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Proterozoic–Paleozoic development of the basement of the Central Andes (18–26°S) — a mobile belt of the South American craton

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Abstract

The Late Precambrian–Early Paleozoic metamorphic basement forms a volumetrically important part of the Andean crust. We investigated its evolution in order to subdivide the area between 18 and 26°S into crustal domains by means of petrological and age data (Sm–Nd isochrons, K–Ar). The metamorphic crystallization ages and t_{DM} ages are not consistent with growth of the Pacific margin north of the Argentine Precordillera by accretion of exotic terranes, but favor a model of a mobile belt of the Pampean Cycle. Peak metamorphic conditions in all scattered outcrop areas between 18 and 26°S are similar and reached the upper amphibolite facies conditions indicated by mineral paragensis and the occurrence of migmatite. Sm–Nd mineral isochrons yielded 525 ± 10, 505 ± 6 and 509 ± 1 Ma for the Chilean Coast Range, the Chilean Precordillera and the Argentine Puna, and 442 ± 9 and 412 ± 18 Ma for the Sierras Pampeanas. Conventional K–Ar cooling age data of amphibole and mica cluster around 400 Ma, but are frequently reset by Late Paleozoic and Jurassic magmatism. Final exhumation of the Early Paleozoic orogen is confirmed by Devonian erosional unconformities. Sm–Nd depleted mantle model ages of felsic rocks from the metamorphic basement range from 1.4 to 2.2 Ga, in northern Chile the average is 1.65 ± 0.16 Ga (1σ; $n = 12$), average t_{DM} of both gneiss and metabasite in NW Argentina is 1.76 ± 0.4 Ga (1σ; $n = 22$), and the isotopic composition excludes major addition of juvenile mantle derived material during the Early Paleozoic metamorphic and magmatic cycle. These new data indicate a largely similar development of the metamorphic basement south of the Arequipa Massif at 18°S and north of the Argentine Precordillera at 28°S. Variations of metamorphic grade and of ages of peak metamorphism are of local importance. The protolith was derived from Early to Middle Proterozoic cratonic areas, similar to the Proterozoic rocks from the Arequipa Massif, which had undergone Grenvillian metamorphism at ca. 1.0 Ga.
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Keywords: Metamorphic basement; Cratonic area; Isotopic composition

1. Introduction

The Paleozoic evolution of the Central Andes has been the subject of considerable debate during recent years. The present Pacific margin of South America is widely considered to be an assemblage of allochthonous or par-autochthonous terranes that were accreted to Gondwana during the Late Proterozoic and/or during the Early to Mid-Paleozoic (e.g. Ramos, 1988, 1995; Unrug, 1996; Astini et al., 1995). Four such terranes were recognized by Omarini et al. (1999) and by; Ramos (1988, 1995; Fig. 1) and three by Bahlburg and Hervé (1997) in the study area. One tectonic scenario is that terrane exchange occurred between the South American (Gondwana) and North American (Laurentia) cratons following Grenvillian collision which formed the hypothetical supercontinent Rodinia (Wasteneys et al., 1995; Tosdal, 1996; Dalziel, 1997). Subsequent separation was followed by a second collision between Laurentia and Gondwana during the Mid-Ordovician Famatinian orogeny in the Andes and the Taconian orogeny in the Appalachians (Dalziel et al., 1994). A modification of the latter model places a small ocean between Laurentia and South America with arcs on or in front of the

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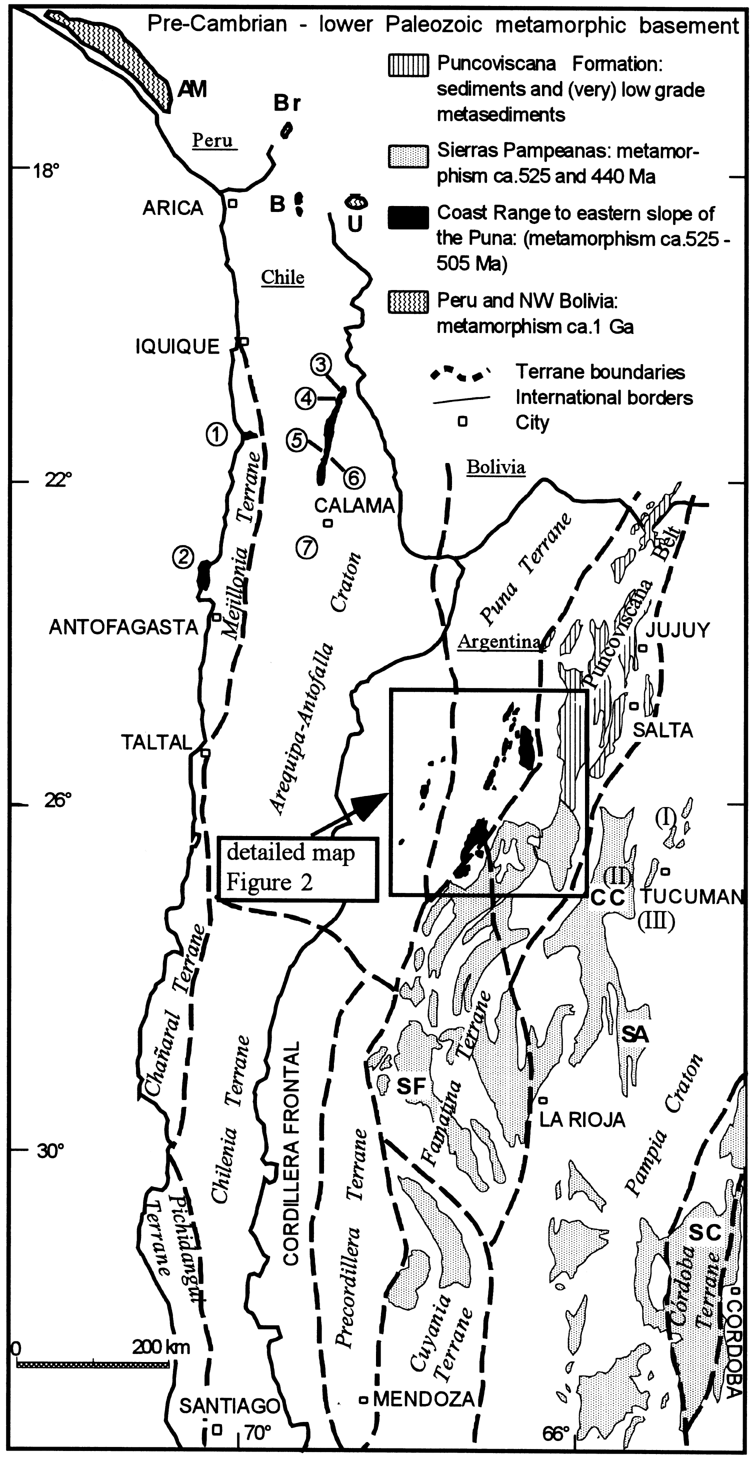

Fig. 1. Map of the major Precambrian to Early Paleozoic units in N. Chile and NW Argentina (compiled from Damm et al., 1990; Lezaun et al., 1997; Lottner and Miller, 1986; Rapela et al., 1990) and of the sample locations cited in the text: 1, Caleta Loa; 2, Mejillones; 3 to 6, Sierra de Moreno (3, Quebrada Choja; 4, Tambillo o Seca; 5, Quebrada Arcas; 6, Quebrada Quinchamale, parallels the eastern slope of southern Sierra de Moreno); 7, Sierra Limón Verde, (I) = Sierra de Medina, (II) = E' Cumbres Calchaquies, (III) = Las Viñas. Other locations (from N to S) are: AM, Arequipa Massif; Br, Berenguela; B, Belen-Tignamar; U, Uyarani; CC, Cumbres Calchaquies; SA, Sierra de Ancasti; SF, Sierra Fiambalá; SC, Sierras de Córdoba. Terrane boundaries (dashed lines) and terminology are from Ramos (1995).

