Crustal Evolution at the Central Andean Continental Margin: a Geochemical Record of Crustal Growth, Recycling and Destruction

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Abstract. Active continental margins are considered as the principal site for growth of the continental crust. However, they are also sites of recycling and destruction of continental crust. The Andean continental margin has been periodically active at least since the early Paleozoic and allows the evaluation of the long-term relevance of these processes. The early Paleozoic orogeny at ca. 0.5 Ga recycled and homogenized the ~2 Ga old early Proterozoic crust of the Brazilian Shield, which was previously orogenized at ca. 1 Ga, consistent with global models of prominent crustal growth at 2 Ga and a near constant mass of continental crust in the Phanerozoic. The metamorphic and magmatic evolution and the isotopic signatures of the early Paleozoic rocks do not indicate significant crustal growth, either by accretion of exotic terranes and island arcs or by juvenile additions from a mantle source. The dominant inferred mode of crustal evolution in the Paleozoic was recycling of older crust. Destruction of continental crust by subduction erosion is prominent in sections of the present active margin and is also likely to have occurred in the past orogens. Voluminous juvenile magmatism is only observed in the Jurassic – lower Cretaceous extensional magmatic arc. Compositions of mantle-derived magmas from the early Paleozoic to the Cainozoic, as well as late Cretaceous mantle xenoliths, indicate that depleted mantle was already present beneath the early Paleozoic orogen. The old subcontinental, enriched mantle related to the Brazilian shield and bordering Proterozoic mobile belts was modified by asthenospheric mantle in the younger subduction systems. In summary, this transect of the Andes is not a site of major continental growth, but a site where long-term processes of growth, recycling and destruction balance out.

3.1 Introduction

Growth of the continental crust is closely linked to the phenomenon of subduction, and active continental margins are generally considered to be the principal sites for the formation of continental crust. Ernst (2000) commented that 'the subduction process … appears to be responsible for the formation and growth of continents. … This constitutes the geologic regime where new continental crust is forming.' Similarly, Middlemost (1997)

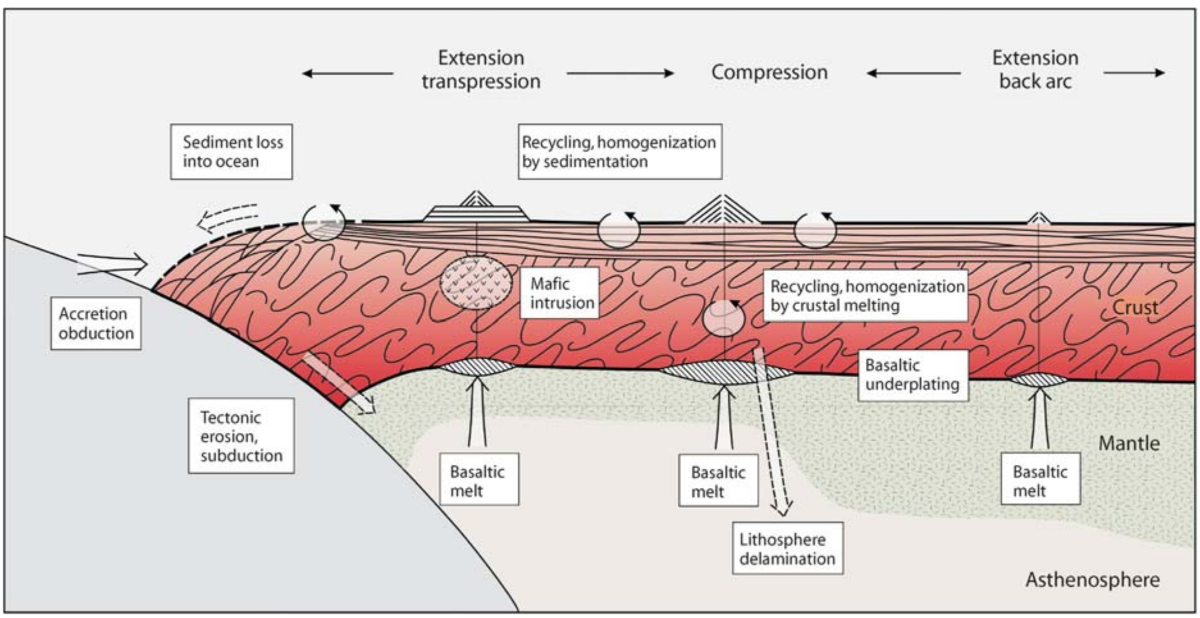

Fig. 3.1. Schematic compilation of the processes of crustal growth, recycling and destruction at an active continental margin; *circles* indicate recycling, *solid arrows* growth and *dashed arrows* destruction of continental crustal material. The transpressional regime is situated in the fore-arc, shown here together with the extensional regime for the sake of simplicity. Tectonic shortening, which leads to thickening of the crust and area reduction and may eventually lead to lithospheric delamination, is not shown

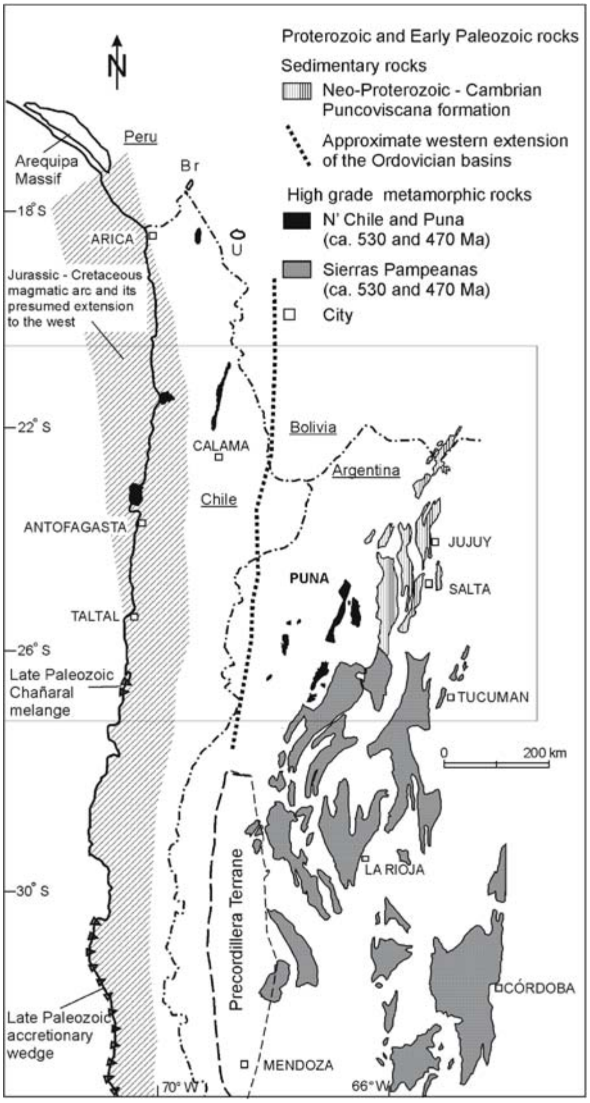

Fig. 3.2a. Distribution of Proterozoic (Arequipa Massif, *Br* = Berenguela, *U* = Uyarani; ca. 1.2–1.0 Ga) and early Paleozoic rocks between ca. ~16 and 33° S and the position of the Precordillera Terrane (modified from Lucassen et al. 2000); northern extension of rocks from the late Paleozoic accretionary wedge of southern and central Chile (Hervé 1988; Willner et al. 2004; this volume) and the late Paleozoic Chañeral Melange (Bell 1987); extension of the early Jurassic – Cretaceous magmatic arc (presumed extension to the west: Von Hillebrandt et al. 2000). The *box* shows the approximate area covered by Fig. 3.2b and by the detailed geological map (see Reutter and Munier 2006, Chap. 27 of this volume; Schnurr et al. 2006, Chap. 29 of this volume)

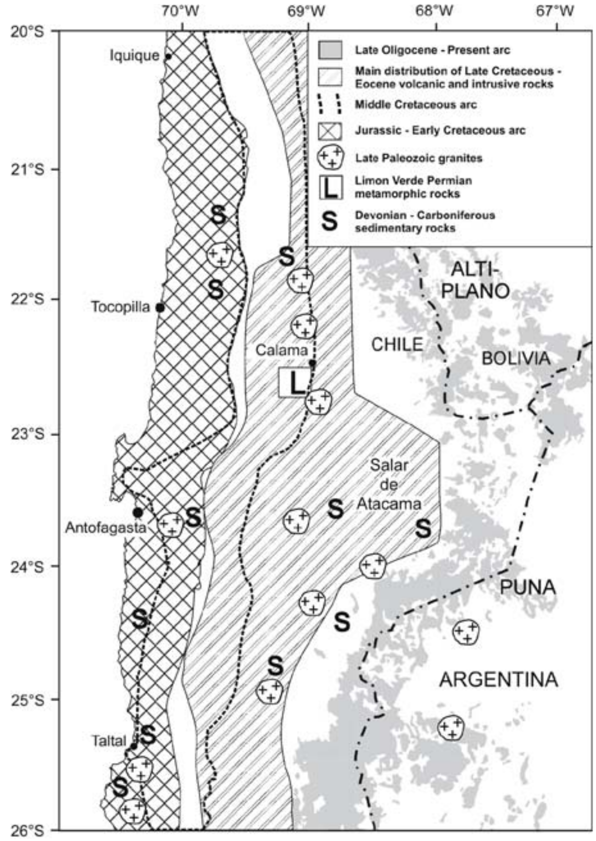

Fig. 3.2b. Late Paleozoic – Cainozoic magmatic rocks from the various magmatic arcs intruded or overlay early Paleozoic crust. Devonian – Carboniferous sedimentary rocks frequently form the host rocks of the late Paleozoic intrusions. The Permian high-grade metamorphic rocks of Limon Verde south of Calama are unique in the area. Map modified after Reutter et al. 1994; E. Scheuber, Freie Universität Berlin, pers. comm.

states for continental margins 'It is also where new continental crust evolves and the process of subduction inserts new chemical heterogeneities and complexities into the upper mantle.'

In a subduction environment, possible processes of continental crustal growth include the addition of magma from the mantle to the crust in volcanic arcs, the tectonic accretion of sedimentary material from the subducting to the overriding plate, obduction of oceanic crust from the lower plate, and the collision with the continental margin of continental or island arc terranes transported on the subducting plate (Fig. 3.1). From a global perspective, only the addition of mantle-derived magma and the accretion of oceanic crust can be strictly considered as net continental growth, because the other processes only redistribute continental material.

