary mantle signature?

Addition of crustal material to mantle-derived magmas could occur in the mantle source by influx of crustal material in a subduction system ('source contamination'), during the ascent and storage of the magmas within the crust ('magma contamination'), and by influx of material from the country rocks during low-grade alteration or weathering at the intrusion/extrusion level ('alteration'). We restrict the discussion to the latter two processes with a focus on the isotopic composition of the samples.

Effects of alteration or obvious crustal contamination are largely avoided by sample selection (Fig. 3a) and a few general considerations are valid for all selected samples of the intra-plate magmatic rocks. (1) The rocks represent, according to their trace element enrichment, small degrees of melting in the mantle and small volumes of melt. They have high Sr and Nd contents (after leaching) which are above typical values of crustal rocks. Any substantial change of Sr or Nd isotope ratios by crustal contamination in magmatic processes or low temperature alteration should be accompanied by major changes in the chemistry or the state of preservation of the rock. (2) The Pb contents of many samples are lower than in typical felsic crustal

Table 2c (continued)

Betanzos			Chaján			Poterillos				South of Tupiza	La Peña
02-25	02-28	02-29	121000-17	121000-19	121000-21	03-13b	03-13a	03-15	03-16	02-58	03-2a
80	80	80	70	70	70	80	80	80	80	20	20
0.703367	0.703875	0.703413	0.704160	0.704176	0.704114	0.704145	0.704136	0.704124	0.704190	0.705066	0.703682
0.000010	0.000007	0.000008	0.000008	0.000008	0.000007	0.000008	0.000008	0.000010	0.000008	0.000007	0.000008
0.055	0.056	0.042	0.116	0.105	0.071	0.067	0.088	0.063	0.057	0.225	0.183
0.70329	0.70380	0.70335	0.70400	0.70403	0.70401	0.70405	0.70401	0.70404	0.70411	0.70500	0.70363
0.512806	0.512800	0.512808	0.512435	0.512437	0.512418	0.512402	0.512374	0.512375	0.512376	0.512666	0.512827
0.000006	0.000006	0.000003	0.000007	0.000010	0.000007	0.000007	0.000006	0.000005	0.000007	0.000006	0.000007
0.1085	0.1033	0.1081	0.0961	0.0940	0.0947	0.1088	0.1087	0.1093	0.1090	0.1192	0.1405
0.51273	0.51273	0.51274	0.51237	0.51238	0.51236	0.51233	0.51230	0.51230	0.51230	0.51265	0.51281
19.88	20.13	19.95	18.51	18.34	18.44	18.51	18.51	18.47	18.49	18.49	18.81
15.62	15.63	15.62	15.50	15.50	15.51	15.55	15.51	15.51	15.51	15.63	15.59
39.64	39.93	39.72	38.35	38.22	38.29	38.12	38.00	38.00	38.01	38.61	38.73
27.2	32.0	46.0	48.6	40.6	44.8	21.4	22.3	15.5	23.9	3.2	4.6
107.0	112.2	168.2	172.5	136.0	138.8	77.3	74.5	49.8	63.9	13.7	19.6
19.45	19.63	19.24	17.75	17.71	17.74	18.17	18.16	18.23	18.12	18.48	18.79
15.60	15.60	15.59	15.46	15.47	15.47	15.53	15.49	15.50	15.50	15.63	15.59
39.11	39.38	38.88	37.50	37.55	37.60	37.74	37.63	37.75	37.70	38.59	38.71

rocks and the Pb isotope system is potentially more sensitive to contamination by crust than the Sr and Nd systems. (3) The radiogenic isotope compositions of the mantle – similar to those of the overlying crust – are subject to time-related changes, i.e., after the change of parent/daughter ratios during metasomatism in the lithospheric part of the mantle.

4.1.1. Nd and Sr isotopes

The crust at the western margin of the Brazilian Shield evolved during mid-Proterozoic and early Palaeozoic orogenies (e.g., Coira et al., 1982; Lucassen et al., 2001). Its composition is that of old felsic crust and Nd and Sr isotope composition of the bulk volume of the crust plots into the lower right quadrant of the Nd–Sr diagram (Fig. 7a). Nd isotope patterns of the Palaeozoic crust are similar at distant sections even on the large regional scale of our sampling area (e.g., Pankhurst and Rapela, 1998; Bock et al., 2000; Lucassen et al., 2001, 2004; Egenhoff and Lucassen, 2003). The Sr isotopic composition can be variable between high grade metamorphic rocks (e.g., granulite facies lower crust of Early Palaeozoic age, ~ 0.715 ; Lucassen et al. 1999, 2001) and other Palaeozoic crustal rocks with more radiogenic Sr isotope compositions ($^{87}\text{Sr}/^{86}\text{Sr} > 0.715$).

Sr and Nd isotopic ratios of many intra-plate samples assemble two restricted compositional fields one indicating a depleted and second indicating an enriched reservoir (Fig. 3b). Systematic compositional variations in these fields are minor and it is not possible to

distinguish small amounts of crustal contamination from small variations in the mantle source composition. Considering the large sampling area, the depleted and enriched mantle sources are of rather uniform Nd and Sr isotope composition (Fig. 3b). Systematic variations of the Sr–Nd isotope compositions in these two groups are restricted to samples from La Peña (two deviating data points in Fig. 3b). Nd and Sr isotope compositions (and correlation of the Pb isotope ratios in these samples; Fig. 6) indicate mixing of melts from depleted mantle with small amounts of an isotopically different material with unradiogenic Nd (and Pb) and radiogenic Sr isotope composition. The isotope ratios of the La Peña samples vary with indicators of magmatic differentiation (not shown), which indicates a process occurring within the crust. The crustal material could have lower crustal composition with a Pb and Nd isotopic composition comparable to that of lower crust xenoliths near 31°30'S (Kay et al., 1996; no Sr isotopes published).

A small number of intra-plate samples have variable high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of up to 0.707 and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios similar to those of the depleted mantle samples (Fig. 3b). Such Nd–Sr isotope signatures are known to be caused by seawater alteration (e.g., Hauff et al., 2003), which, however, is very unlikely in intra-plate magmatic rocks deposited in a continental environment. A second alteration process which could explain the high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is the formation of caliche during weathering under arid conditions. Caliche is not observed in the samples and possible hidden carbonate