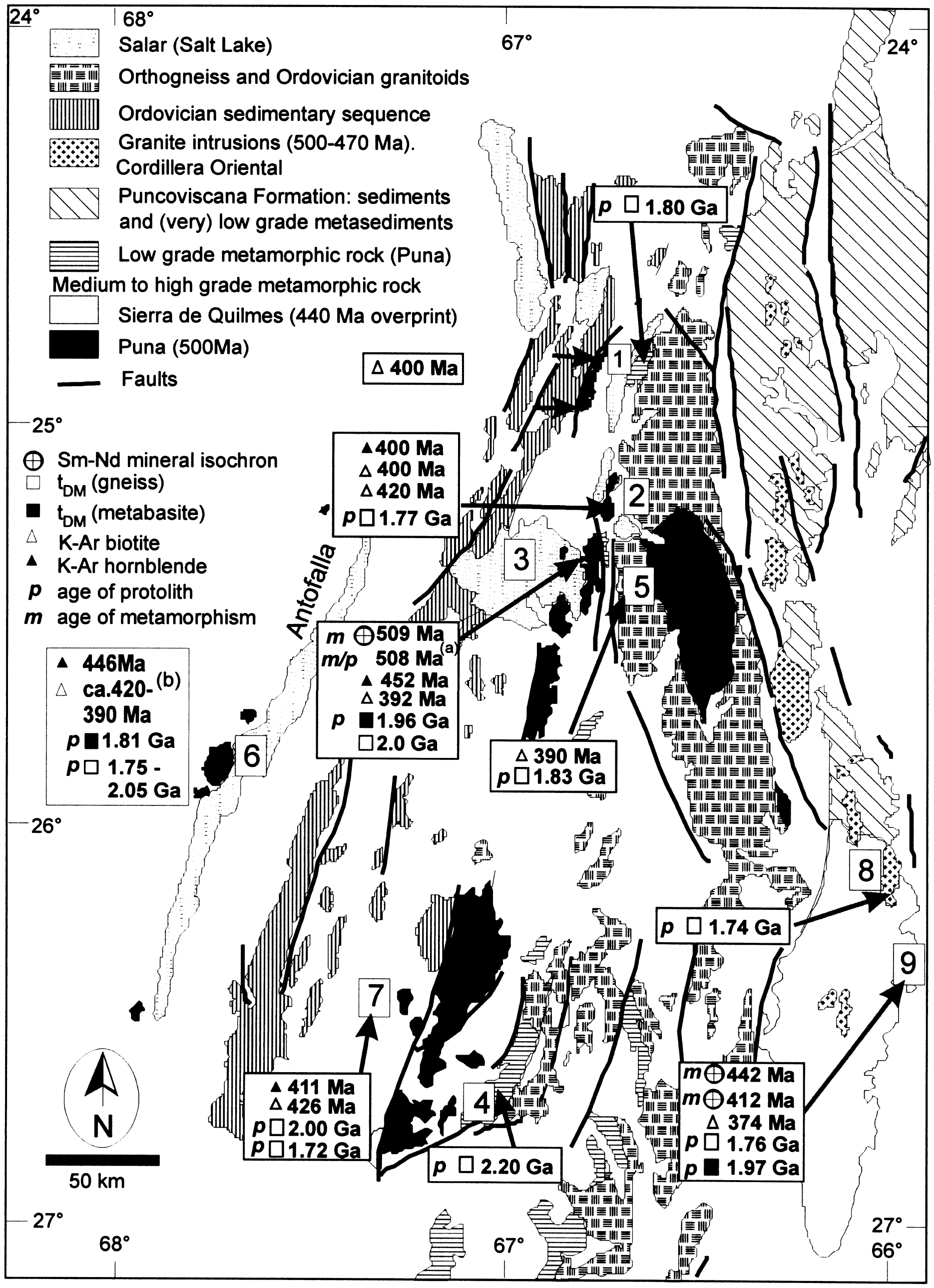

Fig. 2. Map of the major Precambrian to Early Paleozoic units in NW Argentina (modified after Hongn, 1994), including isotopic ages (Tables 1 and 2). 1, Salar Centenario; 2, Salar Ratones and Salar Diablillos, 3, Salar Hombre Muerto; 4, Laguna Blanca; 5, Cerro Blanco; 6, Salar Antofalla; 7, El Jote and El Peñon; 8, Cafayate; 9, Rio Anchillo – Qda. de Quilmes. Additional isotopic age data: (a) from Becchio et al. (1996) and (b) from Kraemer et al. (1999).

respective margins in the Early Ordovician. In Mid-Ordovician followed the possible collision of a continental platform derived from Laurentia including the Argentine Precordillera (Dalziel, 1997). These global tectonic reconstructions of the Paleozoic history of the Central Andes are based primarily on the interpretation of the sedimentary-faunal record, the tectonic–magmatic evolution of the basins (above references) and paleomagnetic evidence (cf. Dalziel, 1997).

Interpretation of the metamorphic history resulting from the accretion events was hampered by the lack of reliable data on the development of the metamorphic basement of the area. In the Andes outcrops of metamorphic rocks are scarce but present in all mountain chains from the Chilean Coast to their eastern slope. In N. Chile continuous outcrops do not exist (Fig. 1; Damm et al., 1990, 1994). In NW Argentina (Figs. 1 and 2) the metamorphic rocks of the Puna region (Viramonte et al., 1993; Miller et al., 1994) are distinguished from the basement of the Sierras Pampeanas (Rapela et al., 1990; Miller et al., 1994 which comprise

Table 1
Results of Sm and Nd isotopic determinations on minerals and whole rocks; calculated mineral isochron ages and t_{DM} model ages (abreviations: MB, metabasite; G, gneiss; O, orthogneiss; S, schist; C, calcsilicate; Gr, granite; wr, whole rock; Grt, garnet; epi, epidote; hbl, hornblende; tit, titanite; cpx, clinopyroxene; opx, orthopyroxene; pl, plagioclase a, depleted mantle model of Goldstein et al. (1984))

	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$[a]	$^{143}Nd/^{144}Nd \pm 2\sigma$	Ages (Ma)		t_{DM} (a) (Ga)
Chile							
Mejillones							
3/379 wr (MB)	3.639	11.740	0.1873	0.512906 ± 5	3/379 all points	3/379 hbl–grt–pl	1.45
3/379 hbl	4.284	12.940	0.2001	0.513021 ± 4	491 ± 147 Ma	525 ± 10 Ma	
3/379 pl	0.354	1.962	0.1090	0.512707 ± 4	Ndi 0.512354 ± 161	Ndi 0.512332 ± 11	
3/379 grt	2.904	10.342	0.1697	0.512913 ± 5	MSWD = 269	MSWD = 1.2	
3/382 wr (G)	4.920	23.100	0.1288	0.512330 ± 5			1.48
Caleta Loa							
3/354 wr (G)	5.686	28.118	0.1223	0.512154 ± 5			1.67
4/354 wr (G)	1.973	9.789	0.1219	0.512231 ± 5			1.53
Qda. Choja							
4/327 wr (C)	5.896	28.321	0.1258	0.512258 ± 6	4/327 all points	4/327 grt–epi–tit	1.56
4/327 grt	4.332	6.497	0.3937	0.513123 ± 4	496 ± 12 Ma	505 ± 6 Ma	
4/327 epi	7.634	18.231	0.2534	0.512660 ± 7	Ndi 0.511841 ± 23	Ndi 0.511821 ± 12	
4/327 tit	136.160	515.600	0.1596	0.512348 ± 8	MSWD = 9	MSWD = 0.14	
4/316 wr (G)[b]	7.415	36.630	0.1224	0.512087 ± 18			1.78
4/329 wr (G)[b]	6.668	34.300	0.1175	0.512096 ± 16			1.68
Qda. Arcas area							
4/76 wr (G)	3.810	14.021	0.1643	0.512618 ± 5			1.66
3/287 wr (G)[b]	7.733	35.360	0.1322	0.512128 ± 16			1.92
3/291 wr (G)	5.218	26.694	0.1182	0.511961 ± 4			1.90
Limon Verde							
6/6wr (MB)	9.803	44.420	0.1334	0.512700 ± 5			0.87
3/303 wr (MB)	5.358	23.753	0.1363	0.512754 ± 5			0.80
3/306wr (G)	3.947	19.164	0.1245	0.512220 ± 9			1.61
3/307wr (G)	5.687	28.151	0.1222	0.512187 ± 5			1.61
5/29wr (G)	4.463	20.664	0.1306	0.512417 ± 10			1.36
Argentina							
Antofalla							
6/147 wr (MB)[b]	1.975	7.774	0.1536	0.512440 ± 13			1.81
6/160 wr (G)[b]	7.438	40.050	0.1123	0.511989 ± 10			1.75
T2 wr (G)[c]	8.01	34.80	0.1392	0.512150 ± 5			2.05
T4 wr (G)[c]	2.70	14.90	0.1096	0.511672 ± 9			1.76
Centenario							
7/49 wr(S)[b]	6.177	30.990	0.1205	0.512053 ± 16			1.80
FC-51 wr (S)	5.335	28.10	0.1148	0.511988 ± 7			1.80
Diablillos							
6/19 wr(O)[b]	5.461	25.490	0.1295	0.512176 ± 13			1.77
6-58 wr (O)[b]	5.947	29.58	0.1216	0.512117 ± 10			1.72
Cerro Blanco							
6/84 wr (O)[b]	3.861	16.998	0.1373	0.512235 ± 10			1.83
Hombre Muerto							
4/164 wr (C)	7.211	33.320	0.1307	0.512086 ± 5	4/164 all points		1.96
4/164 grt	6.842	8.955	0.4619	0.513188 ± 5	509 ± 1 Ma		
4/164 epi	7.478	20.693	0.2184	0.512378 ± 5	Ndi 0.511649 ± 3		
4/164 cpx	0.316	1.843	0.1034	0.511993 ± 11	MSWD = 0.3		
4/164 tit	61.580	404.220	0.0920	0.511955 ± 4			
1/55 wr (G)[c]	6.000	27.00	0.1371	0.512091 ± 14			2.03
mi07 wr (O)[b]	3.445	14.65	0.1422	0.512171 ± 17			2.09
Laguna Blanca							
7/145 wr (S)[b]	6.364	29.46	0.1306	0.511950 ± 12			2.20
El Jote							
6/143 wr (O)[b]	3.178	16.786	0.1145	0.511846 ± 17			2.00
El Peñon							
6/131wr (O)[b]	5.969	31.110	0.1160	0.512051 ± 23			1.72
Sierra de Quilmes							
6/120 wr (G)	5.112	25.768	0.1199	0.512069 ± 4	6/120 all points		1.76
6/120 grt	6.875	17.713	0.2346	0.512403 ± 6	442 ± 9 Ma		
6/120 pl	1.690	9.144	0.1117	0.512048 ± 4	Ndi 0.511723 ± 8		