Active continental margins are also potential sites for the destruction of continental crust and lithosphere, both on a local and a global scale. Surface area reduction of continental crust (which we consider as the local perspective) results when the crust is thickened by tectonic shortening (homogeneous shortening, overthrusting and folding). Volume reduction of crust and associated lithosphere (the global perspective) can occur because of density-driven delamination of mafic material from the base of the crust as a result of shortening (Fig. 3.1). The latter

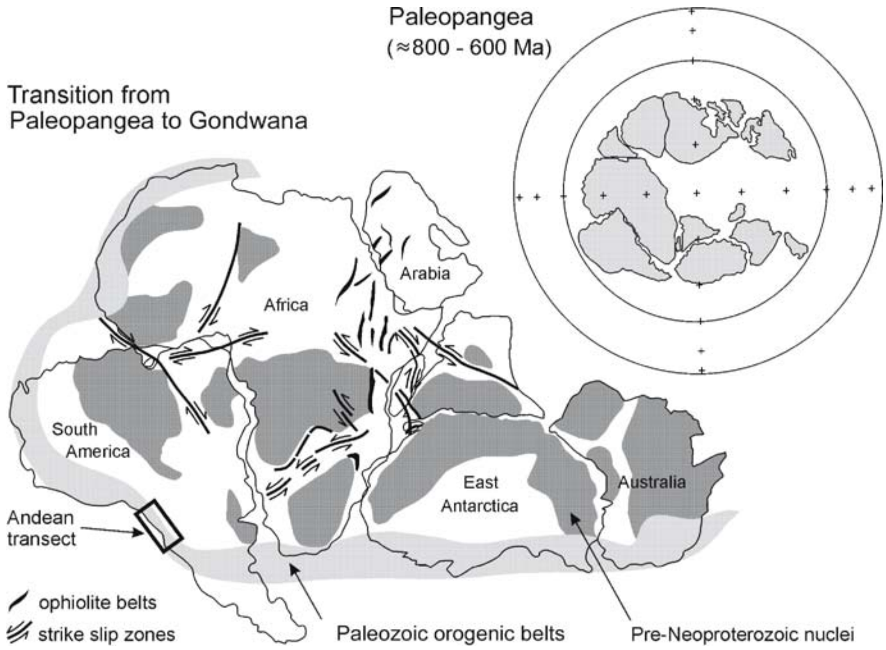

Fig. 3.3.
Paleogeographic reconstructions for Paleopangea and Gondwana (modified from Piper 2000), showing the position of the Andean transect as part of the Paleozoic orogenic belts at the continental margin

process recycles crust (together with mantle lithosphere) into the deeper mantle.

Finally, the destruction of continental crust at active margins can also occur by the subduction of crustal material into the mantle, either as continental sediments deposited in the trench and/or as tectonically eroded crustal rocks from the lip of the overriding plate. Clift and Vanucchi (2004), among others, pointed out that tectonic erosion at continental margins is a major factor in the destruction of continental crust. These processes will leave traces in both the crust and the mantle lithosphere. Most evidence is preserved in the continental crust and revealed through the composition, ages and conditions of formation of metamorphic, magmatic and sedimentary rocks. The evolution of the upper mantle is less tractable from geologic studies, but is an important factor for the formation and destruction of continental material. In this paper we review the available evidence for crustal and mantle evolution in the Central Andes.

Since the first plate tectonic models were developed, the Central Andes have been considered as the type example of an active continental margin (e.g. Mitchell and Reading 1969), and its over-thickened crust and high plateau are the typical example for 'Cordilleran-type orogeny' (Fig. 3.2). The Andean orogeny dominates the present form of this continental margin and its most prominent features (volcanic arc and plateau), but, in fact, the Central Andean crust records at least 600 Ma of history as the leading edge of the South American continent (Table 3.1; Fig. 3.3). It also preserves relics from the 2 Ga history of the western part of the South American craton. This long history is documented by the ages of magmatism and metamorphism, by the depositional ages of Phanerozoic sedimentary rocks (Table 3.1), and by the chemical and isotopic composition of the diverse crustal rocks (Figs. 3.4 to 3.7). The early Paleozoic and Late Paleozoic to Cainozoic history was dominated by prolonged periods of subduction. The mantle evolution of the margin is less well constrained due to the discontinuous record of mantle-derived magmatic rocks, and the fact that mantle xenoliths are essentially absent in arc magmas. Nevertheless, there are enough data to provide some useful constraints on mantle composition.

The purpose of this paper is to summarize and discuss crustal and mantle features of the Central Andes between 21° S and 27° S (as the core area, but extending to 18° S and 32° S) as a case study of the development of an active continental margin. We evaluate how representative the presently active state of the system is regarding long-term evolution, particularly with respect to crustal growth or destruction. The discussion is based on information about the geological history of the area as well as on a large volume of chemical and isotopic data from the continental crust and mantle-derived rocks. Much of the data reviewed in this paper were acquired between 1992 and 2004 in the SFB 267 research program 'Deformation processes in the Andes' and between 1988 and 1991 in an earlier program 'Mobility of active continental margins'. Our work focused on the Phanerozoic crystalline basement and on the chemical-petrological characteristics of Cainozoic magmas. A further aspect was the nature of the upper mantle and its changes with the evolution of the Central Andes. As a large data set has been built up in these years, the present paper can only give a brief over-

Table 3.1. Evolution of the continental crust

Time (Ma)	Evolution of the continental crust
30–Recent	Area of the Early Paleozoic Mobile Belt: Formation of the Puna–Altiplano Plateau. Cenozoic Andean andesite and ignimbrite volcanism sample the Early Paleozoic crust. Strong arc normal shortening and tectonic thickening crust (for references see the text).
<120	Area of the Early Paleozoic Mobile Belt: Eastward movement of the magmatic arcs (1, 2). Possible subduction erosion of the Late Paleozoic, Triassic and Jurassic magmatic arc rocks and Early Paleozoic crust (2 and text). Mainly transpression – transtension (1, 2).
200–120 Start of the 'Andean cycle'	Area of the Early Paleozoic Mobile Belt: Formation of juvenile 'magmatic' crust in the Jurassic–Lower Cretaceous arc. Direct evidence for crustal growth (for references see text). Phases of arc normal extension and transtension and transpression (2).
330–200	Area of the Early Paleozoic Mobile Belt: Widespread granitoid intrusions mainly in N Chile (ca. 300–220 Ma; 4, 5, 6) and an extended Permian rhyolite province (7, 8) recycle Early Paleozoic crust; Late Triassic transition to mantle derived melts in the arc (9, 10) high grade metamorphic rocks of Sierra de Limón Verde (270 Ma; 11). Extension to transpression.
400–330	Early Paleozoic Mobile Belt: Passive margin evolution (12, 13) with sedimentary rocks recycling Early Paleozoic crust (14). Exhumation of the metamorphic basement largely finished. K-Ar cooling ages of metamorphic rocks from micas and first erosional unconformities (15, 16).
500–400 Local name 'Famatinian cycle'	Early Paleozoic Mobile Belt: Formation of the mobile belt; high T/moderate P metamorphism (470–400 Ma; 16, 17, 18, 19, 20, 21, 22) and intense crustal derived magmatism mainly recycles Proterozoic crust (490–400 Ma; 17, 18, 23, 24, 25, 30); oldest K-Ar cooling ages from hornblende in metamorphic rocks (15, 16). Ordovician sediments recycle the Early Paleozoic and Proterozoic crust (14, 26, 27).
ca. (900) 600–500 (400) Panafrican or Brasiliano cycle (Local name: Pampean cycle)	Early Paleozoic Mobile Belt: Formation of the mobile belt; Neoproterozoic–Eocambrian Sediments (28, 29) recycle Proterozoic basement (14, 25, 29) and form part of the protoliths of the Early Paleozoic basement. High T/moderate P metamorphism (530–500 Ma; 16, 17, 18, 19, 20, 21, 22) and crustal derived magmatism recycles Proterozoic crust (17, 18, 23, 24, 25, 30); South American craton: Formation of mobile belts between 700–500 Ma; high T metamorphism and associated granitoid magmatism (31, 32, 33).
ca. 1 300–900 Sunsás cycle (Contemporaneous with the Grenvillian cycle of North America)	South American craton: Major metamorphic–magmatic cycle with mobile belts in-between older cratonic regions (31, 33, 34); granulite facies metamorphism and granitoid magmatism in the Arequipa Massif (Peru) and Bolivia (19, 36, 37, 38). Early Paleozoic Mobile Belt: Upper intercept ages of inherited zircons in the Early Paleozoic crust (17, 18, 19; 38).
ca. 2 000 Early Proterozoic	South American craton: First order crust-formation period at ca. 2 Ga around and west of the Paleoproterozoic to Archean parts of South America; (33, 39, 40). Early Paleozoic Mobile Belt: Nd model ages (6, 14, 16, 23, 25, 26) and upper intercept U-Pb ages of inherited zircons ca. 1.6–2 Ga (8, 19;29, 38).

(1) Haschke et al. 2002; (2) Scheuber et al. 1994; (3) Scheuber and Gonzales 1999; (4) Berg and Baumann 1985; (5) Brown 1991 (6) Lucassen et al. 1999a (7) Kay et al. 1989; (8) Breitkreuz and van Schmus 1996; (9) Morata et al. 2000; (10) Bartsch 2004; (11) Lucassen et al. 1999b; (12) Bahlburg and Hervé 1997; (13) Augustsson and Bahlburg 2003; (14) Bock et al. 2000; (15) Becchio et al. 1999; (16) compilation in Lucassen et al. 2000; (17) Damm et al. 1990; (18) Damm et al. 1994; (19) Wörner et al. 2000; (20) Lucassen and Becchio 2003; (21) Höckenreiner et al. 2003; (22) Büttner et al. 2005; (23) in: Pankhurst and Rapela (eds) 1998; (24) Pankhurst et al. 2000; (25) Lucassen et al. 2001; (26) Egenhoff and Lucassen 2003; (27) Zimmermann and Bahlburg 2003; (28) Aceñolaza et al. 1999; (29) Schwartz and Gromet 2004; (30) Lucassen et al. 2002a; (31) De Brito Neves and Cordani 1991; (32) Trompette 1997; (33) in: Cordani et al. 2000a; (34) Teixeira et al. 1989; (35) Litherland et al. 1989; (36) Wasteneys et al. 1995; (37) Martignole and Martelat 2003 (38) Loewy et al. 2004; (39) Sato and Siga 2002; (40) Rino et al. 2004.

view of the most important features. The interested reader is referred to the previous papers where details from individual studies are published.