								San Andres			
03-3	03-4	03-5	03-10	03-6	03-7	03-9	03-12	02-1	02-4	02-5	02-6
20	20	20	20	20	20	20	20	300	300	300	300
0.704003	0.703931	0.703760	0.703876	0.703851	0.703958	0.704611	0.704521	0.705325	0.705675	0.705041	0.704968
0.000010	0.000008	0.000008	0.000007	0.000007	0.000008	0.000008	0.000008	0.000007	0.000010	0.000008	0.000010
0.152	0.015	0.028	0.027	0.214	0.162	0.121	0.222	0.255	0.349	0.213	0.248
0.70396	0.70393	0.70375	0.70387	0.70379	0.70391	0.70458	0.70446	0.70424	0.70419	0.70413	0.70391
0.512782	0.512778	0.512794	0.512808	0.512777	0.512729	0.512670	0.512614	0.512554	0.512583	0.512584	0.512539
0.000006	0.000005	0.000007	0.000006	0.000004	0.000006	0.000006	0.000004	0.000006	0.000007	0.000005	0.000004
0.1349	0.1328	0.1296	0.1353	0.1201	0.1284	0.1225	0.1127	0.1011	0.1028	0.1015	0.0997
0.51276	0.51276	0.51278	0.51279	0.51276	0.51271	0.51265	0.51260	0.51236	0.51238	0.51238	0.51234
18.91	18.74	19.05	18.85	18.95	18.72	18.55	18.60	23.52	21.15	23.02	20.34
15.63	15.58	15.64	15.59	15.60	15.56	15.56	15.56	15.85	15.74	15.84	15.69
38.85	38.63	39.13	38.84	38.91	38.56	38.42	38.46	48.92	44.39	48.23	42.15
13.9	2.4	34.1	17.0	13.1	14.5	22.4	9.5	57.7	55.9	107.6	45.5
44.1	13.1	126.1	103.8	34.7	42.1	75.1	29.3	374.4	365.8	673.4	267.2
18.87	18.73	18.94	18.80	18.91	18.67	18.48	18.57	20.77	18.49	17.90	18.17
15.62	15.57	15.63	15.59	15.60	15.56	15.56	15.56	15.71	15.60	15.57	15.57
38.81	38.62	39.00	38.74	38.88	38.52	38.34	38.43	43.33	38.92	38.16	38.16

was removed by leaching of the rock powders. Mixing with crustal melts or assimilation of crustal material could principally change the Sr isotope composition. The specific Nd–Sr isotope signature of the samples and their major and trace element composition preclude substantial contamination by an AFC process within the crust. Firstly, crustal reservoirs with radiogenic Nd and radiogenic Sr isotope compositions are largely restricted to mafic (meta)igneous rocks with metasomatic changes of their Rb and/or Sr budget. Partial melts from such rocks have very different major and trace element compositions (e.g., Beard and Lofgren, 1991; Rapp and Watson, 1995) compared with those observed in the intra-plate magmatic rocks. Secondly, up to 50 wt.% additions of crustal material would be needed to explain the observed scatter of Sr isotopes at the given high Sr content of the samples (~three times average crust Tables 2a, 2b and 2c and Table A2) and an initial $^{87}\text{Sr}/^{86}\text{Sr}$ composition of ~0.703. Crustal contamination of that extent should cause prominent changes in the Pb and Nd isotope and major and trace element composition, which is not observed in the high $^{87}\text{Sr}/^{86}\text{Sr}$ samples.

A strong argument for a source with radiogenic Sr isotope composition within the upper mantle comes from the composition of spinel lherzolite xenoliths in the late Cretaceous basanite of the Las Conchas location (Lucassen et al., 2005). Clinopyroxenes from some xenoliths samples have radiogenic Nd and radiogenic Sr isotope compositions similar to the high $^{87}\text{Sr}/^{86}\text{Sr}$ samples of the intra-plate magmatic rocks (Fig. 7a).

The addition of radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ to the clinopyroxene of the mantle xenoliths by weathering can be excluded and the high $^{87}\text{Sr}/^{86}\text{Sr}$ xenoliths therefore represent a variation in the composition of the upper mantle (Lucassen et al., 2005).

Such radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ compositions in mantle rocks can be explained by time-integrated growth of ^{87}Sr from elevated Rb-content, (i.e. in phlogopite with high Rb/Sr ratios). Dehydration melting of phlogopite-bearing lithospheric mantle is a common phenomenon during heating of mantle lithosphere and produces potassic melts (e.g., Comin-Chiaramonti et al., 1997; Gibson et al., 1999). The potassic high Rb and Ba rocks from Matancillas (Tables 2a, 2b and 2c) represent such primary melts from dehydration melting. Clinopyroxene as the principal host of Sr in anhydrous peridotite will preserve the high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios after dehydration melting. Melting of such peridotite will produce non-potassic melts with high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

4.1.2. Pb isotopes

Pb isotope composition of the crust is locally and regionally more variable (e.g., Tilton and Barreiro, 1980; Macfarlane et al., 1990; Aitchison et al., 1995; Tosdal, 1996; Kay et al., 1996; Lucassen et al., 2002b) than the Nd isotope composition. U is easily depleted and redistributed during high grade metamorphism and crustal melting, which occurred repeatedly during the protracted evolution of the Andean crust from Proterozoic onwards (e.g., Lucassen et al., 2002b). In the $^{206}\text{Pb}/^{204}\text{Pb}$ – $^{207}\text{Pb}/^{204}\text{Pb}$ space, the composition of

Palaeozoic and Proterozoic basement rocks from the Andes follows a common trend (Fig. 6a) with consistently higher $^{207}\text{Pb}/^{204}\text{Pb}$ ratios than the Northern Hemisphere Reference Line (Hart, 1984).

The Pb isotope compositions of most intra-plate magmatic rocks are – despite the coincidence of individual samples with crustal compositions – not systematically related to the Pb isotope compositions of the crustal sources (Fig. 6a). Pb provinciality in the study area (CVZ north of 21°S ; see Fig. 9b below; Aitchison et al., 1995), which is evident in crust-contaminated Cenozoic andesite, is absent in the intra-plate samples. Further, the systematic differences between Pb isotope composition of the samples from depleted and enriched mantle sources argue for the dominance of primary differences reflecting Pb compositions of the mantle sources. Intra-plate magmatic rocks from depleted and enriched mantle sources passed through a compositionally similar crust and should, in the case of prominent crustal influence, show a uniform crustal Pb isotope composition. The scatter of $^{206}\text{Pb}/^{204}\text{Pb}$ ratios at similar $^{207}\text{Pb}/^{204}\text{Pb}$ in many clinopyroxenes from mantle xenoliths ratios is comparable to the scatter observed in the intra-plate magmatic rocks (Fig. 7b).

4.2. Influence of an asthenospheric mantle source?

The discussion in the previous chapter presented evidence and interpretations in favour of a mantle lithosphere-origin of the common intra-plate magmatic rocks from depleted sources. Primary potassic melts (Matancillas; San Andres), magmas with high $^{87}\text{Sr}/^{86}\text{Sr}$ and high $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, and intra-plate magmatism with isotopic compositions typical for SCLM from the Brazilian Shield indicate an isotopically variable lithospheric mantle.