Table 1 (*continued*)

	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$[a]	$^{143}Nd/^{144}Nd \pm 2\sigma$	Ages (Ma)		t_{DM} (a) (Ga)
					MSWD = 0.9		
6/111wr (MB)	3.065	11.940	0.1551	0.512426 ± 5	6/111 all points	6/111 pl–opx–cpx	1.97
6/111pl	0.663	5.351	0.0748	0.512177 ± 4	420 ± 58 Ma	412 ±18 Ma	
6/111opx	0.912	3.542	0.1557	0.512400 ± 4	Ndi 0.511975 ± 56	Ndi 0.511976 ± 17	
6/111cpx	12.466	38.541	0.1955	0.512500 ± 5	MSWD = 61	MSWD = 5.6	
Cafayate							
7/176 wr (Gr)[b]	1.837	7.662	0.1450	0.512369 ± 32			1.74
Rancagua							
7/159 wr (S)[b]	7.767	39.20	0.1198	0.512009 ± 11			1.86
Sierra Medina							
7/161 wr (S)[b]	8.043	39.13	0.1243	0.512244 ± 36			1.55
E' Cumbres Calchaqui							
7/189 wr (S)[b]	7.042	34.69	0.1227	0.512119 ± 65			1.73
Las Viñas							
7/200 wr (G)[b]	4.609	22.11	0.1261	0.512234 ± 21			1.60

[a] Measured at GFZ Potsdam.
[b] Measured at Zentrallaboratorium für Geochronologie.
[c] The assumed error on the Sm/Nd ratio is 0.1%.

the upper Precambrian to Lower Cambrian Puncoviscana formation and equivalents and high temperature–low pressure high-grade rocks (Fig. 1). A summary of previous work on metamorphic rocks is given by Damm et al. (1990, 1994) and Miller et al. (1994). The high-grade basement of the southern Sierras Pampeanas in the Sierras de Córdoba (ca. 30–32°S) was investigated by Pankhurst et al. (1998) and Rapela et al. (1998a,b). Our paper presents new petrological descriptions, Sm–Nd mineral — whole rock isochrons, K–Ar ages on minerals and whole rock Sm–Nd systematics from the Coast Range of Chile to the eastern slope of the Argentine Puna. The aim of the paper is to reveal a possible scenario for the Early Paleozoic development from the viewpoint of metamorphic basement development with a focus on the area between 21 and 26°S.

2. Geological setting and petrology

2.1. Northern Chile

The scattered outcrops of the metamorphic basement in N. Chile (Fig. 1) have common lithologic features and contact relations with the surrounding sediments or post-metamorphic igneous rocks. Quartz-rich metasediments and granitic to dioritic orthogneisses are dominant, whereas metabasites are rare and comprise less than 5% of the outcrop area. Ductile deformation caused one major penetrative foliation, with a general N–S strike. Upper Paleozoic intrusions show no penetrative ductile deformation. The dioritic to granitic magmas intruded at high levels into metamorphic and sedimentary rocks. The high level of intrusion is consistent with the general absence of primary white mica in Al-rich granites and by the maximum thickness of Devonian to Carboniferous sediments (<3 km; Breitkreuz, 1986).

A short description of the sample areas for isotopic dating is given below, for more detailed descriptions the reader is referred to Baeza (1984) for Mejillones (locality 2 in Fig. 1), to Lezaun et al. (1996, 1997) for the Belen area and to Lucassen et al. (1994) for the other locations. The Sierra de Limón Verde (locality 7 in Fig. 1) shows a Permian age of the high-grade metamorphism (Lucassen et al., 1999a).

Sierra de Moreno. The most common rock types of the Quebrada Arcas — Quebrada de Quinchamale area (localities 5 and 6 in Fig. 1) are quartz-rich gneisses of upper amphibolite facies with varying proportions of feldspar, biotite and rare muscovite. Their protoliths were graywacke type sediments as inferred from bulk composition (Lucassen et al., 1999b) and compositional layering; we observed series of layers in the metasediments starting with quartzite and grading into mica-rich gneiss or schist, which is repeated several times. Quartzite and calc-silicate rocks that are present in small amounts in the metasediments, dioritic orthogneiss and amphibolite, were found at one location only. Migmatization (interlayering of quartzofeldspathic and biotite-rich layers) occurs locally in these rocks, especially on the western slope of the mountain chain. This metamorphic unit was intruded by granitic to dioritic plutons at around 300 Ma (Maksaev, 1990; Lucassen et al., 1999b). Indications for contact metamorphism were not observed in the metamorphic rocks, but are present where the plutons intrude the Paleozoic sediments. On the western slope, the schists are unconformably overlain by shales of presumed Devonian to lower Carboniferous age (Breitkreuz, 1986) with a basal conglomerate that contains clasts of the metamorphic rocks. In the areas of Quebradas Choja and Tambillo O Seca (localities 3 and 4, Fig. 1), we

Table 2
Results of K–Ar dating

Sample	Location	Rock type	Mineral	K–Ar age (Ma)	K_2O (wt%)
Chile					
Belen	18°25′10″				
4/305	69°33′20″	Amphibolite	hbl	457 ± 7	0.53
3/232		Orthogneiss	hbl	358 ± 10	0.81
Tignamar	18°34′07″				
3/251	69°28′45″	Orthogneiss	hbl	389 ± 11	0.58
Qda. Choja	21°04′33″				
3/184	68°54′20″	Amphibolite	hbl	406 ± 13	0.28
4/325	21°04′32″S	Calc-Silcate	hbl	478 ± 15	0.54
	68°52′45″W				
4/329	68°52′45″W	Migmatite	bt	296 ± 6	9.3
Qda. Tambillo S.	21°30′00″				
3/263	69°01′10″	Migmatite	hbl	422 ± 12	0.68
Qda. Arcas	21°46′48″				
3/288	69°06′20″	Amphibolite	hbl	382 ± 11	0.66
3/289	69°06′20″	Amphibolite	hbl	327 ± 9	0.89
3/291	69°06′20″	Micaschist	bt	284 ± 6	7.43
4/76	21°52′64″	Migmatite	hbl	311 ± 7	0.72
	69°05′52″				
4/72	69°05′52″	Migmatite	bt	272 ± 6	7.77
Caleta Loa	21°25′30″				
3/278	70°00′40″	Orthogneiss	bt	170 ± 3	5.26
4/354	21°26′37″	Migmatite	bt	189 ± 4	7.81
	69°52′24″				
Mejillones	23°18′32″				
3/364	70°34′30″	Amphibolite	hbl	152 ± 5	0.47
3/382	23°15′29″	Gneiss	bt	151 ± 3	8.28
	70°35′07″				
Argentina					
Centenario	24°57′59″	Gneiss	bt	400 ± 8	9.83
6/19	66°49′36″				
Ratones	25°09′05″				
6/44	66°47′15″	Orthogneiss	bt	401 ± 10	9.74
Cerro Blanco	25°29′04″				
6/83	66°45′17″	Gneiss	bt	390 ± 9	9.92
6/84		Orthogneis	bt	378 ± 8	9.4
Diablillos	25°15′18″				
6/48	66°46′42″	Orthogneiss	bt	398 ± 8	9.75
6/55	25°16′59″	Amphibolite	hbl	400 ± 20	0.17
6/58	66°41′09″	Gneiss	bt	420 ± 9	9.83
Hombre Muerto					
4/161	25°19′58″	Amphibolite	hbl	452 ± 12	1.49
4/155	-66°51′00″	Gneiss	bt	392 ± 8	9.64
7/83		Orthogneis	bt	390 ± 9	9.53
El Jote	26°26′56″				
6/139	67°25′56″	Amphibolite	hbl	411 ± 9	0.61
6/142	67°25′56″	Gneiss	bt	426 ± 9	9.22
6/142		Gneiss	ms	371 ± 8	10.3
El Peñón	26°27′75″				
6/131	67°14′84″	Orthogneiss	bt	383 ± 10	8.76
Antofalla	25°43′48″				
6/145	67°46′41″	Amphibolite	hbl	446 ± 11	0.19
Sierra de Quilmes	26°20′10″				
6/120	66°01′00″	Migmatite	bt	374 ± 8	9.39

found essentially the same features: migmatitic ortho- and paragneisses, with small amounts of locally layered amphibolite, schist and rare calc-silicate lenses.