The main part of this paper, therefore, summarizes the key geologic and petrologic features of the crust and upper mantle for this transect of the Central Andes, from the earliest recorded period to the present Andean cycle. We place strong emphasis on the isotope systems Rb-Sr, Sm-Nd and U-Th-Pb because they give the most direct information on mantle versus crustal provenance of the rocks concerned. We will show that the history of the Andean active margin, although it has certainly been a major site of geological activity, did not result in significant net crustal growth. Our perspective is local in a sense, but, because the Central Andes region has been an active continental margin through most of the Phanerozoic, we believe that the long history of crustal development recorded provides useful insights for the global perspective as well.

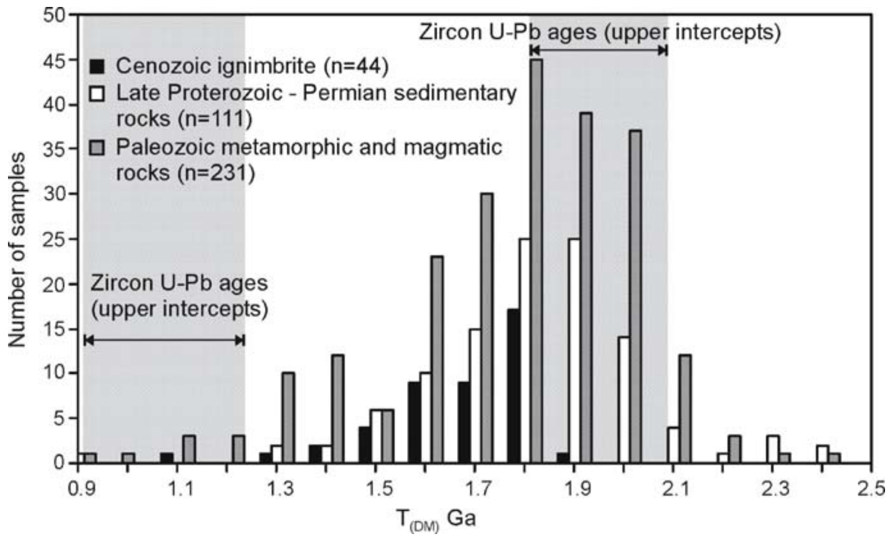

Fig. 3.4. Sm-Nd model ages of metamorphic, magmatic and sedimentary rocks of the Central Andes between 21° S and 32° S (single stage evolution model; Goldstein et al. 1984), compared to the range of U-Pb ages from zircon. The Sm-Nd data ($n = 342$; 4 samples outside the range) indicate the major crustal growth period in the early to middle Proterozoic (see text); the zircon ages indicate two distinct ages of crystallization. The Nd signature is also typical for Paleozoic rocks of the Coastal and Main Cordillera between 36 and 41° S (Lucassen et al. 2004). The Nd isotope signature of the large-scale Cainozoic ignimbrite resembles the isotope signature of the older crust. Figure modified from Lucassen et al. (2001); data south of 27° S from Pankhurst and Rapela (1998) and unpublished data of F. Lucassen and R. Becchio. The range of zircon ages includes the data of Damm et al. (1990, 1994), Wörner et al. (2000), Loewy et al. (2004, and references therein)

3.2 Proterozoic Development of the Central Andes

There is no record of Archean crust in the Central Andes and the earliest information comes from the distribution of Nd model ages of Phanerozoic igneous and metamorphic rocks (T_{DM}; Fig. 3.4). This shows a well-defined peak at 1.8–2.0 Ga, and a shoulder towards ages of 1.3 Ga. Model ages older and younger than this range are scarce. The material reworked in the Phanerozoic orogenies formed during a major period of crustal growth between 1.6 and 2.0 Ga, which is evident in many areas of the South American craton and common west of the Early Proterozoic and Archean cores (e.g., Goldstein et al. 1997; Cordani et al. 2000b; Sato and Siga 2002). Similar age ranges (2.2–1.9 Ga and 1.5–1.0 Ga) were also identified from analysis of detrital zircon populations from the Amazon (Rino et al. 2004) and Orinoco (Goldstein et al. 1997) river sediments, which – in contrast to the Phanerozoic rocks at the western edge of the continent – show the whole spectrum of ages including the Earliest Proterozoic and Archean ages (peak at 2.6 to 2.8 Ga and tail to Archean ages > 3 Ga) of the South American craton.

The major orogenic cycle during the Mesoproterozoic in the western section of the continent was the Sunsás orogeny at 0.9–1.3 Ga. Within the transect described in this paper (Fig. 3.2), metamorphic or magmatic crystallization ages corresponding to the Sunsás orogeny are absent, but such rocks do occur immediately to the north in the Arequipa Massif (southern Peru) and in a few small isolated outcrops in Bolivia (Table 3.1; Fig. 3.2a). Their ages are similar to those in the Sunsás Belt, which is located east of the Bolivian Andes (Litherland et al. 1989) and in the Andes of Columbia (Restrepo-Pace et al. 1997). The peak metamorphism is of high to ultra-high temperature type, the main lithologies described are prevailingly felsic, and protolith ages are mainly Paleoproterozoic to Mesoproterozoic (e.g. Wasteneys et al. 1995; Restrepo-Pace et al. 1997; Wörner et al. 2000; Martignole and Martelat 2003). The distribution of T_{DM} model ages indicates that juvenile additions to the crust in this period are of minor importance compared with the older episodes (Fig. 3.4) and we conclude, that if the Sunsas orogeny was important in the transect, it did mainly rework ca. 2.0 Ga old basement.

Evidence for inherited material of ca. 1–2 Ga of age is common in the Paleozoic and younger magmatic, metamorphic and sedimentary rocks in the Central Andes. The inherited material is evident in the upper intercepts of zircon U-Pb discordia arrays, as cores of single zircons revealed by in-situ dating methods, or as Sm-Nd average crustal residence ages described above (see also Fig. 3.4; Table 3.1). Existing models for Paleo- to Mesoproterozoic supercycles of crustal evolution in South America (Wasteneys et al. 1995; Goldstein et al. 1997; Restrepo-Pace et al. 1997; Wörner et al. 2000; Martignole and Martelat 2003; Condie 2002; Rino et al. 2004) are in good agreement with these Proterozoic ages for crustal formation and metamorphism in the Central Andes transect. Their similarity with those of the North American Grenvillian Belt, and the fact that their occurrence extends considerably N-S along the Andes, led to speculation of a possible collision of North and South America in Grenvillian/Sunsás time (e.g. Restrepo-Pace et al. 1997, and references therein). We do not consider the Proterozoic evolution further in this paper since key areas for Proterozoic crust occur outside the transect and the basement rocks within the transect have been completely reworked in the Paleozoic.

Little is known about the crustal evolution of our transect area for the time between the Sunsás orogeny and the early Paleozoic, which is described in the following section. The oldest sedimentary rocks in this age range are Neoproterozoic to early Cambrian (e.g. Puncoviscana Formation; Aceñolaza et al. 1999). Paleogeographic reconstructions (e.g. Unrug et al. 1996; Piper 2000) and geotectonic interpretations (e.g. Astini et al. 1995; Aceñolaza et al. 2002) suggest that the section of the margin considered here was located at the western edge of the South American Plate from the late Proterozoic onwards (Fig. 3.3).

Evidence for the presence – at least of remnants – of Proterozoic mantle lithosphere in the region is found in some alkaline rocks along the Cretaceous rift (see below) in the present back-arc region south of 27° S and north of 21° S (Lucassen et al. 2002b, 2005). Their isotopic characteristics are typical of old enriched mantle lithosphere, with $^{87}Sr/^{86}Sr$ initial ratios between 0.7040 and 0.7065, ε_{Nd} values between –1 and –9, and relatively unradiogenic Pb-isotope ratios. Similar isotope compositions of alkaline basaltic magmas are known from the cratonic parts of South America, e.g. the Paraná province of the Brazilian shield (e.g. Carlson et al. 1996; Gibson et al. 1996, 1999). In the region of 21 to 27° S, however, there is no evidence for such an enriched lithospheric mantle in the rift magmas or their xenoliths. If enriched mantle was once present in this region, it has apparently been replaced or overprinted by the depleted mantle that now dominates the region (see below).

3.3 The Early Paleozoic Orogeny

Early Paleozoic rocks, both sedimentary and metamorphic, cover large areas of southern Bolivia and northwestern Argentina but are scarce in northern Chile (Fig. 3.2). The Neoproterozoic – early Cambrian sedimentary rocks partly form the protoliths of the high-temperature metamorphic rocks in northwestern Argentina (Willner et al. 1985; Lucassen et al. 2001; Schwartz and Gromet 2004). Studies of these rocks over an area between ~18 and 32° S (5×10^5 km^2) show that the crust is surprisingly homogeneous in lithology and composition (with some very local exceptions). The dominant rock types in the high-grade basement of northwestern Argentina and northern Chile are felsic gneisses and migmatites of upper amphibolite facies, locally transitional to granulite facies.

Uniform pressure estimates between 5 and 7 kbar in the region indicate uniform mid-crustal exposure levels. High-pressure (high-P) rocks indicating exhumation from the root of the orogen, or those typical of accretionary complexes, seem to be absent. The uniformity of high-temperature (high-T) metamorphic conditions and absence of high-P rocks indicate large-scale uniform crustal thickening and possible plateau formation similar to the present Andean orogen (Lucassen and Franz 2005b).

Ages of metamorphic crystallization (Table 3.1) cluster in two groups, 530–500 Ma (Pampean) and ~470–400 Ma (Famatinian). Granitoid magmatism is widespread and voluminous, especially in the early Ordovician (560–400 Ma, see Table 3.1) whereas mafic igneous rocks are rare. The apparent grouping of metamorphic crystallization ages could be affected by the small size of the database. A continuous high temperature regime in the crust is indicated by voluminous and widespread granitoid magmatism (560–400 Ma, see Table 3.1), especially in the Early Ordovician (500–470 Ma).

Magmatic rocks occur in the same areas as the high-T metamorphism (Fig. 3.2). Most magmatic rocks are granitoid intrusions, and their chemical composition indicates various depths of melt generation and hybridization in the lower to mid crust (e.g. Damm et al. 1994; Coira et al. 1999; Pankhurst and Rapela 1998; Pankhurst et al. 2000). All melts show considerable contribution of crustal material (see below) or are crustal melts compositionally related to the metasediments. The composition of scarce early Paleozoic mafic intrusions in the plateau area, and west of it, resembles rocks from arc magmatism (Damm et al. 1990; Coira et al. 1999; Kleine et al. 2004; Zimmermann and Bahlburg 2003).

Detritus from the evolving early Paleozoic orogen is partly preserved in extensive Ordovician siliciclastic sediments of southern Bolivia (e.g. Egenhoff and Lucassen 2003, and references therein), which extend into northwestern Argentina. The Argentine Ordovician sequences contain some volcaniclastic rocks (e.g. Bock et al. 2000; Zimmermann and Bahlburg 2003, and references therein), but they are volumetrically unimportant and, in any case, do not represent significant mafic, potentially juvenile magmatism during this time. In northern Chile and the northern section of the Sierras Pampeanas, uplift and erosion of the orogen was completed in the Devonian (Lucassen et al. 2000, and references therein). In the southern section of the Sierras Pampeanas erosional unconformities occur in the Carboniferous (e.g. Ramos 2000).