However, the influence of material from an asthenospheric source cannot be excluded a priori, especially in the intra-plate magmatic rocks from depleted sources. The isotopic composition of the average Pacific MORB is considered as a proxy for the composition of asthenospheric mantle. Compared with this asthenospheric mantle, the Nd isotope evolution in the depleted mantle source of the intra-plate magmatic rocks is retarded and Sr and Pb isotope compositions are more radiogenic (Fig. 7). Mixing trajectories between asthenospheric mantle MORB type mantle and depleted intra-plate magmatic rocks are absent, however (Figs. 3 and 6). The dominance of a depleted lithospheric mantle source is a reasonable assumption to explain the isotope composition of the depleted intra-plate rocks, if a MORB-type mantle source is assumed for the asthenosphere. The variations in the

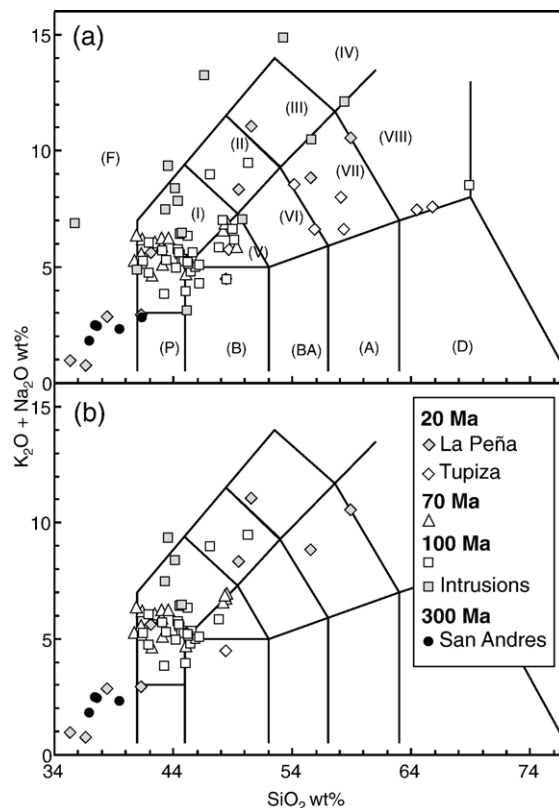


Fig. 2. Compositional variations of the intra-plate magmatic rocks in the TAS classification diagram in terms of age groups; compositions are recalculated to dry 100%. (a) Data for all investigated silicic rocks (see Table A2 in the Appendix); (b) selected silicic rocks based on the alteration and contamination criteria as discussed in the text; the complete selection includes carbonatites from Ayopaya not shown in this diagram. Key to the compositional fields (a,b): (F): foidite; (P): picro-basalt; (B): basalt; (BA): basaltic andesite; (A): andesite; (D): dacite; (R): rhyolite (I): basanite; (II): phono-tephrite; (III): tephri-phonolite; (IV): phonolite; (V): trachy-basalt; (VI): basaltic trachy-andesite; (VII): trachy-andesite; (VIII): trachyte; Le Maitre (1989).

isotope signatures in the depleted intra-plate magmatic rocks and the mantle xenoliths could be explained by time-integrated growth — as outlined below for the case of uraniumogenic Pb in a variably metasomatized lithospheric mantle. Relatively small scale compositional heterogeneities are more likely preserved over a long time in lithospheric than asthenospheric, convective mantle.

4.2.1. Comparison with sub-arc asthenosphere

The assumption of a MORB-type source composition for the convective mantle is not a unique option at an active continental margin. The composition of the convective mantle in the mantle wedge is modified by metasomatism as long as subduction continues. Mantle-derived rocks, which give insight into the composition

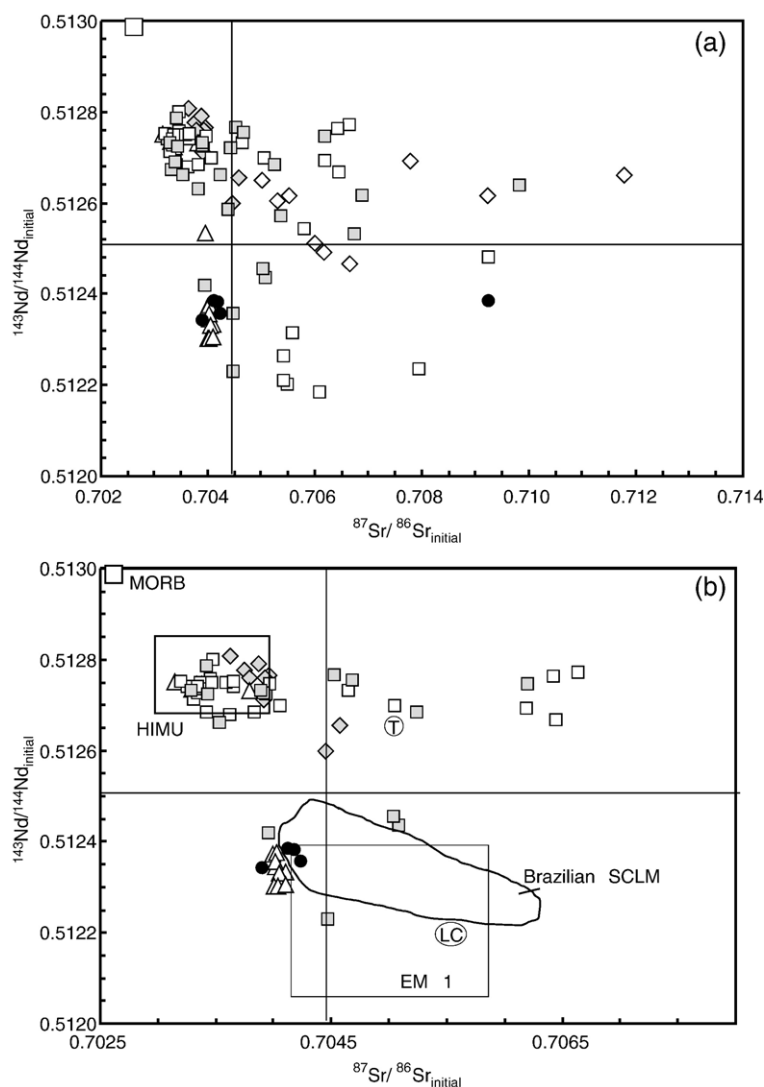


Fig. 3. Nd and Sr isotope ratios of the intra-plate magmatic rocks in terms of age groups (see Fig. 1), corrected for radioactive in-situ decay using the ages from Table 1; (a) all samples (b) sample selection based on the alteration and contamination criteria as discussed in the text, but including carbonatites from Ayopaya and the least contaminated samples from Tupiza (T) and Los Condores (LC); note different scale on the $^{87}\text{Sr}/^{86}\text{Sr}$ -axis. For comparison, the fields of common mantle 'endmember compositions' are shown; MORB = average Pacific MORB compiled from the Petrological Database of the Ocean Floor <http://www.petdb.org>; the Nd isotope ratio of average MORB is corrected for in-situ decay to 100 Ma using a $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.214; HIMU from Hofmann (1997); EM1 Zindler and Hart (1986); SCLM = sub-continental lithospheric mantle beneath the Brazilian Shield (Carlson et al., 1996; Gibson et al., 1996, 1999; Comin-Chiaramonti et al., 1997, 2005; Velázquez et al., 2006); Sr isotopic composition of bulk silicate Earth is calculated at 160 Ma assuming Rb/Sr ratio of 0.03 (McDonough and Sun, 1995) and an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.69898 (Allegre et al., 1975).

of the mantle wedge (asthenosphere) after subduction derived metasomatism, have been studied along the Mesozoic to Recent arcs (18° – 40°S ; Fig. 1; e.g., Hildreth and Moorbath, 1988; Rogers and Hawkesworth 1989; Morata and Aguirre, 2003; Lucassen et al., 2004, 2006; Kramer et al., 2005; and references therein). They occur in the Jurassic–lower Cretaceous ($\sim 18^{\circ}$ – 32°S) arc and in the Jurassic–Recent magmatic rocks of the