Caleta Loa. The metamorphic rocks — layered migmatites and minor homogeneous metadiorites of upper amphibolite facies — are exposed for 25 km along the bottom of the Rio Loa Valley (area 1, Fig. 1; Wilke et al., 1997). In the west, the migmatites are intruded by the Mesozoic Coastal batholith, and in the east they are covered by volcanic rocks. The metamorphic unit is locally intruded by small gabbros and diorites of Mesozoic age and by a granite of late Paleozoic age (Maksaev, 1990). The Paleozoic granite also intrudes into Devonian to Carboniferous sediments (Breitkreuz, 1986).

Mejillones Peninsula (area 2, Fig. 1). Gneiss and schist of upper amphibolite facies with minor intercalations of garnet–amphibolite dominate the central part of the peninsula (Baeza, 1984). These rocks are similar to those of the areas described above, although migmatization is absent. The southern part, which is also separated from the central part by a NW–SE striking fault zone, consists of Mesozoic mafic intrusions and their metamorphosed equivalents (Lucassen and Franz, 1992, 1994).

2.2. NW Argentina

Figs. 1 and 2 show the distribution of major occurrences of high-grade metamorphic rocks of upper amphibolite facies in NW Argentina north of 26°S (Toselli et al., 1978; Allmendinger et al., 1982; Viramonte et al., 1993; Hongn, 1994; Becchio et al., 1999). There are also low-grade equivalents of the same lithologic units (Adams et al., 1989; Quenardelle, 1990; Gonzalez et al., 1991). Rock types of the metamorphic basement are similar to those of N. Chile with abundant quartz-rich metasediments and orthogneiss of granitoid composition, especially on the Puna plateau. Generally, aluminum-rich metapelite is more frequent than in N. Chile. Metabasites are rare in both areas. The foliation strikes around N–S, and dips to the west. In the Salar de Centenario and El Jote areas (localities 1 and 7 in Fig. 2) the metamorphic basement is thrust on Ordovician sediments and their low-grade equivalents during Ordovician deformation (Hongn, 1992, 1994; Mon and Hongn, 1988, 1991, 1996).

Salar Centenario (area 1 in Fig. 2).The metamorphic unit at the western border of the Salar consists of quartz-rich, sillimanite–andalusite–kyanite schist (Viramonte et al., 1976), granitic orthogneiss and subordinate deformed pegmatite, amphibolite, migmatite and staurolite–garnet-bearing schist. Borders between schist and orthogneiss are commonly zones of intensive deformation and mylonite formation under ductile conditions. Mafic dikes crosscut the foliation and are locally deformed and folded. Granitoid intrusions into the metamorphic unit caused contact metamorphism especially in the phyllite and greenschist facies rocks. The metamorphic unit continues to the south: monotonous orthogneisses with subordinate amphibolite and local migmatite are widespread at the eastern slope of the Salar de Ratones and along the Salar de Diablillos (area 2 in Fig. 2) where a unit with paragneisses and micaschists occurs at the southern end of the salar.

Salar de Hombre Muerto (area 3 in Fig. 2). The eastern slope of the salar and a small mountain range in the salar (Tetas de Pachamama) are comprised of ortho- and paragneisses grading locally into migmatite with tourmaline-rich pegmatitic pods, minor amphibolite, calc-silicate lenses and coarse-grained marble (dm–m scale). Mafic dikes crosscut the foliation of metamorphic rocks. At least two different generations of dikes occur with the older generation deformed and locally folded and the younger generation unaffected by ductile deformation. Lithologically similar rocks in the south of the salar are of lower metamorphic grade and migmatite is absent (details in: Aramayo, 1986; Quenardelle, 1990). East of the salar at Cerro Blanco (area 5 in Fig. 2), tourmaline, garnet and alumino-silicate-bearing metapegmatites are locally abundant in the paragneisses.

Salar de Antofalla (area 6 in Fig. 2). Discontinuous outcrops of schist, orthogneiss, migmatite and amphibolite occur at the western slope of the salar. Sierra de Campo Negro (Palma, 1990) is the largest occurrence of metamorphic rocks of this range. The metamorphic unit is intruded by granite and unconformably overlain by presumed Devonian sediments. The same sediments with metamorphic and granitic clasts at their bases unconformably overlie Ordovician sediments at Quebrada Honda (Allmendinger et al., 1982; Palma, 1990). The rock types of the metamorphic basement and their contact relations are very similar to those found in the Chilean Sierra de Moreno.

El Peñon–El Jote (area 7 in Fig. 2). The major metamorphic rock types in the Sierra de El Peñon are quartz-rich schist, orthogneiss (augengneiss) and minor pegmatite. Calc-silicate layers and marble lenses are <2 m in thickness. The lithologies at El Peñon are similar to those of Salar de Hombre Muerto and ascribed to one unit by Hongn (1992) and Viramonte et al. (1993). The rocks of El Jote area consist of a homogeneous fine-grained orthogneiss at Cerro El Jote with rare garnet-bearing amphibolite and schist. In its southern continuation the sequence comprises abundant layers of calc-silicate boudins in paragneiss and minor staurolite–garnet schists both restricted to small shear zones (m-scale).

Cafayate and northern Sierra de Quilmes (areas 8 and 9 in Fig. 2). Cordierite and garnet-bearing migmatitic para- and orthogneisses with mafic boudins (dm — 10 m) of two-pyroxene granulite (Toselli et al., 1978) occur at the eastern slope of the Sierra de Quilmes. Compositional layering and foliation strike in N–S directions. Locally, strongly deformed tourmaline- and garnet-bearing pegmatite layers (cm–m) comprise more than 20 vol% of the rock. In the north the transition in metamorphic grade from the high-grade rocks to the low-grade Puncoviscana Formation (Fig. 2) seems gradual and without major faults. Willner et al. (1985) suggested that the Puncoviscana Formation is the protolith of the high-grade rocks based on chemical composition. At Cafayate the metamorphic unit is intruded by granite with clearly intrusional contacts into migmatite (lit–par–lit cutting migmatitic textures; Steffen Büttner, Berlin, pers. commun.) in the southern part of the pluton. Ductile deformation also affected the granite with the same N–S striking direction as in the gneiss (Oyarazabal, 1988).