The chemical (Fig. 3.5) and Nd-Sr isotope compositions (Fig. 3.6a) of various early Paleozoic sedimentary, felsic igneous and metamorphic rocks show similar compositional features for the different lithological or age groups in the region of 21 to 27° S. Although the metamorphic conditions indicate a deep middle crust, the rock

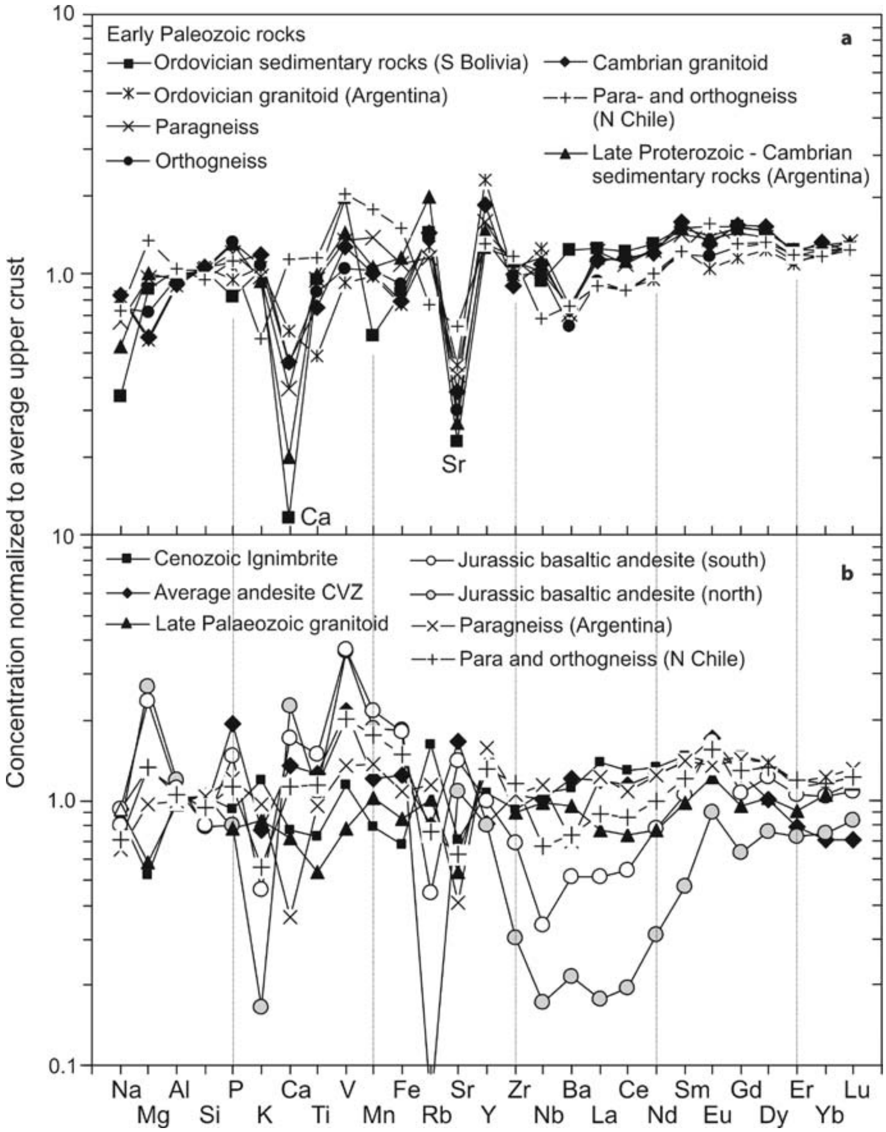

Fig. 3.5. Average major and trace element composition of rock units of the Central Andes from 21–27° S (normalized to average upper crust; Taylor and McLennan 1995). **a** The early Paleozoic crust has a uniform felsic composition similar to average upper-continental crust, except for Ca, Na and Sr. Early Paleozoic gneisses from northern Chile show a different chemical composition compared with the early Paleozoic rocks from Argentina and Bolivia, indicating a slightly less evolved protolith for the Chilean rocks. **b** Late Paleozoic granitoid and Mesozoic – Recent rocks of the various magmatic arcs and, for comparison, the gneisses of northern Chile and the paragneisses of Argentina. The Jurassic rocks, which are mantle derived, differ strongly from average upper crust, late Paleozoic granitoids, and typical rocks of the Cainozoic arc. Large-volume Cainozoic ignimbrites are compositionally similar to the Argentine early Paleozoic crust. Data Sources: Early and late Paleozoic rocks (Lucassen et al. 1999a; Lucassen et al. 2001; Egenhoff and Lucassen 2003), Jurassic magmatic arc (Lucassen et al. 2002b, Kramer et al. 2004), Cainozoic large-volume ignimbrite (Lindsay et al. 2001), average Central Volcanic Zone (CVZ) andesite (GEOROC database)

compositions are close to average upper-crustal values of Taylor and McLennan (1995). The main compositional differences from average upper crust are the lower Na, Ca and Sr contents in the rocks from the Central Andes transect, which may indicate a lower content of plagioclase (Eu is also slightly depleted compared to upper crust). The depletions in Sr, Eu and Ca relative to average upper crust, along with the slightly higher Rb and rare earth element (REE) contents suggest that the Central Andean Paleozoic crust was already chemically evolved. As discussed below, the isotopic and trace element data generally show compositional features typical for an old continental crust (Bock et al. 2000; Lucassen et al. 2001, 2002a; Egenhoff and Lucassen 2003).

The inherited age pattern of this crust indicates several metamorphic-magmatic-sedimentary cycles. The result of these processes of homogenization is well displayed by the uranogenic Pb isotope compositions of the igneous rocks. The Pb isotope ratios indicate a progressive homogenization over time of the initially varied Pb isotope compositions during the early Paleozoic, leading finally to a rather uniform Pb isotope composition of the crust (Fig. 3.7a). Compositional variations in the uranogenic Pb isotopes can be explained by long-term separation of crustal sections which formed earlier, e.g. in the Sunsás or older metamorphic-magmatic events.

In summary, crustal evolution during the early Paleozoic orogeny appears to have involved primarily reworking or recycling of pre-existing, ~2 Ga old felsic crust. Juvenile additions of magma from the mantle in this time are small. Extensive production of crustal melts during high-T regional metamorphism contributed to compositional homogenization on a regional scale, especially for Pb isotopes but probably also for the Sr isotope systems

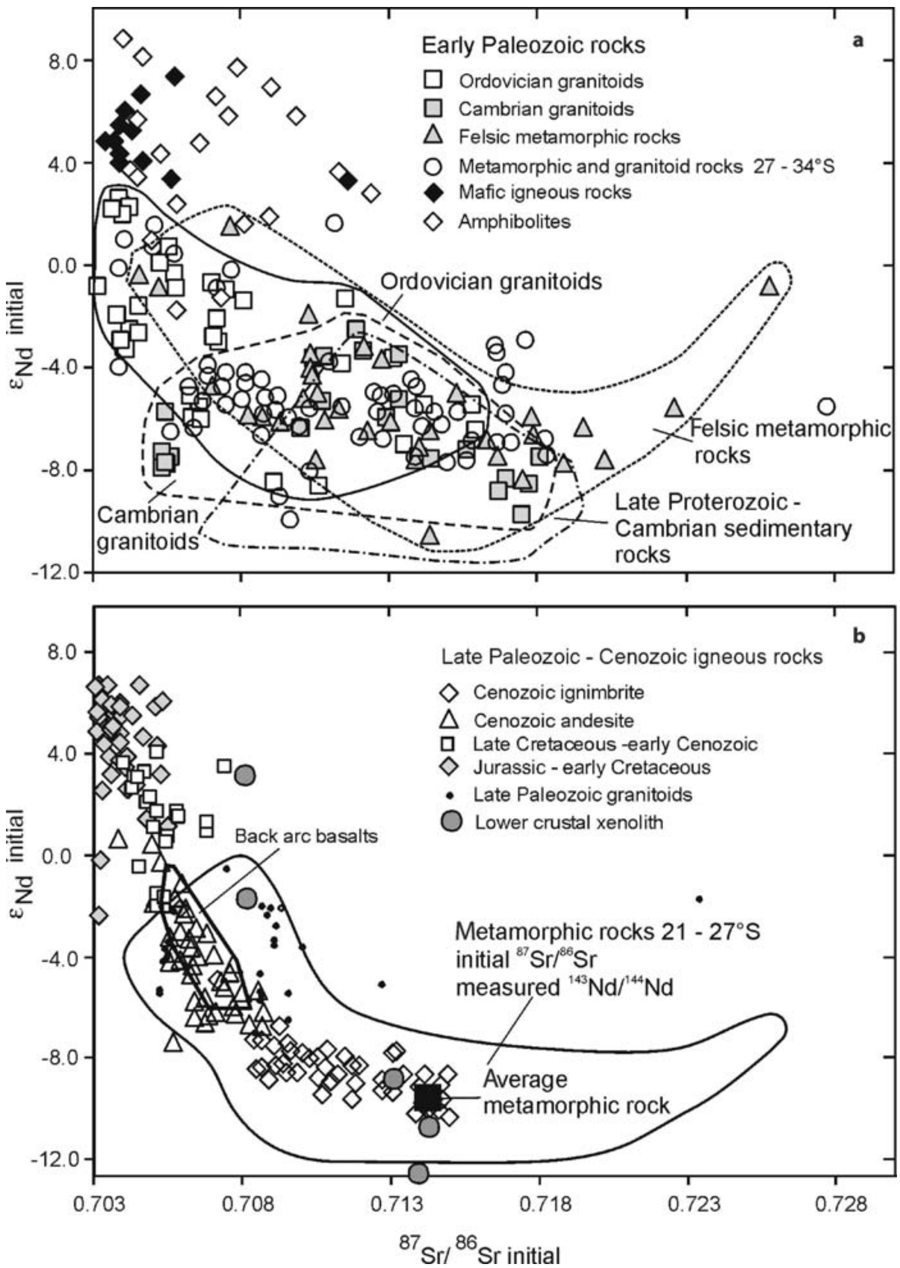

Fig. 3.6. Initial Nd-Sr isotope composition of major rock units of the Central Andes from 21 to 27° S. **a** Isotope ratios of rock groups of different age from Bolivia, Chile and Argentina of the early Paleozoic felsic magmatic and metamorphic basement, compared with the basement between 27–34° S (shown as a single group), and mafic magmatic rocks and amphibolite. The two regional groups show a very similar compositional range. The magmatic rocks are either crustal melts or a hybrid containing depleted mantle and felsic crust. The mafic rocks represent the composition of the mantle below the early Paleozoic arc, which is of the depleted mantle type. The Sr isotope ratios of many amphibolites and some mafic igneous rocks scatter to values of the felsic crust caused by Rb and Sr mobility during metamorphism, but the Sm-Nd composition of most rocks does not seem to be affected. **b** Late Paleozoic – Recent rocks show similar compositional trends between a depleted sub-arc mantle (represented by the Jurassic – early Cretaceous igneous rocks) and the early Paleozoic felsic crust. The possible isotopic composition of the Paleozoic lower crust is indicated by lower-crustal xenoliths, which plot into the field of the initial isotope ratios of the early Paleozoic rocks. The isotopic composition of the back-arc basalts is similar to the composition of the andesites. Data sources: (**a**) modified from Lucassen et al. (2001), additional data from Pankhurst and Rapela (1998), Egenhoff and Lucassen (2003), Kleine et al. (2004), and F. Lucassen and R. Becchio (unpublished data). (**b**) Rogers and Hawkesworth (1989), Kay et al. (1994), Trumbull et al. (1999), Lindsay et al. (2000, and references therein), Lucassen et al. (1999c, 2001, 2002b, and references therein), Haschke et al. (2002), Kramer et al. (2004) and Bartsch (2004)