Southern Volcanic Zone (Fig. 1b). The age difference between the Jurassic to lower Cretaceous mantle wedge-derived magmas and the Cretaceous intra-plate magmatism does only cause smaller differences in radiogenic isotope composition from time-integrated radiogenic growth in possible common mantle sources. It should be mentioned, however, that the location of the Cretaceous intra-plate magmatism was up to 800 km east of

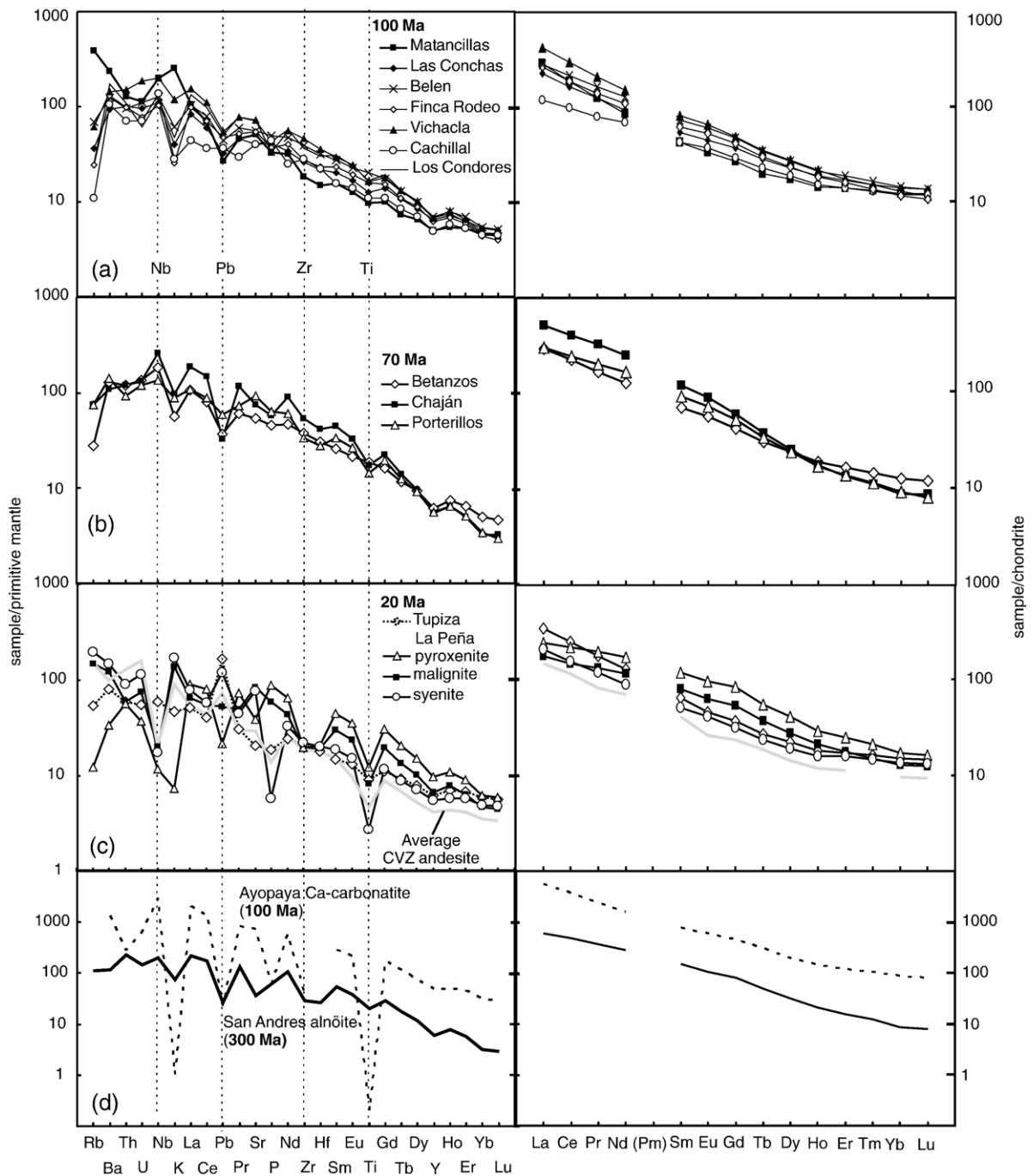


Fig. 4. Selected major and trace element patterns normalized to the primitive mantle (McDonough and Sun, 1995) and REE patterns normalized to the C1 chondrite (McDonough and Sun, 1995). The elements are listed with increasing compatibility in the mantle residue to the right. (a) Average compositions of the 100 Ma group samples. The Matancillas sample shows pronounced enrichment of Rb, Ba, and K. (b) Average compositions of the 70 Ma group samples; they are similar as the 100 Ma group, except for the steeper REE pattern of the Chaján samples. (c) Average compositions of the 20 Ma group samples; pyroxenite from La Peña represents an early cumulate, whereas malignite and syenite represent more evolved magmas. The average composition of Cenozoic Andesite from the Central Volcanic Zone is shown for comparison (compiled from GEOROC data base; <http://georoc.mpch-mainz.gwdg.de/georoc>). (d) Extreme composition of the Ayopaya carbonatite and the San Andres alnöite (average values).

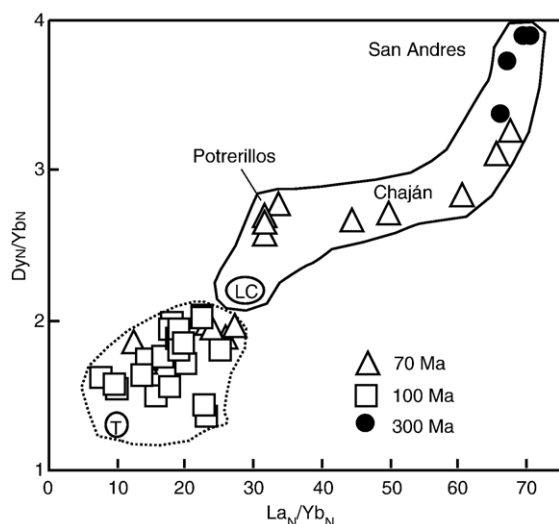


Fig. 5. La_N/Yb_N and Dy_N/Yb_N ratios. Samples from enriched sources (Chaján, San Andres, and Potrerillos) indicate garnet in the source. All other samples show variations, which could be explained either by garnet free spinel lherzolite or small amounts of garnet in the mantle source.

the Jurassic magmatic arc (before shortening of the lithosphere in the Cenozoic orogeny) and the use of the sub-arc mantle as a proxy for the regional sub-continental asthenosphere remains speculative (but certainly not less ‘realistic’ than the comparison with compositions of standard mantle ‘endmembers’ or the Pacific MORB source).