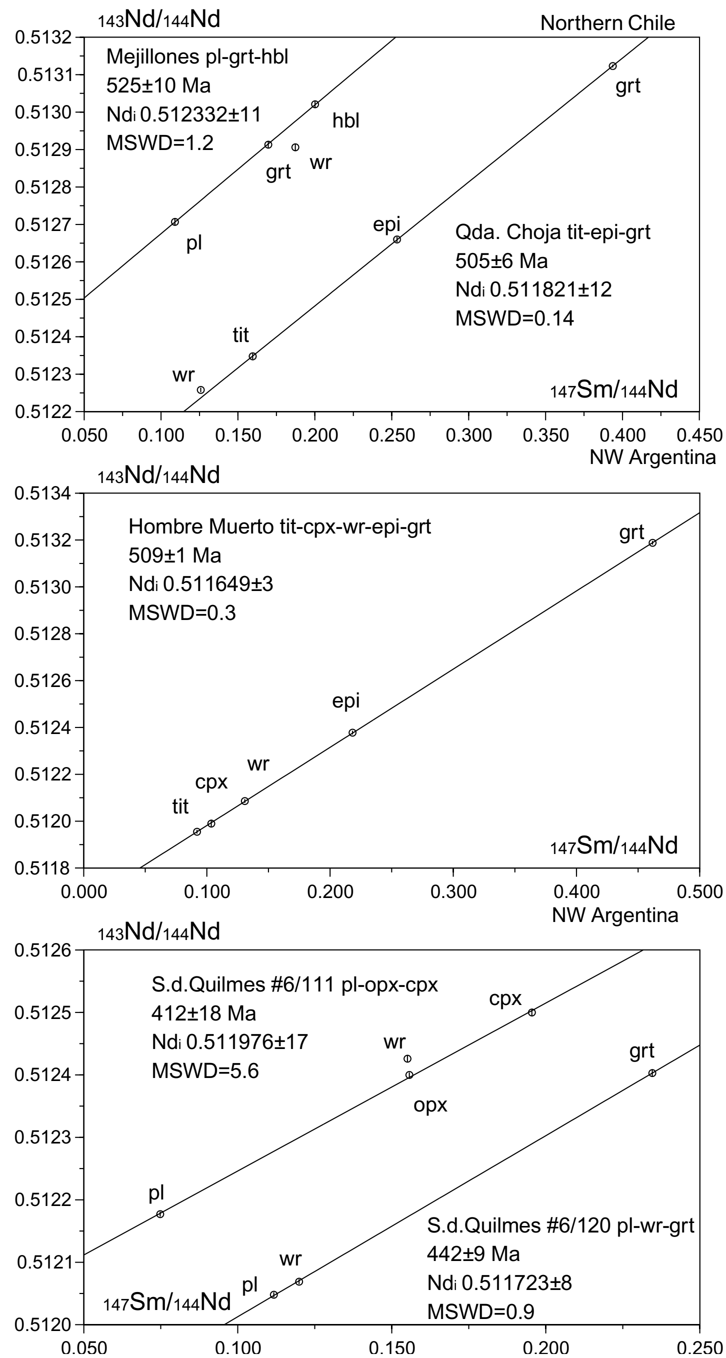

Fig. 3. Sm–Nd mineral isochrons for amphibolite, calcsilicate rocks, gneiss and mafic granulite (Table 1); see Figs. 1 and 2 for locations. Errors on ages and initial values (Nd_i) are given on the 95% confidence level (2 σ). The time of regional metamorphism (525–505 Ma) is consistent from the Chilean Coast Range to the Argentine Puna region, metamorphism further to the east was ca. 442 and 412 in the Sierras Pampeanas.

2.3. Metamorphic conditions

The mineral assemblage of the gneiss and migmatite from N. Chile and most areas of the Argentine Puna is quartz–plagioclase–biotite and minor garnet. The assemblage of the metabasite is hornblende and plagioclase and — rarely — garnet. Quantitative *P–T* estimation for the peak-metamorphism in the gneiss and metabasite is impossible at many sample sites. Migmatitzation shows minimum temperature above the granite solidus (Johannes and Holtz, 1996). Pressure was low as indicated by no or minor garnet in the gneiss and amphibolite. Upper amphibolite facies temperature (650–750°C) at 5–7 kbar pressure are reported from Belen (Lucassen et al., 1996; Lezaun et al., 1997) and Mejillones (Lucassen et al., 1996), both N. Chile, and from the Eastern Puna (Becchio et al., 1999) and Sierra de Quilmes (Toselli et al., 1978; Rossi de Toselli et al., 1987; Becchio et al., 1999), both NW Argentina.

Conditions of Early Paleozoic metamorphism in the Sierras de Cordoba at ca. 31°S show slightly higher pressure (ca. 8 kbar) at similar temperature (Rapela et al., 1998b).

We conclude from the presence of migmatite and the absence of any high pressure mineral assemblage, that maximum P–T conditions have been similar at the other locations. A detailed account to the metamorphic fabric, mineral chemistry and P–T data will be given elsewhere.

3. Geochronology

3.1. Sm–Nd dating

Sample preparation and analytical techniques are described in Appendix A. The method of Sm–Nd dating based on mineral isochrons is a reliable method to get information about the timing of near-peak conditions of high-grade metamorphism due to the high closure temperature of >650°C (Thöni and Jagoutz, 1992) or 600°C (Mezger et al., 1992). In many cases zircon shows several growth stages, monazite is generally absent in these rocks, and the Rb–Sr system could be severely disturbed by hydrothermal processes and would only record late stage cooling. Samples were selected (for areas see Figs. 1 and 2, for exact locations see Table 2) to get information about the oldest possible metamorphic crystallization event: a garnet–hornblende–plagioclase metabasite (Mejillones), a garnet–clinopyroxene–epidote–titanite calc-silicate (Sierra de Moreno), a garnet–clinopyroxene–epidote–titanite–calc-silicate (Salar Hombre Muerto), orthopyroxene–clinopyroxene–plagioclase metabasite and garnet–plagioclase gneiss (both Sierra de Quilmes). Minerals have been checked for compositional homogeneity by microprobe analysis and were found to be homogeneous. Sm and Nd concentrations, Nd isotope ratios, isochron ages and ages of separation from the depleted mantle are listed in Table 1 together with Sm–Nd systematics of additional whole rocks.

Regression lines of isochrons are presented in Fig. 3. Discarding the whole rock analysis from Mejillones and Sierra de Quilmes (Table 1) improves the mean squared weighted deviates (MSWD) dramatically, for the isochron from Quebrada Choja, exclusion of the whole rock sample yields a slightly better MSWD. The reason for this slight disturbance of the Sm–Nd system between whole rock and the major mineral assemblage is unknown. It may be speculated that e.g. during a static thermal overprint with temperature above the closure temperature of the K–Ar system in hornblende (as observed in the Mejillones sample, see below) or in biotite (Quebrada Choja sample, see below) new minerals such as epidote with high REE concentrations may form in small amounts. In all cases the absolute age is within the limits of error identical to the age calculated with the whole rock sample.

Similar ages as those from Mejillones, Sierra de Moreno and Hombre Muerto of ca. 505–525 Ma are also known from high-grade metamorphic rocks from the continuation of the Sierras Pampeanas to the south at Sierra Fiambalá (near 27°S; DeBari, 1994 and references therein), Sierra de Ancasti (28°S Fig. 1; Knüver, 1983), Sierra de Córdoba (ca. 525–520 Ma; Fig. 1 near 31°S, Rapela et al., 1998b), from N. Chile at Cordillera Frontal south of 30°S (Caminos et al., 1982), and the Belen area in N. Chile (Damm et al., 1990; Basei et al., 1996; Lezaun et al., 1997). The Early Paleozoic metamorphism is part of the Pampean Cycle, defined by Pankhurst and Rapela (1998) as post-Grenville to 490 Ma. Bachmann et al. (1986) suggested an early stage of metamorphism at ca. 550 Ma at Cumbres Calchaquies and Sierra de Ancasti (Fig. 1), which could be correlated with the ca. 560–540 Ma K–Ar ages of (very) low-grade metamorphism of the Puncoviscana Formation (Fig. 1) and equivalents (Adams et al., 1989). These ages of peak metamorphism coincide with the late phase (550–500 Ma) of the important Brasiliano–Pan-African orogeny which overprinted and formed many parts of the South American craton (e.g. Cordani et al., 1988; De Brito Neves and Cordani, 1991; Trompette et al., 1992; Van Schmus et al., 1995) as well as the African Proterozoic areas (e.g. review for for NE-Africa by Schandelmeier and Reynolds, 1997).

Lower intercept data of U–Pb zircon ages of a migmatite from Qda Choja (466 + 8/−7 Ma) and from an orthogneiss (415 + 36/−38 Ma) were interpreted as ages of high-grade overprint (Damm et al., 1990). The higher Sm–Nd age presented here is more likely near to the metamorphic age, because lower intercept ages of zircon with a complicated geological history are more difficult to interpret. There are no differences in lithology and metamorphic conditions between the northern and southern parts of Sierra de Moreno and the peak metamorphic event was probably at ca. 500 Ma in both areas. The Sm–Nd mineral isochron age of the sample from Salar Hombre Muerto (509 Ma) is confirmed by a U–Pb age of 508 ± 19 Ma from the same area (orthogneiss, single zircon method, Becchio et al., 1996), that could indicate the age of intrusion or metamorphism.

Sm–Nd isochron ages at Sierra de Quilmes are distinctively younger: ca. 442 Ma in a gneiss and ca. 412 Ma in a metabasite despite similar metamorphic conditions. Ages of high-grade metamorphism between ca. 410–450 Ma are also known from the Cumbres Calchaquies and from the Sierra de Ancasti (28°S; Bachmann et al., 1986; Miller et al., 1994). This metamorphism is part of the Famatinian Cycle (490 Ma to Mid-Carboniferous as defined by Pankhurst and Rapela, 1998) with prominent widespread magmatism at ca. 470 Ma (e.g. Rapela et al., 1990; Miller, 1996; Pankhurst et al., 1998). The ca. 440 and 410 Ma isotopic ages of high-grade metamorphism in Sierra de Quilmes are interpreted as a thermal overprint during the Famatinian Cycle which reset the Cambrian–Ordovician ages. The large error of 18 Ma for the 412 Ma isochron (same locality as the 442 Ma isochron) and the fact that the whole rock does not lie on the isochron points to an incomplete resetting. Ductily deformed granites (Rb–Sr whole rock age of 475 ± 7 Ma, Rapela et al., 1982), which intruded into the already deformed and migmatized high-grade basement ca. 30 km north of our sample location, testify the complex magmatic–metamorphic history. If the metamorphic basement of Sierra de Quilmes grades into the low-grade Precambrian–Lower Cambrian Puncoviscana Formation to the north (Fig. 1), this would give an upper age limit for the metamorphic evolution.