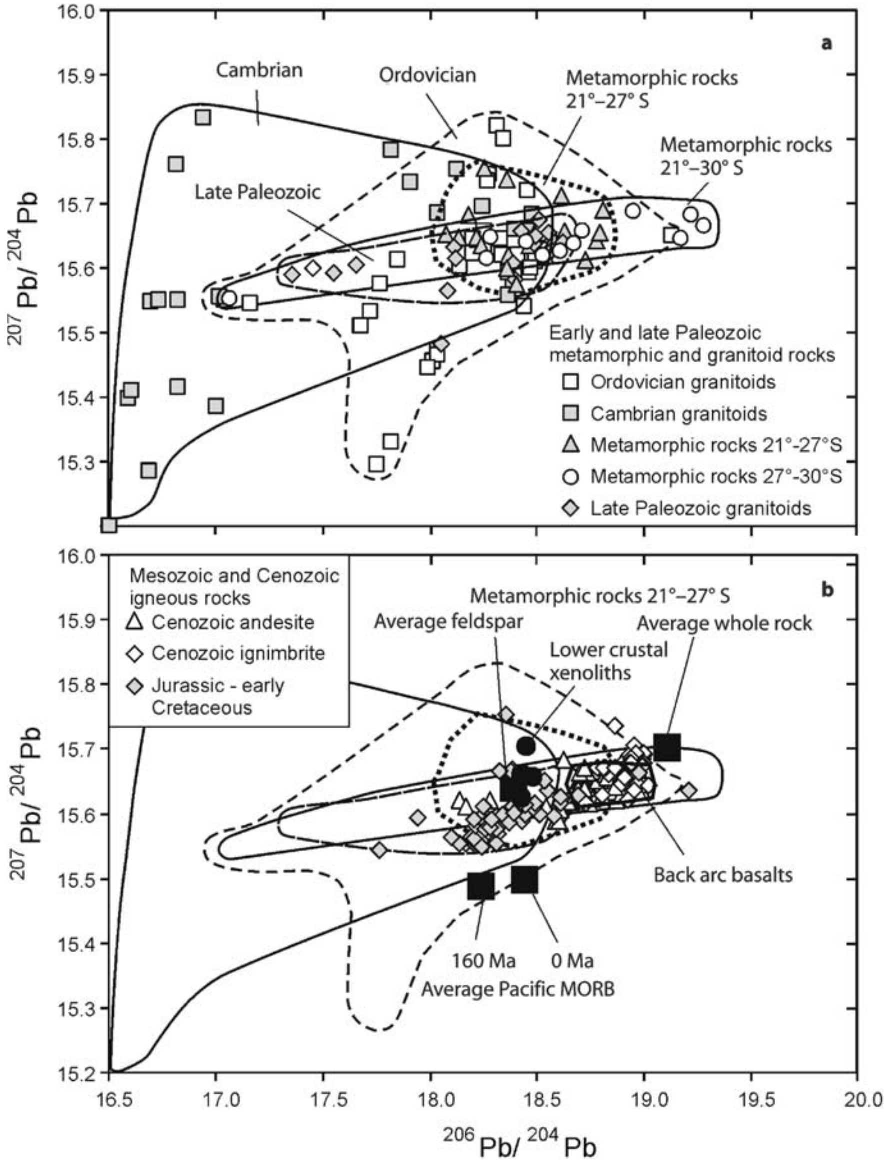

Fig. 3.7.
Uranogenic Pb initial isotope composition (**a**) Paleozoic rocks and (**b**) Mesozoic – Recent rocks. **a** All data shown are from feldspar. The Pb isotopic heterogeneities seen in the Cambrian and Ordovician rocks were homogenized during Paleozoic high-grade metamorphism and crustal melting. **b** The present-day Pb isotope composition of the crust is bracketed by the average feldspar composition of the metamorphic rocks and the lower-crust xenoliths, both representing the U-depleted lower crust and the average whole-rock composition representing the undepleted upper crust. The isotopic composition of andesite and ignimbrite represents this variation between these lower- and upper-crustal sources. Jurassic rocks show some affinity to the average Pacific MORB, which is the best available proxy for the Pb isotope composition of the sub-arc mantle. Modified from Lucassen et al. (2002a). Other data sources (see Fig. 3.6b). Average Pacific MORB from the data compilation in Lucassen et al. (2002b)

(Figs. 3.6a and 3.7a). Crystallization ages of the high-grade metamorphic and magmatic rocks indicate a long-standing regime (> 100 Ma) of high temperature (> 600 °C) in the mid-crust over a very large area. With the present data set, no distinctive thermal peaks or spatial patterns in metamorphic conditions can be distinguished. Thermal anomalies of this size and duration (i.e. equal to the lithospheric time constant) indicate a long-lasting and deep-seated origin, probably a shallow asthenosphere related to the active mantle wedge of a subduction zone.

Information about the mantle component involved in the early Paleozoic orogeny comes from meta-igneous mafic rocks, which are small in volume but not uncommon. Compositionally, the rocks are subalkaline basalt to basaltic andesite (Lucassen et al. 2001, and references therein). Ages of this magmatism are unknown but many of the metabasic rocks represent former dikes and their structural position with respect to the regional metamorphic fabric indicates an early Paleozoic emplacement. The principal mantle component in these rocks was compositionally very different from the enriched subcontinental mantle of the Proterozoic. Instead, most mafic magmas originated from depleted mantle according to their Sr and Nd isotopic compositions (Figs. 3.6a and 3.8a) and their source(s) are interpreted as the mantle wedge active in the various periods of early Paleozoic subduction. The array of the early Paleozoic magmatic rocks in the Sr–Nd isotope diagram (Fig. 3.6a) has the same form as that for the Andean Mesozoic – Cainozoic arc magmatism (Fig. 3.6b), which supports the inference of arc activity in the early Paleozoic.

3.4 Late Paleozoic – Triassic

The uplift and erosion of the early Paleozoic orogen was followed in the Devonian to lower Carboniferous by an episode of passive margin-type sedimentation and only minor magmatism (Bahlburg and Hervé 1997). The Devonian passive margin extends as far south as 48° S (Augustsson and Bahlburg 2003) and a largely uniform evolution of this large section of the margin is assumed at least until the Jurassic (Glodny et al. 2006, Chap. 19 of this volume). A renewed onset of magmatism began contemporaneously along the margin at ca. 300 Ma, between at least 21 and 40° S, and continued into the Permian and Triassic (Table 3.1; Berg and Baumann 1985; Parada 1990; Brown 1991; Lucassen et al. 1999a, 2004; Glodny et al. 2006, Chap. 19 of this volume).

Isolated late Paleozoic intrusions in the uppermost levels of the crust, mainly granite with minor diorite (Fig. 3.2), have similar compositions to the early Paleozoic granites (Fig. 3.5b). The isotopic signatures of these granites indicate considerable proportions of recycled early Paleozoic continental crust (Figs. 3.6b and 3.7b). The same is true for the silicic volcanic rocks from the extensive Permian silicic igneous province of central Chile (there known as the Choiyoi group; Kay et al. 1989; Llambias et al. 2003), the northern part of which extends into the plateau region (Breitkreuz and Van Schmus 1996). Triassic magmatism is fairly common but of minor volume and it has been poorly investigated. Available evidence suggests that recycling of older crust dominated (Berg and Baumann 1985; 32° S, Morata et al. 2000), and only for the latest Triassic have mantle-derived volcanic rocks been described (Morata et al. 2000; Bartsch 2004).

Large-scale crustal thickening after exhumation and erosion of the early Paleozoic orogen, and before the Cainozoic plateau formation, is ruled out by the sedimentary record for the late Paleozoic (e.g. Bahlburg and Breitkreuz 1991) and Mesozoic (Prinz et al. 1994; Salfity and Marquillas 1994). The record indicates low average-sedimentation rates in continental platform or shallow marine environments. The dominant tectonic style in the late Paleozoic to Mesozoic was transtension to transpression, and there are no indications for large-scale compression (Scheuber et al. 1994; Ramos 2000).

Metamorphism accompanied by deformation in the late Paleozoic–Mesozoic sedimentary rocks is absent or of very low grade. An exception is the Sierra de Limón Verde (Fig. 3.2b), where Permian high-grade metamorphism (ca. 700 °C, 12 kbar; Lucassen et al. 1999b) is unique in the region. The tectonic mechanism for the exhumation of the Limón Verde rocks is speculative, but could well be related to the Permian transpressional to transtensional regime. The local character of the exhumation of lowermost crust is underlined by the contemporaneous sedimentation of detritus from Limón Verde rocks during final uplift in the Triassic (Lucassen et al. 1999b).

Good constraints on the nature and evolution of the subcontinental mantle during this time are unavailable. We can assume from the overall geologic quiescence that the mantle composition did not change significantly from its state in the early Paleozoic, i.e., a depleted mantle persisted.

The magmatic record of the late Paleozoic into the Triassic indicates crustal recycling as an important process and the variable but generally moderate contents of mantle derived material in the granites (e.g. Lucassen et al. 1999a, 2004) resulted only in minor juvenile addition to the crust, because the magmatism is not voluminous. There is also no apparent continental growth by accretion in this time period. A late Paleozoic accretionary wedge is well known in the southern Central Andes (e.g. Hervé 1988; Glodny et al. 2006, Chap. 19 of this volume; Fig. 3.2a), but this was either not formed, or subsequently destroyed, in our area. Only the late Paleozoic Chañaral Melange, exposed south of 27° S, is a possible remnant of an accretionary wedge that once extended farther to the north (Bell 1987; Fig. 3.2a).