The Nd and Sr isotopic compositions of the arc-related rocks, the intra-plate magmatic rocks from a depleted source, and many mantle xenoliths are very similar (Fig. 7a). Mixing between melts from such lithospheric mantle and sub-arc asthenospheric mantle cannot be detected in the Nd and Sr isotope signatures of the respective hybrid magmatic rocks. The $^{206}Pb/^{204}Pb$ ratios are substantially lower and more uniform in the arc-related rocks than in the intra-plate rocks and most mantle xenoliths (Fig. 7b). The variation of the $^{207}Pb/^{204}Pb$ ratios is similar in arc-related and intra-plate rocks (Fig. 7b). Provided the highest $^{206}Pb/^{204}Pb$ ratios of intra-plate magmatic rocks and mantle xenoliths indicate one ‘endmember’ composition in the lithospheric mantle (Fig. 7b), mixing of melts from such source with melts from the sub-arc mantle could account for some of the scatter in the $^{206}Pb/^{204}Pb$ ratios of the intra-plate rocks from depleted sources.

4.3. The Cretaceous depleted lithospheric mantle

As shown above, the isotopic compositions of the Cretaceous intra-plate magmatic rocks from depleted

mantle sources are different from published endmember compositions (HIMU, EM1), average Pacific MORB mantle or any consistent mixing arrays between them (Figs. 3b and 6). They are, however, similar to the isotopic compositions of mantle xenoliths from local lithospheric mantle. The variations in $^{206}Pb/^{204}Pb$ (Figs. 7 and 8) and rare high $^{87}Sr/^{86}Sr$ ratios are more readily explained by lithospheric mantle sources than asthenospheric mantle. This depleted lithospheric mantle likely formed during the early Palaeozoic orogeny at the active early Palaeozoic continental margin. It is compositional different from Brazilian SCLM, which has been separated from depleted convective mantle since more than 1 Ga (e.g., Comin-Chiaromonte et al., 2001).

4.3.1. A conceptual model for the Cretaceous depleted lithosphere

The time span between early Palaeozoic and Cretaceous could explain peculiarities observed in the Nd, Pb, and Sr isotope systems by time-integrated radiogenic growth after relative Nd, U, and Rb enrichment in the early Palaeozoic depleted sub-arc mantle. The origin of early Palaeozoic replacement of older lithospheric mantle by depleted mantle is speculative. The formation of a mantle wedge during early Palaeozoic subduction is certainly one process of replacement of older lithosphere, but replacement may have included the same processes which were invoked for other orogens with overthickened crust and active margins. Delamination of mafic lower crust and mantle lithosphere with the subsequent influx of convective mantle (asthenosphere) seems to be common in fossil and modern orogens (e.g., Meissner and Mooney 1998; for the present Central Andes: Kay et al., 1994; Sobolev et al., 2006). Metasomatism of the sub-arc mantle by slab derived trace elements (e.g., Grove et al., 2002 among many others) is continuous as long as subduction proceeds. The Sr, Nd, and Pb isotope compositions of the Mesozoic to Recent mantle wedge at the Andean margin are an example of such process, which produced a relatively uniform signature different from other convective mantle reservoirs as e.g., MORB-type mantle (Lucassen et al., 2006 and references therein). The modified composition in the mantle wedge may be frozen after cease of subduction or change of the focus of the subduction zone by conversion of convective to lithospheric mantle during cooling and form the initial ‘baseline’ composition for the further evolution. Metasomatism at decreasing temperatures in the waning subduction system could raise the U/Pb and Rb/Sr and lower Sm/Nd ratios with respect to convective mantle values. Fig. 8 illustrates a simple conceptual model for the growth of radiogenic Pb in the source of the intra-plate rocks from early Palaeozoic onwards. The transformation of

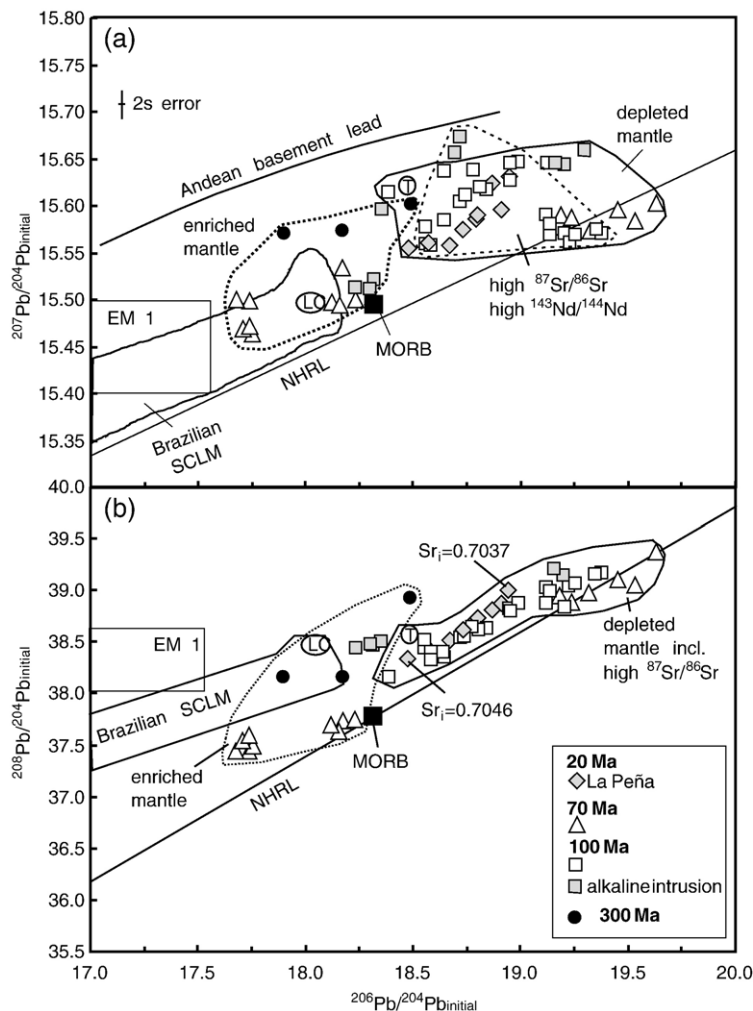


Fig. 6. Pb initial isotope compositions of the samples in comparison to the composition of Andean crust, Brazilian SCLM, average Pacific MORB, and EM1 mantle endmembers. (a) $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. (b) $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$. Three carbonatite samples from Ayopaya with very high $^{208}\text{Pb}/^{204}\text{Pb}$ ratios ($\sim 41\text{--}42.7$; Table A2 in the Appendix) are not shown. Both sources, depleted and enriched mantle as defined by the Nd–Sr-isotope ratios are also distinguished in the Pb isotope ratios. Simple mixing relations between a depleted MORB-type or an enriched (EM1) type mantle with the Paleozoic crust are absent. Only the samples from La Peña (grey diamonds) form a consistent array from a depleted mantle with a crustal component, the latter with a strongly unradiogenic Pb isotope composition (Fig. 6b; see text). Data sources as in Fig. 3; Andean basement lead evolution line from Lucassen et al. (2002b).

convective sub-arc into lithospheric mantle is followed by metasomatism in Early Paleozoic. The initial $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ of the Pacific MORB source at 470 Ma is assumed as a hypothetic starting condition in the early Palaeozoic depleted sub-arc mantle (Fig. 8). Slightly heterogeneous initial Pb isotope ratios and a rise of the μ ratio to values of 20 to 30 at ~ 470 Ma would result in radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ composition (at rather uniform $^{207}\text{Pb}/^{204}\text{Pb}$) at 100 Ma (Fig. 8). Such model could explain the compositional variation of the intra-plate magmatic rocks from depleted sources in the $^{207}\text{Pb}/^{204}\text{Pb}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ space by compositional heterogeneities within the lithospheric mantle (Fig. 8).