3.2. K–Ar dating

In order to constrain the final cooling of the rocks, K–Ar dating on biotite and hornblende was performed on identical samples or samples from the locations used for Sm–Nd dating, and from additional outcrops in Sierra de Moreno, Belen, Caleta Loa and the Puna.

We were also interested in evaluating the extent of isotopic resetting by later events (e.g. by granitoid igneous activity) and the potential effect on the Sm–Nd system. Results and exact sample locations are listed in Table 2, approximate positions are shown in Figs. 1 and 2. The observed spectrum of ages from the large area of investigation with its long geological history is wide and ranges from ~480 to 170 Ma. However, all data can be explained together with other isotope data in their geological context. The K–Ar ages indicate cooling from the peak temperature of metamorphism, because the closure temperatures of the Ar isotope system of 530°C in hornblende (Harrison, 1981) and 300°C in biotite (e.g. von Blanckenburg et al., 1989) are much lower than the peak metamorphic temperature. These cooling ages can be disturbed by reheating during magmatic activity with partial or complete resetting. All our samples follow a geologically meaningful sequence in their age relations, i.e. the ages of Sm–Nd are higher than the K–Ar ages, and K–Ar ages of amphibole are higher than those of biotite or similar, and it is therefore unlikely that excess Ar has to be considered as a major factor.

Sierra de Moreno (N. Chile). In the Quebrada Choja area the 478 Ma age of hornblende are close to the 500 Ma age of metamorphism and similar as a previously determined age of 452 Ma (K–Ar, Maksaev, 1990). However, K–Ar ages on hornblende from other locations in the same area are distinctly younger (422 and 406 Ma) as are K–Ar ages on mica (296 Ma, Table 2; 269 Ma, Maksaev, 1990). These ages were found in the vicinity of Late Paleozoic high level intrusions with K–Ar ages of 330–270 Ma (Maksaev, 1990) and 300 Ma (Rb–Sr dating; Lucassen et al., 1999b) in and south of the Quebrada Arcas. The ages of intrusions coincide with the biotite K–Ar age of the country rocks (micaschist and migmatite, samples 3/291 and 4/72, Table 2) close to the contact with granite or are distinctively younger than ages on mica from samples more distant to the granite outcrops (ca. 420 Ma; Maksaev, 1990). Those hornblende ages, which are intermediate between the Early Paleozoic metamorphic age and the Late Paleozoic magmatism are therefore interpreted as partially reset (samples 4/76, 3/288, 3/289, Table 2).

Belen and Tignamar (N. Chile, Fig. 1) The K–Ar ages of hornblende from Belen are 457 ± 7 Ma (amphibolite, sample 4/305, Table 2) and 358 ± 10 Ma (dioritic orthogneiss, sample 3/232, Table 2). Two U–Pb lower intercept ages of 456 ± 4 and 366 ± 3 Ma from the same area (Lezaun et al., 1997) are very similar compared with our K–Ar ages. The coincidence of the K–Ar and U–Pb lower intercept ages at Belen could indicate two thermal events above the closure temperature of hornblende. A K–Ar age from Tignamar, ca. 10 km south of Belen is 389 ± 11 Ma (hornblende from dioritic orthogneiss, sample 3/251, Table 2). Published K–Ar ages are between 417 and 365 Ma from the Belen and between 536 and 516 Ma from Tignamar (mineral type and errors not quoted; Basei et al., 1996). The broad scatter in the K–Ar ages reflects disturbance of the K–Ar system by post-metamorphic intrusional activity and low-grade, static metamorphic overprint indicated by the formation of abundant chlorite and sericite in many rocks of the area.

Chilean Coast Range (Mejillones peninsula, Caleta Loa). K–Ar ages (Table 2) of 151 ± 3, 170 ± 3, 189 ± 4 (biotite) and 152 ± 5 Ma (hornblende) are in the same range as previously published K–Ar ages of metamorphic rocks from the central block of Mejillones (whole rock 177 ± 8 Ma, hornblende 167 ± 8 Ma; Maksaev, 1990) and coincide with ages of cooling of the Mesozoic magmatic belt in the Coast Range (Maksaev, 1990; Scheuber et al., 1994; Lucassen and Thirlwall, 1998). They indicate a thorough reset of the K–Ar system in the Paleozoic metamorphic rocks and the identical ages of biotite and hornblende from Mejillones indicate fast cooling.

NW argentina. The K–Ar ages of all samples from the Puna high plateau are less variable than those from N. Chile. They cluster around 400 Ma (Table 2) and a few similar ages are reported in the literature: the K–Ar age of muscovite from a gneiss from Sierra Campo Negro at Salar de Antofalla is 419 ± 8 Ma and of granite from the same location is 417 ± 8 Ma (Kraemer et al., 1999), in accordance with our K–Ar age of 446 ± 11 Ma on hornblende of amphibolite from the same area. South of Sierra Campo Negro a K–Ar age of granite is 389 ± 8 Ma (muscovite; Kraemer et al., 1999). The lowest age of 371 ± 8 Ma obtained so far is from a secondary muscovite in El Jote (sample 6/142, Table 2); biotite from the same sample and amphibole from El Jote have a similarly high age like the other samples.

Only one age from the Sierra de Quilmes (Rio Anchillo-Quebrada de Quilmes) was obtained up to now. The age on biotite (374 ± 8 Ma) of the same migmatite sample, which yielded a Sm–Nd age of 442 Ma, is identical to the young muscovite age of El Jote. A similar K–Ar age of 383 ± 5 was found in the Puncoviscana Formation (Adams et al., 1989) and in granites of Sierra de Quilmes (395–373 Ma; Toselli et al., 1978).

3.3. Nd model ages of the basement

We present single stage t_{DM} ages (Table 1; for locations see Figs. 1 and 2) assuming a linear evolution of the ^{143}Nd in the mantle to a present day $^{143}Nd/^{144}Nd$ ratio of 0.51316 and $^{147}Sm/^{144}Nd$ ratio of 0.214 (Goldstein et al., 1984). Calculated t_{DM} ages using this model are ca. 10% higher for the given ages than ages calculated using the nonlinear model of DePaolo (1981).t_{DM}

The t_{DM} model ages for felsic metmorphic rocks from the metamorphic basement range from 1.36 to 2.2 Ga (Table 1) without a gradient in their regional distribution. In N. Chile the average is 1.65 ± 0.16 Ga (1σ; $n = 12$). Most metabasite samples have younger t_{DM} ages than the gneiss: at Mejillones 1.45 Ga (Table 1) and 0.85–1.14 Ga (recalculated data of Damm et al., 1990), at Sierra de Limón Verde 0.75 and 0.82 Ga (Table 1; Lucassen et al., 1999a) and at Cordón de Lila 0.86–1.4 Ga (recalculated data of Damm et al., 1991). At Belen ages of metabasite are between 1.91–2.05 and 0.67–1.14 Ga (recalculated data of Damm et al., 1990; additional ages of 1.54 and 1.75 Ga are reported by Basei et al., 1996, from the Belen schists at 18°S, but the method of calculation is not quoted). The average t_{DM} age of both gneisses, including additional locations of the Sierras Pampeanas (localities I–III, Fig. 1, Table 1), and metabasite in NW Argentina is 1.76 ± 0.4 Ga (1σ; range 1.55–2.2 Ga; $n = 22$; Table 1). The same t_{DM} ages of 1.6–1.7 Ga are reported from Ordovician sediments of the northern Puna by Bock et al. (1998; method of calculation not quoted).