3.5 Jurassic – Lower Tertiary

A dramatic change in the nature of arc-related magmas took place in late Triassic – early Jurassic and this is considered as the onset of the Andean Cycle, although subduction had already commenced in the late Carboniferous after the passive margin configuration. Large volumes of mantle-derived magmas were extruded and intruded during the Jurassic to lower Cretaceous, mainly in the Coastal Cordillera (Fig. 3.1; e.g. Palacios 1978; Buchelt and Tellez 1988; Rogers and Hawkesworth 1989; Pichowiak 1994; Lucassen and Franz 1994; Lucassen et al. 2002b; Kramer et al. 2005). They form the extensive La Negra volcanic province and the related coastal batholith.

This magmatic belt is also prominent in the southern Central Andes (e.g. Vergara et al. 1995). The time span of the La Negra volcanism (early to middle Jurassic, ca. 180–150 Ma) is constrained mainly by stratigraphic evidence from intercalated marine sediments (Table 3.1; von Hillebrandt et al. 2000), by the formation ages of the batholith (ca. 190–140 Ma; Pichowiak et al. 1994; Dallmeyer et al. 1996) and by the early Cretaceous ages of dikes that represent the last magmatic phase (Lucassen and Franz 1994).

The La Negra magmas extruded mainly from fissure eruptions close to sea level (Bartsch 2004), and lava sequences up to several thousand meters thick accumulated during subsidence in a generally extensional tectonic regime (Scheuber et al. 1994). The time-space evolution of this magmatism shows a westward shift of activity in the mid-Jurassic in the northern part of the transect at Iquique (Kossler 1998; Kramer et al. 2005), but this migration is not obvious further south. The principle host rocks are early Paleozoic metamorphic rocks, and late Paleozoic sediments and intrusions, indicating that magmas intruded or were deposited on old continental crust.

The La Negra rocks comprise mainly basaltic andesite, with subordinate amounts of both more basic and more evolved compositions that locally can be volumetrically important. Local occurrences of alkaline volcanic rocks, commonly rich in sodium, can be interpreted as either melts with an important slab component (Kramer et al. 2005) or as magmas formed by local remelting of underplated material (see Petford and Gallagher 2001) from the Jurassic – lower Cretaceous magmatism. The coastal batholith is dominantly of basaltic andesite to andesite composition, but it also contains layered (ultra)-mafic intrusions and locally important bodies of granodiorite to granite.

Negative Nb-Ta anomalies in trace element distribution-patterns and a relatively low La/Yb ratio for the sub-alkaline volcanic and intrusive rocks are indicative of a magmatic arc setting. The low total REE abundances and flat REE patterns (Fig. 3.5b) argue for a high degree of melting (e.g. Lucassen and Franz 1994; Kramer et al. 2005; Bartsch 2004). Despite differences in detail among Jurassic igneous rocks, in terms of composition, age and spatial distribution (described for the area N of 20° S by Kramer et al. 2005; and at ca. 25° S by Bartsch 2004), they show considerable similarities along strike and constitute a compositionally distinct group, different from both the older (late Paleozoic) and younger (Cainozoic) magmatic suites in the region (Fig. 3.5b).

Initial Nd-Sr isotopic compositions of Jurassic – Lower Cretaceous magmatic rocks fall within in the range of the depleted mantle (Fig. 3.6b). Some samples show an elevated Sr isotope ratio, attributed to seawater alteration at the time of extrusion (Kramer et al. 2005). Initial Nd isotopic compositions for most samples show little variation ($+3.5 < \varepsilon_{Nd} < +7.5$) considering the long time span and large area represented by the dataset (Fig. 3.6b). The high values for Nd isotope ratios indicate important amounts of juvenile additions to the crust. Crustal contributions are considered to be small because the Pb isotope ratios of these mantle-derived magmas (Fig. 3.7b), which are very sensitive to additions of crustal material, are less radiogenic than those of the surrounding Paleozoic crust or average subducted sediment, which we take as a proxy for the upper crust.

Rocks from deeper sections of the coastal batholith (ca. 10 to 15 km depth) are low-P-granulite facies orthogneiss of Jurassic age (170–140 Ma; Lucassen and Thirlwall 1998), partially retrogressed to amphibolite (Lucassen and Franz 1996; Scheuber and Gonzales 1999). The metamorphism occurred in an extensional regime during long-standing magmatic activity, deformation and heating related to periodic additions to the batholith (Lucassen et al. 1996). The granulite and amphibolite are compositionally indistinguishable from their magmatic protoliths (Lucassen and Franz 1994; Lucassen et al. 2002b).

The late Cretaceous – lower Tertiary magmatic arc migrated east with time (e.g. Rogers and Hawkesworth 1989; Döbel et al. 1992; Scheuber et al. 1994; Haschke et al. 2002) and the magmas produced show an increasing contribution of Paleozoic crustal material in their isotope signatures (Fig. 3.6b). The variable composition of these magmas, and the fact that their occurrence is dispersed over a larger area, and their minor volume compared with the Jurassic – lower Cretaceous magmatism, makes an estimate of their volumetric importance to crustal growth difficult. In difference to the Jurassic – lower Cretaceous arc- there are no voluminous additions of mafic material seen in the geophysical image of the crust or in the composition of respective magmas (references or see below). The eastward shift of the magmatic arc with time is generally attributed to subduction erosion at the continental leading edge, but it is uncertain exactly how many kilometers of continental crust have been eroded since the Late Triassic. The earliest volcanic arc of the Andean Cycle must have been, in part, west of the present coastline (Fig. 3.2a; von Hillebrandt et al. 2000; Bartsch 2004), and it can be presumed that of the order of 200 km of continental crust, including parts of the Jurassic juvenile additions, have been removed by subduction erosion.

An important feature of the Cretaceous evolution, with potentially crucial consequences for understanding the Andean orogeny in the Cainozoic, is the development of a N-S-trending rift system quite far inland from the arc. The rift is now preserved as inverted basins in the eastern part of the cordillera, i.e., the Salta Rift in northwestern Argentina (Galliski and Viramonte 1988; Viramonte et al. 1999), but the rift was once more extensive, reaching as far west as the Chilean Precordillera (Salfity and Marquillas 1994). Late Cretaceous, alkaline, basaltic rocks erupted along this rift, and the upper-mantle/lower-crustal xenoliths they contain provide good constraints on the composition of the mantle and on the thermal structure of the lithosphere (Lucassen et al. 1999c, 2002b, 2005, submitted). Thermobarometry studies and isotopic dating of the xenoliths indicate high temperatures in the lower crust (ca. 900 °C) and upper mantle (> 1 000 °C) during the late Cretaceous (ca. 100 Ma), meaning that the Cainozoic orogeny was preceded by a N-S-oriented thermal anomaly in an area that now forms the eastern edge of the Andean Plateau. The basanite host rocks have depleted mantle signatures (Fig. 3.8) and the isotopic compositions of most mantle xenoliths also indicate a depleted mantle source, with some variations in isotopic composition. The variations in the U-Th-Pb systems indicate the addition of crustal material to the source and subsequent radiogenic growth probably since the early Paleozoic (Fig. 3.8).