4.4. Implications for the Cenozoic Andean arc magmatism

4.4.1. Contributions of SCLM

The radiogenic isotope compositions of dominantly Cretaceous (but also some late Palaeozoic and Tertiary) intra-plate magmatic rocks indicate two mantle types, a depleted mantle and an enriched mantle. The isotopic composition of the enriched mantle is comparable with compositions known from Brazilian SCLM and similar to the EM1 endmember (Figs. 3b and 6a). The influence of melts from this mantle type is not evident in the isotopic compositions of magmatic rocks from the Andean arcs. Mesozoic to modern arc magmas follow a uniform trend

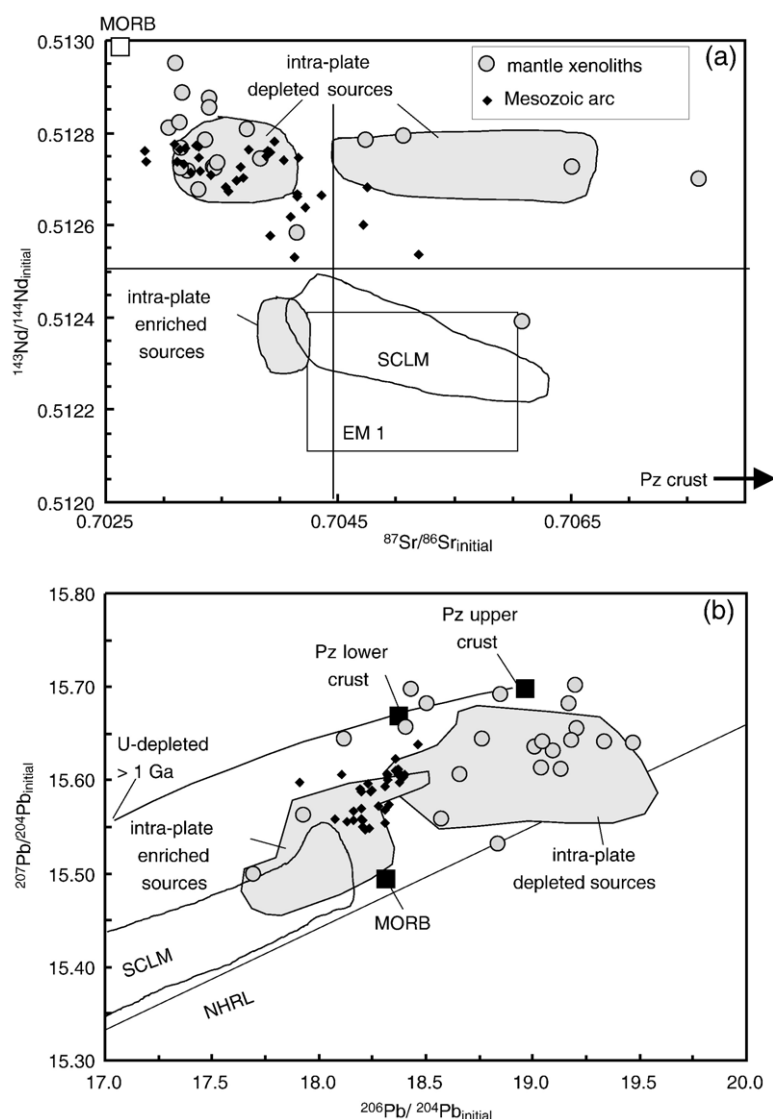


Fig. 7. (a) Sr–Nd isotope and (b) uraniumogenic Pb isotope compositions of Cretaceous lithospheric mantle xenoliths ($\sim 26^\circ\text{S}$; Las Conchas locations; Lucassen et al., 2005), Jurassic to early Cretaceous magmatic rocks from the arc ($18^\circ\text{--}27^\circ\text{S}$; Lucassen et al. 2002a; Kramer et al., 2005; Lucassen et al., 2006) compared with the depleted and enriched sources of the intra-plate magmatic rocks (grey shaded fields); upper (average) Palaeozoic and lower Palaeozoic crust ($21^\circ\text{--}27^\circ\text{S}$; Lucassen et al., 1999; Lucassen et al., 2002b), corrected for in-situ decay to 100 Ma. Other sources and endmember compositions see Figs. 3 and 6; U depleted source (e.g., Arequipa Massif of southern Peru; Tilton and Barreiro (1980)). The xenoliths show a similar range of Nd–Sr–Pb compositions as the depleted mantle source, as defined for the intra-plate magmatic rocks.

of crustal contamination in their Sr–Nd isotope composition (Fig. 9a). Their Pb isotope compositions follow either uniform mixing patterns in between a depleted mantle source and the common Palaeozoic crust or indicate Pb provinciality related to regional heterogeneities in the crustal composition (Fig. 9b; e.g., Macfarlane et al., 1990; Aitchison et al., 1995).

The occurrences of enriched mantle lithosphere could either represent Proterozoic remnants within mainly

Palaeozoic mantle (the Palaeozoic orogeny reworked mainly cratonic material) or mark the transition to the Brazilian Shield. The border between Proterozoic craton and the early Palaeozoic orogen is unknown from the surface geology. Recent seismic images of the lithosphere show that underthrust of the Brazilian Shield beneath the Cenozoic orogen does not necessarily occur at all locations (Heit, 2005 and references therein) and place the western edge of the Brazilian Shield at 21°S near

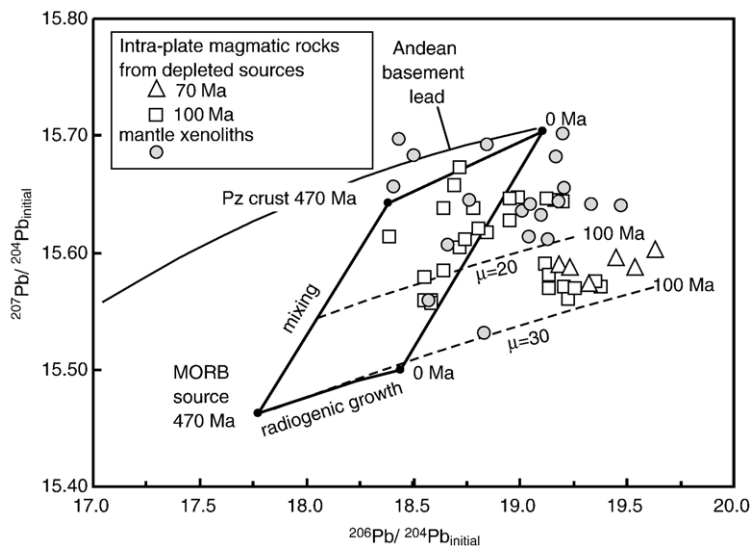


Fig. 8. Model for the evolution of the uraniumogenic Pb in the intra-plate magmatic rocks and many of the xenoliths. The Pacific MORB source is considered as a proxy of depleted mantle, the Mesozoic sub-arc mantle represents the ‘true’ asthenospheric beneath the arc; The Pacific MORB is corrected for radioactive in-situ decay using the present average Pb isotope composition of Pacific MORB and a μ ratio of 8.1. The average early Palaeozoic magmatic and metamorphic basement (Pz crust; 21°–27°S; Lucassen et al., 2001; 2002b) is a good approximation to a common crustal endmember in the Central and Southern Andes. Deviations from this scheme are less radiogenic than this crust at 470 Ma, e.g., old (>1 Ga) U-depleted rocks (e.g., Arequipa Massif of southern Peru; Tilton and Barreiro, 1980). The Pb isotope composition the Pz crust at 470 Ma is represented by Pb isotope composition of feldspar and may be also common in the U-depleted lower crust up to Present (Lucassen et al., 1999, 2002b). The lines labelled 20 μ and 30 μ represent 370 Ma of Pb growths starting from different initial Pb composition at 470 Ma.

longitude 65°W. This is where SCLM was sampled in late Palaeozoic at the San Andres location (Fig. 1b).