The average t_{DM} ages of the gneiss samples are indistinguishable at the 67% confidence level in all areas. We interpret the protolith of the high-grade metamorphic rocks as sediments and crustal melts mainly derived from Mid-Proterozoic sources. The Paleozoic gneiss between northwestern Argentina and the Chilean Coast appears isotopically homogeneous. Also, some of the metabasite samples have Mid-Proterozoic t_{DM} ages. Various Early Paleozoic intrusions from N. Chile and NW Argentina between 18 and 24°S have Sr and Nd isotope ratios that are indistinguishable from those of the high-grade basement (Damm et al., 1990, 1994). In summary, the majority of the basement exposures consists of material added to the crust before the Late Precambrian. Additions of juvenile material, e.g. during the activity of possible Late Proterozoic–Early Paleozoic magmatic arcs, were certainly of minor importance and may be found only in the distinctively younger t_{DM} of some metabasite samples. Rapela et al. (1998b) and Pankhurst et al. (1998) reported numerous Sm–Nd isotope data from the Sierras de Córdoba (ca. 30–32°S; metamorphic and intrusive rocks from the Pampean Cycle, intrusive rocks from the Famatinian Cycle) and the average t_{DM} of 1.78 ± 0.18 Ga (1σ; range 1.48–2.18 Ga; $n = 46$; recalculated to the Goldstein et al., 1984, model, excluding three samples with ages >2.2 Ga and six samples with ages ~1 Ga or smaller) is in excellent agreement with our results from the north.

4. Discussion

Fig. 4 gives an overview of the various magmatic–metamorphic events in the region during the Proterozoic and Paleozoic. Despite the large uncertainties of Sm–Nd t_{DM} model ages due to model assumptions and to possible changes in the Sm/Nd ratio during a complicated metamorphic–magmatic history, gneiss, migmatite and granitoids of the Early Paleozoic high-grade basement from N. Chile to southern Sierras Pampeanas are isotopically homogeneous without regional gradients. The significance of the model ages may be evaluated using available U–Pb zircon ages from the area and the agreement of these ages is rather good. Ages quoted below are upper or lower discordia intercept ages. The oldest ages of ca. 1.7–2.0 Ga are from Uyarani (Bolivia), Belen (N. Chile) and the Arequipa Massif (S. Peru; Fig. 1; Lezaun et al., 1997; Wasteneys et al., 1995) interpreted by the respective authors as ages of the protoliths. Grenvillian protolith ages and ages of metamorphism (0.97–1.2 Ga) are known from the same locations and from the Bolivian Altiplano (Tosdal, 1996); Litherland et al. (1989) also reported protolith ages of ca. 2 Ga and magmatism and metamorphism between 1.3 and 1.0 Ga from the Proterozoic basement of eastern Bolivia (ca. 14–19°S; 63–58°W).

The Pampean Cycle (Pankhurst and Rapela, 1998) with its high-grade metamorphism at ca. 525–500 Ma was the major event of crustal consolidation. Relics of an older metamorphism, such as the Grenvillian age of ca. 1 Ga, which is confirmed for the area north of 18°S (Wasteneys et al., 1995; Tosdal, 1996; Lezaun et al., 1997), are restricted to the U–Pb ages in zircon. Relic zircons from the Early Paleozoic metamorphic basement of N. Chile (1.2 Ga, Quebrada Choja, Damm et al., 1990) and NW Argentina (1.1 Ga, Hombre Muerto, Becchio et al., 1996; Figs. 1 and 2) support the incorporation of Grenvillian material into the sedimentary and igneous protolith. Such old material is probably also recycled during Cenozoic magmatism in the Andean crust (Wörner et al., 1994).

The peak metamorphic conditions of the Pampean Cycle reached temperatures high enough for migmatization in many ares, and the Sm–Nd mineral isochrons should be close to the ages of crystallization of the high-grade metamorphic assemblages. It is unlikely that the high temperatures above the granite solidus were sustained over a long period of time with a possible strong diffusional reequilibration of the isotope system, because metamorphism occurred at mid-crustal levels (5–7 kbar) in the whole area and considering the Paleozoic equilibrium geotherm not beeing significantly higher than recent equilibrium geotherms. In this shallow level, cooling is rather rapid due to the steep thermal gradient in the upper crust.

The age data from Sierra de Quilmes (NW Argentina) are distinctly younger at ca. 442–412 Ma (Fig. 4) than in N. Chile and in the Puna high plateau. This Silurian (= Famatinian Cycle according to Pankhurst and Rapela,

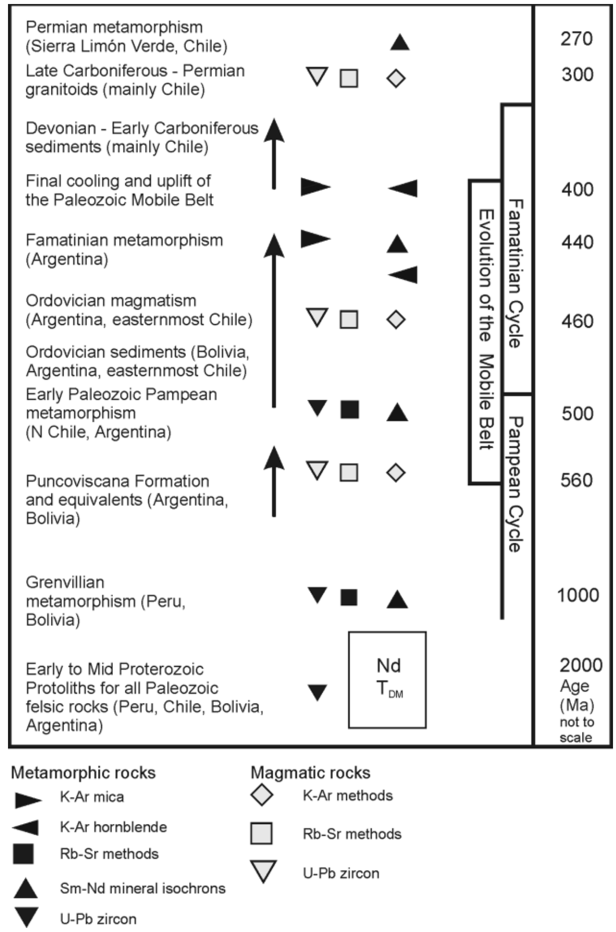

Fig. 4. Principal sedimentary, magmatic and metamorphic events known from the Premesozoic basement in the Central Andes in S. Peru, N. Chile, SW Bolivia and NW Argentina. Only the peaks of magmatic and metamorphic development are indicated. Isotopic ages and the range of depositional ages (indicated by arrows) are from the references in the text and our work; compare to Fig. 2 in Rapela et al., 1998a.

1998) high T–low P metamorphism certainly overprinted the Pampean metamorphism as shown by the intrusion of a ca. 475 Ma old granite (Rapela et al., 1982) into already existing migmatites. An investigation of the timing of this polymetamorphic development and the related late Ordovican–Silurian granitic magmatism is currently undertaken (Büttner et al., 2000). In the context of the present paper, we want to restrict the interpretation to the fact that the high-grade metamorphism is Early Paleozoic and not Precambrian, which was already suggested by Toselli et al. (1978), who first reported granulite facies metamorphism from the Sierra de Quilmes.

K–Ar ages on hornblende and biotite (Table 2) give a rough temperature–time scale for the cooling history according to the respective closure temperatures and to possible resetting by later thermal events. The data confirm the resetting due to the widespread and long lasting magmatic activity in the Central Andes. In the Coast Range the K–Ar system in biotite and hornblende is reset to 150–190 Ma during Jurassic contact metamorphism. In the Chilean Precordillera most hornblende ages are in the range between 420 and 300 Ma, biotite ages between 300 and 270 Ma, and all these ages are probably influenced by the major igneous activity around 300 Ma (Lucassen et al., 1999b). At these locations in NW Chile, lower intercepts of the U–Pb ages of zircon are between 460 and 400 Ma (and younger) and referred as the ‘Paleozoic thermal overprint’. However, there are also older K–Ar ages as in Quebrada Choja (Sierra de Moreno, Fig. 1) and west of this area (Quebrada Arcas; Maksaev, 1990) and in Belen (Basei et al., 1996; Lezaun et al., 1997). Here we assume that the Late Paleozoic thermal overprint was less important. Also, in the outcrops of NW Argentina, where plutonism at around 300 Ma is much less abundant, the hornblende ages are between 450 and 400 Ma, biotite ages between 420 and 400 Ma. Cooling seems later at 370 Ma in Sierra de Quilmes, consistent with the distinctly later high-grade overprint at 440–420 Ma.