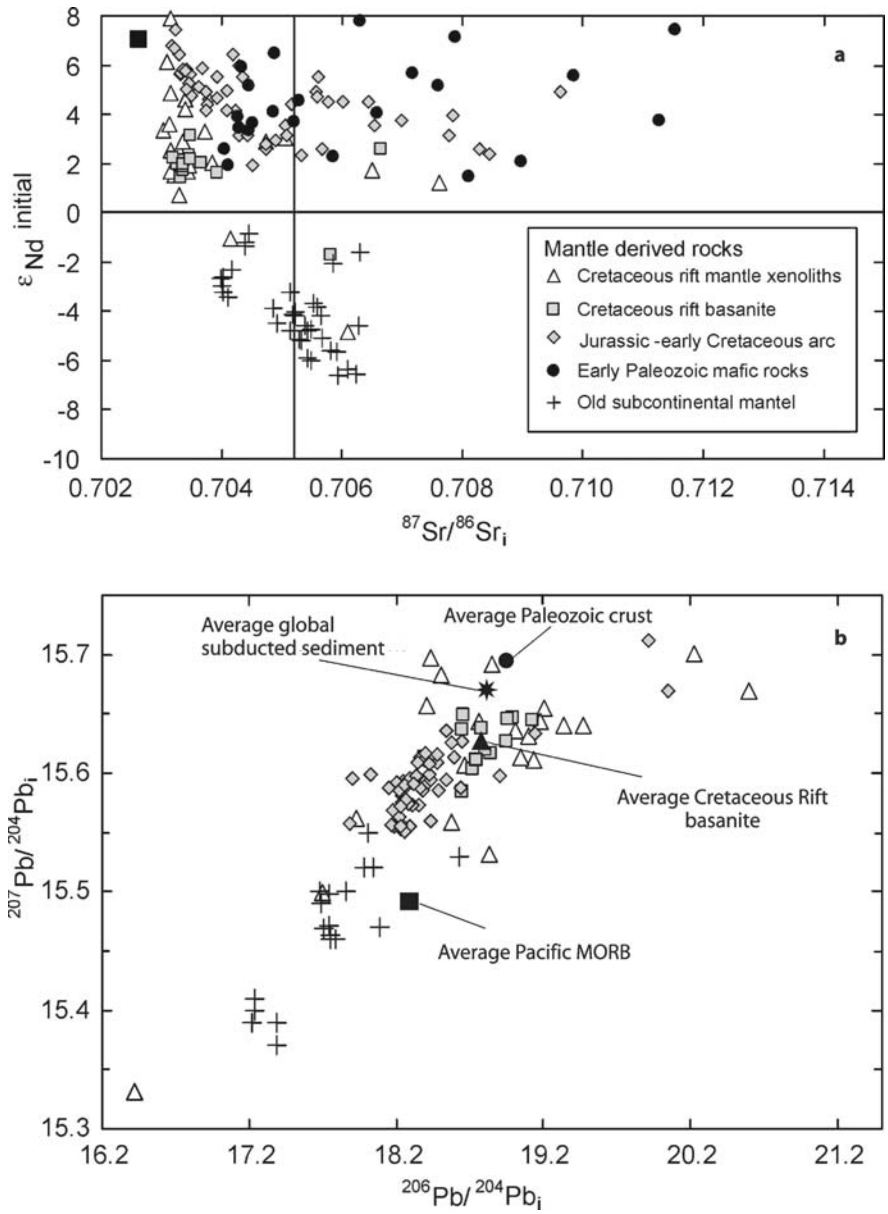

Fig. 3.8. Nd-Sr isotope (**a**) and uranogenic Pb isotope (**b**) composition of mantle-derived rocks. **a** The Jurassic – lower Cretaceous igneous rocks show compositional affinities to average Pacific MORB; many Sr isotope ratios reflect seawater alteration. The Cretaceous upper-mantle xenoliths from the rift have a broad range of Nd isotope compositions, but most samples resemble that of contemporaneous basanite from the rift. The mantle below the rift is also depleted, but slightly less radiogenic in its Nd isotope composition than the sub-arc mantle. The early Paleozoic mafic rocks are shown for comparison. There is no evidence for substantial contribution of old subcontinental mantle between 21–27° S. **b** Considering a MORB type source for the Jurassic – lower Cretaceous igneous rocks, the contributions of the crustal sources to these rocks are minor because the Pb isotope composition of many of these rocks is not dominated by crust. Pb isotope compositions, especially the $^{206}Pb/^{204}Pb$ ratios of many mantle xenoliths and basanite samples, are more radiogenic than those of the Jurassic to lower Cretaceous rocks. They are also different from the local crustal sources. The old subcontinental mantle is less radiogenic in uranogenic Pb, but shows much higher $^{208}Pb/^{204}Pb$ ratios at given $^{206}Pb/^{204}Pb$ than our samples (not shown; Lucassen et al. 2002b) and a substantial contribution of this mantle to our samples can be excluded. Data sources: Carlson et al. (1996), Gibson et al. (1996, 1999), Lucassen et al. (2001, 2002b), Kramer et al. (2004) and Bartsch (2004); for some Jurassic rocks, F. Lucassen and W. Kramer (unpublished data); average global subducted sediment, Plank and Langmuir (1998); Jurassic – lower Cretaceous igneous rocks and MORB are corrected for in situ decay to 160 Ma, all other Cretaceous rocks to 100 Ma, the early Paleozoic mafic rocks to 450 Ma

The upper mantle sampled by Late Cretaceous rift magmas must comprise at least part of the lithosphere that was later tectonically shortened in the Cainozoic and is now located under the Central Andean Plateau. From the evidence summarized above, we can conclude that the upper mantle from the Paleozoic to the Cretaceous (and most probably till the present) from 21 to 27° S has a depleted signature. The range of isotopic compositions of basic igneous rocks from the Jurassic to lower Cretaceous magmatic arc of the Coastal Cordillera, with ε_{Nd} of +3.5 to +7.5 (Figs. 3.6a and 3.8a), indicate a mantle composition significantly less depleted than the contemporaneous mid-ocean-ridge-basalt (MORB) source (Fig. 3.8) and, also, the composition is irregularly variable in time and space. This heterogeneity is especially well displayed in the trace element and isotope composition of clinopyroxene from mantle xenoliths (Lucassen et al. 2005a). We argue that these variations are due to incompletely homogenized mixtures of components in the lithosphere that reflect its geologic history, perhaps even including scarce remnants of the Proterozoic lithosphere.

For the main issue of crustal growth versus destruction, the rift-related mafic rocks represent negligible net addition to the crust because their volumes are small and occurrences restricted to small plugs and dikes. Even if one assumes an order of magnitude larger amount of intruded or underplated material at the base of the crust, the early Cretaceous rift magmatism is volumetrically small. However, the rift magmatism and its underlying thermal anomaly are highly significant in terms of their weakening effect on the lithosphere and on localizing deformation in the subsequent Cainozoic orogeny that, as is discussed below, may eventually have triggered lithospheric delamination.

3.6 The Cainozoic Orogeny

During the Cainozoic, arc magmatism continued to migrate eastward, albeit with important variations in intensity. Paleocene and Eocene magmatism was located in the present Precordillera of Chile. The composition and isotope ratios of these early Tertiary magmatic belts are similar to those of the Neogene arc (dominantly andesite) and suggest that the arc magmas were strongly affected by contamination from the crust. Juvenile (basaltic) magmas in these belts are rare, although it is reasonable to assume that the andesites are associated with more mafic intrusive and/or cumulate material in the lower crust (Haschke et al. 2002). Nevertheless, there is no magmatism in this period to compare with the juvenile Jurassic crust in the present fore-arc, and seismic velocity data from the Precordillera indicate that high-density material does not contribute much to the total crustal thickness. Oligocene and younger volcanic rocks are located in the West Cordillera and the Altiplano-Puna Plateau region to the east. There are variations in the distribution of volcanism across this region that suggest a broad Oligocene – early Miocene arc, which subsequently narrowed and became focused in the area of the present volcanic chain (Coira et al. 1982, 1993). This variation in arc location and width was interpreted to reflect a change in slab dip from shallow in the Oligocene to steeper today.

The intense tectonic shortening that led to the near doubling of crustal thickness in the Central Andes, and the development of the world's second-largest continental plateau, began around 30 Ma (e.g. Almendinger et al. 1997). The contribution of Cainozoic juvenile magmas to the crustal thickness is considered to be negligible (e.g. Francis and Hawkesworth 1994). The erupted magmas from the Tertiary and Quaternary arcs are dominantly andesite to silicic andesite, and these magmas have incorporated 20–30% of continental crust (e.g. Harmon et al. 1984; Hildreth and Moorbath 1988; Wörner et al. 1994; Trumbull et al. 1999; Siebel et al. 2001). The volume contribution to crustal growth by the remaining 70–80% of magma, which was derived from the mantle wedge, is impossible to evaluate with any certainty, but geophysical evidence from both seismic and gravity studies (Beck and Zandt 2002; Yuan et al. 2000) demonstrate that only the lowermost part of the 60–80 km-thick crust can be mafic.

Two other volcanic associations are found in the Central Andes in addition to the andesite volcanoes of the frontal arc, but neither has a significant impact on crustal growth, albeit for different reasons. The first of these comprises the regionally extensive silicic ignimbrites that erupted from large resurgent caldera complexes primarily in the late Miocene to Pliocene (Coira et al. 1993; de Silva 1989; Francis et al. 1989; Ort et al. 1996; Lindsay et al. 2001). The ignimbrite magmas are dominated by crustal sources (> 70%; e.g. Francis et al. 1989) and therefore do not contribute greatly to crustal growth despite their large volumes. The second volcanic association comprises basaltic magmas that erupted primarily behind the arc. There are isolated examples of basaltic volcanism in the late Oligocene to early Miocene from Chile (Segerstrom basalts; Kay et al. 1994) and Bolivia (Chiar Khollu; Davidson and de Silva 1995) but the greatest concentration is in the late Pliocene and Quaternary in the back-arc region of northwestern Argentina (Kay et al. 1994). These basalts represent juvenile magmas, but their volumes are too small to contribute to crustal growth. On the contrary, these back-arc basalts have been suggested to indicate an episode of delamination of the lower crust and lithospheric mantle beneath the Central Andean Plateau (Kay et al. 1994).

The Nd, Sr and uranogenic Pb isotope ratios of the late Cainozoic andesite-dacite and ignimbrite associations show a rather restricted range of compositions, suggesting a mixing trend between an arc-source and the early Paleozoic crust (Figs. 3.6b and 3.7b). These diagrams illustrate the problem that estimating the degree of crustal involvement in a given suite of rocks depends on assumptions about the end-members. For example, there may be contributions to the crustal 'component' from lower crustal rocks that were depleted in Rb and U during the early Paleozoic orogeny. These will have less radiogenic Sr and Pb isotope ratios (Fig. 3.6a,b) than their counterparts in the mid-crust.

The composition of the sub-arc mantle wedge is poorly constrained because all of the arc magmas are differentiated and have been contaminated by crustal material either in the source via subduction processes, or during their ascent through the crust. Nevertheless, Nd and Sr isotope signatures of andesites from the main arc, the ignimbrites and most basalts from the back-arc region follow the same mixing trajectory between Paleozoic crust and a depleted mantle. Helium isotope signatures indicate an asthenospheric source for the primary mantle magma of the hybrid back-arc basalts (Pilz et al. 2001).

3.7 Discussion

The long record of compositional variations, metamorphism and tectonics preserved at the Andean continental margin (Table 3.1) is the basis for identifying the processes contributing to growth, destruction or modification of the continental crust and upper mantle, and for assessing their long-term importance. This discussion is focussed on the processes of continental growth and destruction at active margins as outlined in the introduction.

3.7.1 Crustal Growth: Juvenile Magmas and Tectonic Accretion

The data on age structure, metamorphic grade and isotopic-geochemical characteristics of the early Paleozoic metamorphic, magmatic and sedimentary rocks in the Central Andes (Table 3.1) support the interpretation of a coherent, autochthonous early Paleozoic mobile belt at the western leading edge of South America (Fig. 3.2; Aceñolaza and Miller 1982; Damm et al. 1994). No early Paleozoic exotic terranes have been recognized between 18 and 27° S. The geotectonic setting at this time was probably a non-collisional active continental margin with an orogen comparable to the Cainozoic orogen (Lucassen and Franz 2005b).

This interpretation contrasts with suggestions that the Paleozoic basement is a mosaic of accreted terranes (e.g. Ramos et al. 1986; Ramos 2000). The arguments against late Proterozoic to Phanerozoic terrane accretion models for the areas between 18 and 27° S have been discussed in detail before (Becchio et al. 1999; Bock et al. 2000; Lucassen et al. 2000; Franz and Lucassen 2001) and rely mainly on the similar ages of metamorphic and magmatic crystallization, the age pattern of the inherited material, chemical and isotopic composition of sedimentary, magmatic and metamorphic rocks, and the type and distribution of early Paleozoic metamorphism, e.g. the absence of high pressure rocks typical of suture zones.

South of 28° S, the Argentine Precordillera is widely believed to be a Laurentia-derived exotic terrane based on its stratigraphic and faunal record (see reviews in Ramos and Keppie 1999). If the Argentine Precordillera accreted in the Ordovician as proposed, this would constitute a significant contribution to growth of the local crust. Our data are not inconsistent with terrane accretion south of 28° S, but, if it took place, we can conclude that the process is not discernable from the early Paleozoic pressure-temperature-time record of the adjacent high-grade metamorphic basement (Fig. 3.2; Lucassen and Becchio 2003).