4.4.2. Depleted lithospheric mantle

The isotopic characteristics of the Mesozoic sub-arc mantle evolved consistently up to Recent and indicate a uniform composition of the mantle wedge (incoming ‘fresh’ asthenosphere plus element input from subduction) beneath this section of the Andean margin at least since Mesozoic (Fig. 9; 18°–40°S; recent summary in Lucassen et al., 2006). The only marked difference in the isotope composition between the sub-arc and depleted lithospheric mantle sources (with unradiogenic $^{87}\text{Sr}/^{86}\text{Sr}$) is the more radiogenic Pb isotope composition of the latter (Fig. 7). Contributions of the depleted lithospheric mantle during Cenozoic to Recent shortening in the Central Andes would not be seen in the Sr and Nd isotopic composition of arc magmas from such mixed mantle source (Fig. 9a). The Pb isotope signatures of the hybrid magmatic rocks from the compressive arc of the CVZ are dominated by crustal Pb (Fig. 9b), which likely superimpose possible Pb isotope signatures from mixing of low Pb mantle sources.

The most depleted Sr, Nd (and Pb) isotopic compositions of the Miocene alkaline La Peña samples

are indistinguishable from the Cretaceous intra-plate samples from depleted sources (Figs. 3a and 6a). Approximately 200 km north of the La Peña location, the Miocene Maquinas alkali basalts in the Andean back arc were investigated by Kay et al. (1991) and Kay and Abbruzzi (1996). The source of the basalts has been interpreted as lithospheric mantle. One of the samples has Nd and Sr isotope compositions identical to the least contaminated La Peña sample, but a Pb isotope composition similar to the most contaminated La Peña samples. The least contaminated sample from the Miocene Tupiza alkaline rocks in southern Bolivia also shows affinities to the depleted mantle sources and to the trace element patterns of the intra-plate magmatic rocks. Sources comparable to those of the Cretaceous intra-plate magmatism were still active in the Miocene back arc at distant localities (Fig. 1).

5. Summary and conclusion

The new isotope data on the lithospheric mantle in the Cretaceous back arc suggest the dominance of depleted compositions in the mantle lithosphere on a large regional scale in the Central Andes. Isotope signatures of our sample suite show no systematic mixing relations with

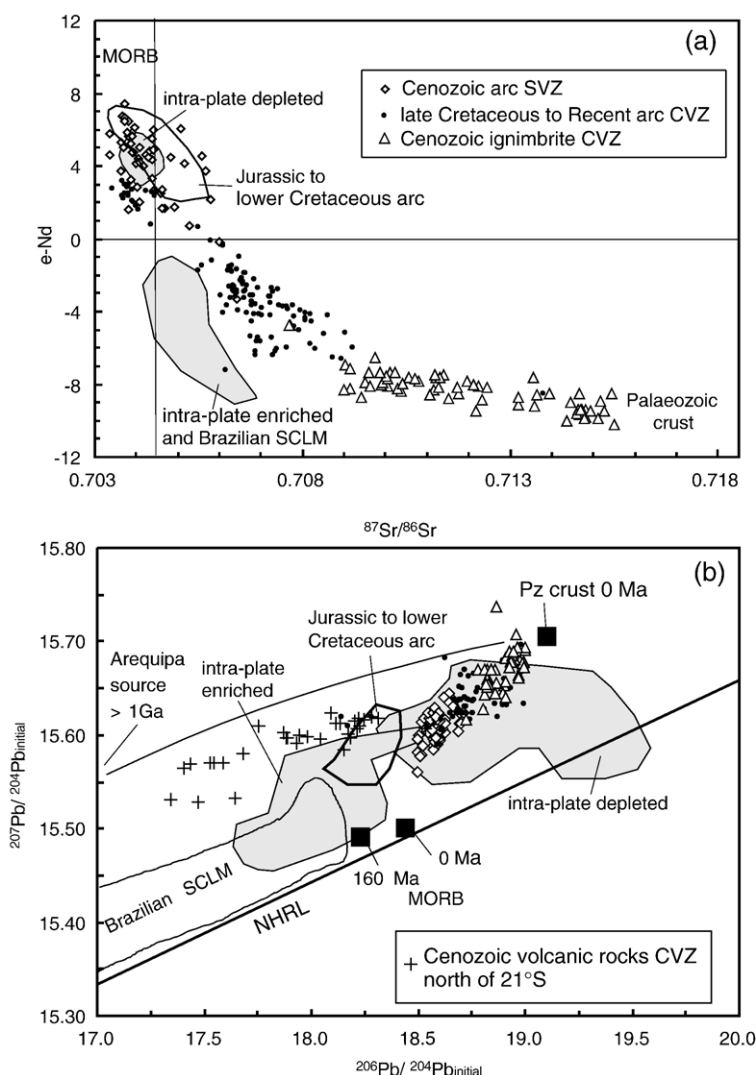


Fig. 9. Relations between the Nd, Sr, and Pb isotopic composition of mantle-derived and crust-contaminated (hybrid) magmatic rocks of the Mesozoic to Recent arcs (18° – 27°S ; CVZ and 34 – 40°S ; SVZ; Fig. 1b) and the composition of the intra-plate magmatism. (a) Nd–Sr isotopic compositions follow a uniform trend between a depleted mantle source, which is very similar in the CVZ and SVZ, and the ambient Palaeozoic crust. Major influence of an enriched mantle source is not seen. (b) Systematic differences between the Pb isotope composition of the Jurassic and late Cenozoic (age < 20 Ma) mantle-derived rocks could be explained by an evolving mantle source (comparable to the evolution of MORB between 160 to 0 Ma) plus crustal contribution in the subduction system. Andesite of the CVZ south of 21°S are intermediate in composition between mantle-derived rocks from SVZ (representing the modern sub-arc mantle composition) and ignimbrite from CVZ, which contain high proportions of melts from the Palaeozoic crust. Pb provinciality of the crust is known from the Pb isotope composition of Cenozoic andesite from CVZ north of 21°S (Aitchison et al., 1995), but not seen in the intra-plate igneous rocks which also comprise some locations north of 21°S . Data sources: magmatic rocks of the Jurassic to lower Cretaceous magmatic arcs (Lucassen et al., 2002a, Kramer et al., 2005; Lucassen et al., 2006); the late Cretaceous to early Cenozoic magmatic rocks and late Cenozoic andesite and ignimbrite from the CVZ (Rogers and Hawkesworth, 1989; Francis et al., 1989; Aitchison et al., 1995; Ort et al., 1996; Trumbull et al., 1999; Lindsay et al., 2001; Haschke et al., 2002); SVZ volcanic rocks (<http://georoc.mpch-mainz.gwdg.de/georoc/> and Lucassen et al., 2004 and references therein); other sources as in Figs. 3 and 6.

isotopic compositions of common mantle ‘endmembers’ (Figs. 3b and 6). The Cretaceous depleted lithospheric mantle presumably represents metasomatized sub-arc mantle of the early Palaeozoic arc. A contribution of

sub-continental asthenospheric mantle to the isotopic composition of our sample suite is either not visible due to compositional similarity (Mesozoic to Recent Andean sub-arc mantle composition) or very limited (MORB-type

source). Subduction could create regionally important mantle reservoirs *sui generis* in the upper mantle.