Continental growth by accretion of oceanic crust and continental sediments from the lower plate is considered to be globally minor, based on observations of presently active margins (Clift and Vanucchi 2004, and references therein), and the Central Andean record is no exception. The existence and preservation of accretion complexes could be an important indicator of long-term stability of the trench-margin relationships. The development of accretionary wedges depends mainly on convergence geometry and rates, morphology of the lower plate and sediment thickness in the trench, which in turn may be linked to climate. These factors are unlikely to remain constant over the time span considered here, so it is probable that accretionary and non-accretionary periods alternated during the history of the active margin, as observed in the fore-arc of the southern Central Andes (Glodny et al. 2006, Chap. 19 of this volume). However, Phanerozoic accretionary units are generally absent in the transect area, except for the local Chañaral melange.

Early Paleozoic magmatism was extensive and has been well studied, but the addition of mantle magmas was apparently volumetrically unimportant to the growth of the continental crust throughout this period. The magmas produced were largely granitic and their isotopic composition clearly shows that recycling of Proterozoic crust was the dominant source. The same is true for late Paleozoic igneous rocks, which were derived from early Paleozoic and Proterozoic crustal sources. There could, potentially, have been substantial intrusions of juvenile magmas in the lower crust that are not now exposed. However, the excellent geophysical coverage of the Central Andes shows that dense mafic material is a very minor component of the crust today (Asch et al. 2006, Chap. 21 of this volume), and there are no indications in the composition of Mesozoic magmas for a thick Paleozoic mafic lower crust. Igneous rocks derived from partial melting of mafic crust are compositionally distinct (e.g. in the coastal batholith of Peru: Atherton and Petford 1993; Petford and Atherton 1996; and tonalite-trondhjemite-granodiorite rocks in the early Paleozoic orogen: Pankhurst et al. 2000) and such rocks are rare or absent among the Paleozoic intrusions as well as among the Cainozoic volcanic rocks.

The occurrence of magmas derived from mafic (lower) crust in the Central Andes seems restricted to the Mesozoic province of the Coastal Cordillera and Precordillera, where a mafic to intermediate crust formed in the Jurassic and Lower Cretaceous (Lucassen et al. 1994, 2002b; Haschke et al. 2002). The tectonic regime operating at this period was extensional to transpressional and caused by strongly oblique subduction (e.g. Scheuber et al. 1994). This mafic crust in the Coastal Cordillera has a clear geophysical expression (seismic velocity and gravity), which increases our confidence in concluding that thick mafic lower crust is absent in the Paleozoic belt east of the Coastal Cordillera (see other contributions of this volume). Indeed, the Jurassic – lower Cretaceous magmatism in the Coastal Cordillera is the only important episode of crustal growth from mantle magmas in this section of the Central Andes. Some reworking of mafic crust is found in the Paleocene and Eocene magmatic arcs of the Precordillera (Haschke et al. 2002).

3.7.2 Crustal Destruction: Subduction Erosion and Lithospheric Delamination

The process of subduction erosion in the Central Andes today is related to morphological roughness of the oceanic plate and the climate-related lack of sediment infill of the trench. It is evidenced by the young tectonic features of the fore-arc and continental slope (e.g. von Huene and Ranero 2001; Kukowski and Oncken 2006, Chap. 10 of this volume). The climate, which plays a key role in the sediment supply from the continent, was likely dominated by arid conditions from the Jurassic onwards (Hartley et al. 2005 and references therein) and adds a long-term component to the subduction erosion model, which is based mainly on observations of active features. Presently, in northern Chile the loss of crust by subduction erosion is greater than the replenishment by magmatic additions, and the result produces a trench retreat of ~3 km Ma^{-1}, or 40 to 45 km^3 per Ma and km at 22° S during the last 8 Ma (Kukowski and Oncken 2006, Chap. 10 of this volume). Clift and Vanucchi (2004) show that arc-retreat rates of this order can be maintained for the short-term (tens of Ma), but for the long-term (200 Ma) history relevant to the Central Andes, an average retreat rate of 1 km Ma^{-1} seems realistic.

The overall volumetric importance of subduction erosion in the Central Andes since the early Paleozoic is difficult to assess because it depends on knowing the initial configuration of the continental margin and convergence parameters, as well as the assumption of a constant distance between trench and magmatic front of the eastward-migrating Mesozoic magmatic arcs (e.g. Scheuber et al. 1994). A rough estimate based on the latter assumption, is that up to ~200 km of crust have been lost since the late Mesozoic (Scheuber et al. 1994, and references therein; Kramer et al. 2005; Kukowski and Oncken 2006, Chap. 10 of this volume).This eroded crust comprised parts of the Jurassic magmatic arc, as well as the original late Paleozoic fore-arc, based on the assumption of a northern continuation of the Chañaral Melange. This estimate implies that at least a part of the subducted continental crust was juvenile crust formed in the Mesozoic.

A second potential mechanism for destruction of continental crust at active margins is delamination of the roots of overthickened crust into the mantle (e.g. Kay and Kay 1993; Meissner and Mooney 1998, and references therein). This mechanism is density-driven and therefore restricted to mafic rocks that develop sufficient high density by high-P-phase transitions (basalt-eclogite) to sink into the upper mantle. Delamination is suggested to have taken place under the ultra-thick Central Andean Plateau in the late Pliocene (Kay et al. 1994). The arguments for this proposal come from the occurrence and composition of back-arc basalts, from seismic evidence of low-velocity upper mantle with high S-wave attenuation, and from the coincidence of high plateau elevation with only moderate tectonic shortening. Sobolev et al. (2006, Chap. 25 of this volume) can show by numerical modeling that lithospheric delamination is in fact a realistic scenario for the Central Andes.

Delamination may also have been an important process for removing mafic crust in the early Paleozoic orogen. In a period of more than 100 Ma the Paleozoic belt that extended over thousands of kilometers along the continental margin (Fig. 3.3) must have formed an impressive orogen and it is very likely that in such a subduction system lithosphere was delaminated and destroyed parts of the lower crust. There is no direct petrological evidence for delamination, but, on the other hand, geophysical evidence suggests a dominantly felsic crust under the now-exposed Paleozoic belt, and there are no younger igneous rocks with compositions consistent with a mafic crustal source. By contrast, the Jurassic–early Cretaceous crust of the Coastal Cordillera in the Central Andes and the Cainozoic crust in the Southern Andes have the mafic to intermediate compositions that would be expected for active margins where subduction magmatism continuously delivers mafic magma to the base of the crust. A subduction setting is certain for the Cainozoic orogen in the Central Andes and was also most likely the case in the early Paleozoic, so the apparent lack of mafic crust in these two examples probably means that it has been lost. In summary, even if there are good indications for mafic additions to the crust by andesite magmatism in the Cainozoic, this is likely to be balanced out by lithospheric delamination.

Loss of crustal material into the mantle requires some return flow. Both aspects are relevant for the long-term evolution of the crustal and mantle reservoirs in general (e.g. Jacobsen 1988; Kramers and Tolstikhin 1997), but in this paper we restrict ourselves to the specific aspects of arc-evolution. As described above, the composition of the Andean upper mantle differs from that of the major MORB-source reservoir despite its generally depleted character. The compositional differences could reflect input of crustal material into the mantle since the early Paleozoic, either by subduction or by delamination. Long-term separation from the convective mantle could have contributed to the specific evolution of the Sr, Pb, Nd radiogenic isotope systems of the lithospheric mantle, as indicated by the trace element contents and isotope signatures of mantle xenoliths (Lucassen et al. 2002b, 2005, submitted). The pre-Paleozoic lithospheric mantle was possibly similar to that beneath other sections of the Brazilian shield and, if so, a nearly complete exchange with depleted sub-arc mantle took place starting in the early Paleozoic orogeny.

The present data set and isotopic data from mantle-derived igneous rocks in the southern Central Andes (Lucassen et al. 2004, and references therein) allow us to speculate that the convective asthenospheric mantle beneath the Andean continental margin has developed a uniform composition, over a large space and time domain, which is distinct from a common Pacific MORB source. The Andean margin has been active for a very long time in a more or less constant position at the edge of old subcontinental lithosphere. The uniform composition of the mantle-derived magmas can be created by mixing material from a MORB source with old continental crust or with isotopically similar continental sediments deposited on the oceanic plate. Alternatively, the rocks could represent magmas derived from a uniform, compositionally modified MORB-like source.

During the Phanerozoic evolution of the Andean active margin, processes of crustal recycling dominate, but processes of crustal growth and destruction are occasionally important. Global estimates of crustal growth assume a nearly constant volume or a slight volume increase of the crust since the Precambrian, with 80 to 90 vol% of the present crust having formed by the beginning of the Phanerozoic (e.g. Jacobsen 1988; Kramer and Tolstikhin 1997; Collerson and Kamber 1999; Condie 2002). The global system of mid-ocean ridges and subduction zones has been the principle setting for mass flux between the mantle and crust, at least for the Phanerozoic. Provided that subduction erosion is not a short-lived transitional state of subduction zones, but is characteristic for their long-term evolution, the condition of constant crustal volume would require considerable addition of juvenile material from the mantle to balance the loss. The place for compensating crustal losses must be the continental and intra-oceanic arcs.

In the Central Andes between 21 and 27° S, the isotope and trace element compositions of all major Paleozoic magmatic, metamorphic and sedimentary units indicate recycling of a Proterozoic felsic crust formed ca. 2 Ga ago (Figs. 3.4, 3.6 and 3.7; and reworked in the Mesoproterozoic, in accordance with major cycles of crustal growth and orogenic reworking of South America (Cordani et al. 2000b; Sato and Siga 2002) The dominant processes in the early Paleozoic imply conservation of the existing felsic crustal material. The same holds true for the Cainozoic orogen, where juvenile additions to the crust by arc magmas are of minor importance volumetrically.

Regionally, significant crustal growth was restricted to the extensional-transpressional Jurassic to lower Cretaceous arc. This may imply that large-scale crustal growth by juvenile magmatic additions in continental arcs requires certain geotectonic situations. An oblique-subduction setting with extensional tectonics may be necessary to allow efficient mantle melting and intrusion of magmas into the crust. Conversely, orthogonal subduction may favor destruction of the crust by subduction erosion under conditions of arid climate and lack of trench sediments. The strong influence of the long-term prevailing climate conditions is indicated by the increasing preservation of a late Paleozoic – early Mesozoic accretionary wedge towards the south. A near steady state situation of the active margin since Mesozoic is described south of 38° S (Glodny et al. 2006, Chap. 19 of this volume, and references therein). Where, as in the Central Andes, subduction orogeny leads to major crustal shortening and thickening, further loss of the crust can occur by lithospheric delamination. This delaminated lower crust is prevailing juvenile, mafic and relatively young because it was probably replenished between or during the respective orogenies by magmatic underplating (c. 1200–900 Ma, 560–400 Ma, 30 Ma) and only the more differentiated juvenile rocks contributed to the crust. The statement that active continental margins and the continental subduction environment are the major site for growth of the continents and the continental crust is, in this simplified form, definitely not correct.

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