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Appendix A

A.1. Sample preparation, analytical methods, and data quality

Samples were first crushed using a hammer and, after separation of parts with visible alteration or xenoliths, passed through a jaw crusher. The crushed samples were cleaned in distilled water in order to remove the dust fraction. Most volcanic and sub-volcanic rock samples are fine grained and we selected ~100 g of fragments (~5–10 mm) under the binocular microscope, avoiding altered material as filled fractures, amygdalae, and weathered parts if present. Samples were ground in a tungsten-carbide shatterbox for 5 min. The loss on ignition was determined on ~2 g of dry (110 °C) powdered sample at 1000 °C. Whole-rock samples were analysed for major elements and some trace elements (Sr, Ba, Cr, Nb, Ni, V, Zr) by XRF (PHILIPS PW 1404; Tables 2a, 2b and 2c and A2) at Technical University Berlin on fused glass disks using Oxiquant and X-40 Philips software. Reproducibility of the in-house standard is ($n=90$; wt.% ± 1 SD): SiO₂: 60.1 $\pm$ 0.75; Al₂O₃: 19.0 $\pm$ 0.27; Fe₂O₃: 5.8 $\pm$ 0.11; MnO: 0.13 $\pm$ 0.004; MgO: 1.1 $\pm$ 0.13; CaO: 5.1 $\pm$ 0.06; Na₂O: 4.6 $\pm$ 0.23; K₂O: 2.2 $\pm$ 0.04; TiO₂: 0.56 $\pm$ 0.012; P₂O₅: 0.20 $\pm$ 0.008.

Powdered samples of silicate rocks (except carbonatite and alnöites, which contain primary carbonate) for isotope and ICP-MS analysis were treated with hot acetic acid or diluted (2 N) HCl to remove possible secondary carbonate. Sr contents bound to carbonate should be lowered by leaching and substantial Sr loss should be seen in the ICP-MS Sr contents in comparison to the values obtained from XRF on unleached samples. The difference between XRF and ICP-MS Sr-values is small in most samples (except Vichacla location, probably due to dissolution of some nepheline in HCl; Table A2) and not correlated with the Sr isotope composition. The effect of a previous leaching experiment on one of the high ⁸⁷Sr/⁸⁶Sr from the Las Conchas location on Sr isotope ratios is small and does not affect any conclusions (untreated sample: 0.706708, Sr 1160 ppm; leached sample 0.706566, Sr 1123 ppm; leachate: 0.711770; Nd and Pb isotope ratios vary within the analytical error; Lucassen et al., 2002a). The dried samples were weighed into Savillex® screw top beakers. Samples were dissolved in 52% HF on the hot plate at 180 °C for at least 36 h. After drying, the samples were redissolved in 6 N HCl and split into two fractions one for Pb isotope determination and one for Sr and Nd isotope analysis. Pb was separated using standard HBr–HCl chemistry; Sr and the REE fractions were separated in a first step from the same solution and Nd was isolated in a second step from the other REE using HDEHP coated Teflon® powder. Laboratory procedures at GFZ-Potsdam are described in Romer et al. (2005). Rare earth element (REE), U, Th, Pb, Rb, and Sr contents were determined by ICP-MS at GFZ-Potsdam (experimental procedure at GFZ-Potsdam and accuracy from international reference material see Dulski, 2001). The Pb isotope ratios of the Ayopaya carbonatite were determined on carbonate separates.

All isotope ratios were determined using TIMS. Nd and Sr isotope were measured using dynamic multi-collection on a MAT 262 mass spectrometer and on a VG 54 Sector mass spectrometer at the GFZ-Potsdam, respectively. Nd isotopic ratios were normalized to ¹⁴⁶Nd/¹⁴⁴Nd=0.7219, Sr isotopic ratios to ⁸⁶Sr/⁸⁸Sr=0.1194. During the analytical work at GFZ-Potsdam the NBS 987 Sr yielded ⁸⁷Sr/⁸⁶Sr=0.710265 $\pm$ 28 (2 σ ; $n=47$) and the La Jolla Nd standard yielded ¹⁴³Nd/¹⁴⁴Nd=0.511850 $\pm$ 8 (2 σ , $n=46$). Pb isotope ratios at GFZ-Potsdam were measured using static multi-collection on a MAT 262 mass spectrometer at controlled temperatures between 1220 and 1250 °C. The 2 σ reproducibility of all Pb isotope ratios of the NBS SRM 981 standard (²⁰⁶Pb/²⁰⁴Pb=16.902 $\pm$ 0.015, $n=58$) is better than 0.1% and a 2 σ error of 0.1% is assumed for the measured samples. Instrumental mass-fractionation

has been corrected using 0.1% per a.m.u. on the basis of the NBS SRM 981 values. Procedural blanks were <30 pg for Pb, <50 pg Nd and <100 pg for Sr. No blank corrections have been applied to the measured ratios because blank contribution was insignificant in comparison to the amount of the respective elements in the sample.

A.2. K–Ar dating

Minerals were separated magnetically and by handpicking under the binocular microscope. Purified biotite was ground in pure alcohol to remove altered rims that might have experienced a loss of Ar or K. Details of sample preparation, argon and potassium analyses for the laboratory in Göttingen are given in Wemmer (1991). Potassium was determined in duplicate by flame photometry using an Eppendorf Elex 63/61. The samples were dissolved in a mixture of HF and HNO₃ according to the technique of Heinrichs and Hermann (1990). CsCl and LiCl were added as an ionization buffer and internal standard, respectively.

The argon isotopic composition was measured in a pyrex glass extraction and purification line couple to a VG 1200 C noble gas mass spectrometer operating in static mode. The amount of radiogenic ⁴⁰Ar was determined by isotope dilution method using a highly enriched ³⁸Ar spike (Schuhmacher, 1975). The spike is calibrated against the biotite standard HD-B1 (Fuhrmann et al., 1987). The age calculations are based on the constants recommended by the IUGS quoted in Steiger and Jäger (1977). The stated analytical error for the K/Ar age calculations is given at the 95% confidence level (2σ; Table A1).

Appendix B. Supplementary data

Supplementary data associated with this article can be found, in the online version, at [doi:10.1016/j.lithos.2007.06.007](https://doi.org/10.1016/j.lithos.2007.06.007).

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