Summarizing all K–Ar results we conclude that the rocks passed the closure temperature of ca. 500°C for hornblende at 480–450 Ma and the closure temperature of ca. 300°C for biotite at ca. 400 Ma at many locations. This means that the basement must have been exhumed close to the present erosional surface between 400 and 300 Ma (Fig. 4), which

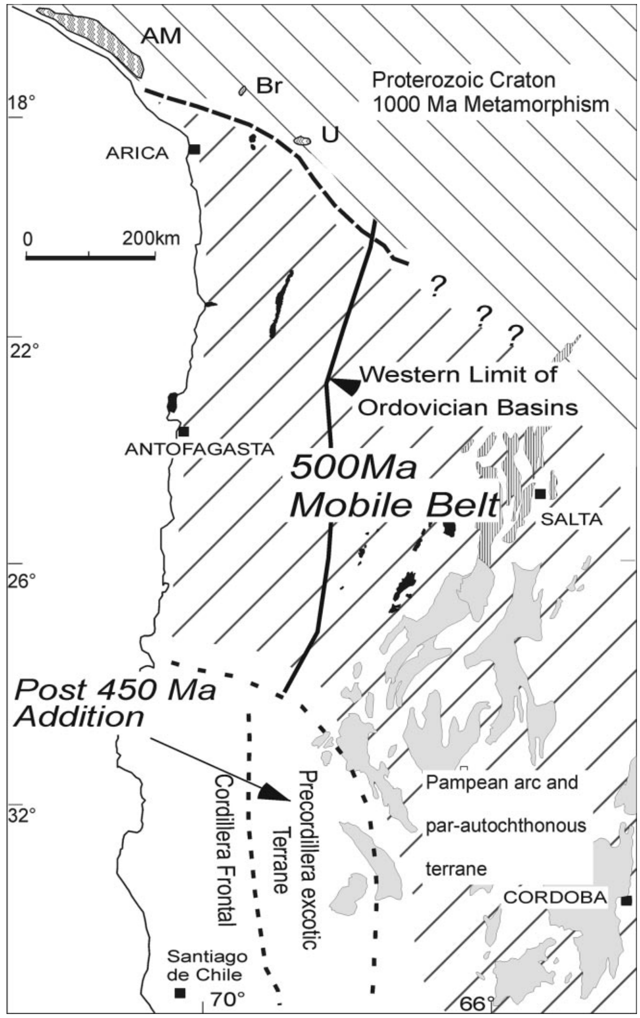

Fig. 5. Basement domains in the Central Andes on the basis of their age relations. The northern Proterozoic cratonic domain with Arequipa (AM), Berenguela (Br), Uyarani (U) is separated (dashed line, with unknown extension towards the south–east) from a ca. 500 Ma old mobile belt. To the south, boundaries of the mobile belt to the adjacent areas are interpreted as terrane boundaries (see text). The 'Pampean par-autochtonous terrane and arc' (Rapela et al., 1998a,b) lies in the continuation of the Puncoviscana Formation. The westernmost Sierras Pampeanas are probably part of the Argentine Precordillera (see Pankhurst and Rapela, 1998). The approximate 'western limit of Ordovician basins' might represent the border of the morphological high parts of the mobile belt west of this line that could have been a source region of the Ordovician sediments.

is confirmed by Devonian, Carboniferous and Permian erosional unconformities between gneiss and sediments (Quebrada Arcas area, Breitkreuz, 1986; Belen, Lezaun et al., 1996) in N. Chile. Also, at Caleta Loa Early Paleozoic migmatite as well as Devonian to lower Carboniferous sediments were intruded by the same Carboniferous (ca. 320 Ma) pluton. Therefore the migmatite unit must have been near to the surface at this time. The same is true for NW Argentina, where an erosional unconformity of presumed Devonian age is described from the Salar de Antofalla area (Allmendinger et al., 1982; Palma, 1990). Reheating by the late Paleozoic magmatism at approximately 300 Ma was significant but locally variable, and partly reset the K–Ar system. This was the last major thermal influence of regional extent in the basement of the Chilean Precordillera and the Argentine Puna plateau. In contrast, in the Coastal Cordillera no old K–Ar data were found up to now, indicating that during the Jurassic magmatism the K–Ar system was completely reset.

5. Conclusions

The observed petrological similarities and the age relations in the high-grade basement between 21 and 26°S constrain the possible geodynamic scenarios for the Late Proterozoic to Early Paleozoic development of what represents the present Andean continental margin, and what was the continental margin of Gondwana. It is concluded that the high *T* metamorphic rocks with ages of metamorphism between ca. 525 and 500 Ma formed during one major metamorphic–magmatic cycle (broadly referred to as the

Brasiliano Cycle, or more specifically the Pampean Cycle) from Mid-Proterozoic protoliths. In N. Chile these rocks border against a realm with prevailing Grenvillian metamorphism (Fig. 5), which is considered as a cratonic area. Some rocks from this area have Late Precambrian–Cambrian ages of ca. 600–520 Ma (Dalmaryac et al., 1977; Litherland et al., 1989), indicating a Pampean overprinting of the craton. Therefore the boundary is not a terrane boundary, but a distinction between the Grenvillian and Paleozoic metamorphosed parts of a compositionally homogeneous Mid-Proterozoic protolith. West of the present Pacific Coast line there could have been a continental margin (see Fig. 6), the continuation of the mobile belt or the cratonic counterpart of the mobile belt. In the basement between 18 and 26°S *mineral parageneses* of Grenvillian ages of metamorphism are unknown and all Grenvillian ages are from relic zircons. Following this argumentation, the postulation of an Arequipa–Antofalla Craton (numerous references up to present e.g. Ramos, 1988, 1995; Coira et al., 1998; Omarini et al., 1999) is meaningless because it does not take into account the Pampean metamorphism that substantially reworked the preexisting rocks. Therefore, we suggest a distinction between a *cratonic area* and a *mobile belt* on the basis of metamorphic ages (Fig. 5). The possible border of the mobile belt towards the south against the 'Precordillera Exotic Terrane' with a presumeable docking at ca. 450 Ma (Pankhurst and Rapela, 1998) and the Cordillera Frontal are interpreted as terrane boundaries (dotted signature; "Pampean parautochtonous terrane and arc"; Rapela et al., 1998a,b). The westernmost Sierras Pampeanas are probably part of the Argentine Precordillera.

This interpretation also inhibits the accretion of various exotic or parauthochthonous terranes (in the sense of formation or existence of oceanic crust, with subsequent subduction of this oceanic crust and collision of the microcontinent) for the area between 18 and 26°S, with ages becoming younger to the west (from Late Precambrian to Carboniferous (?) e.g. Ramos, 1988 counts 4 terranes, his Fig. 1; Bahlburg and Hervé, 1997 count 3 terranes, their Fig. 12). Recently proposed terrane boundaries (Ramos, 1995) are shown in Fig. 1. We think that the accretion of magmatic arcs (e.g. Ramos, 1988; Conti et al., 1996; Dalziel, 1997; Omarini et al., 1999) seems to be unlikely for the area between 18 and 26°S, because the typical geological indicators for suture zones are absent; there is no undoubted evidence for the existence of former oceanic crust. Mafic rocks are rare and if present, they are not related to oceanic crust but form part of the Pampean metamorphic basement. In the ancient environment basaltic–andesitic magmatism was relatively unimportant. Typical alpine-type ultramafic rocks are absent. For most of the rare mafic–ultramafic rocks in the area between 18 and 26°S (Coira et al., 1998, on the ultramafic–mafic suites of the Puna) reliable age constraints and geochemical data are not available. New geochemical data by Bahlburg et al. (1997) from the Puna strongly suggest that the ultramafic rocks are not ophiolites. Mon and Hongn (1991) had already pointed out that due to structural observations the age of these 'ophiolites' is probably Precambrian and questioned the terrane concept of Ramos et al. (1986).

From the viewpoint of metamorphic ages there is no reason to separate the basement of Mejillones peninsula as a 'Mejillonia terrane'. Also, there seems no reason to separate a Pampean terrane (distinguished in Fig. 1 as the

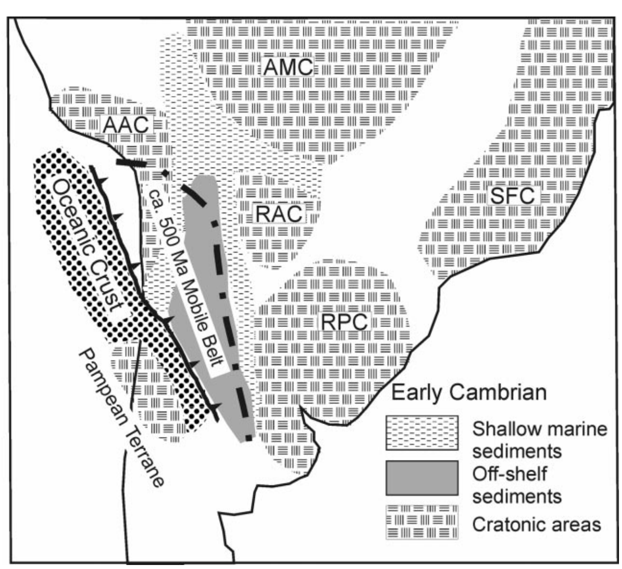

Fig. 6. Possible geodynamic scenario of the Early Cambrian (modified after Rapela et al., 1998a,b) with the distribution of cratonic areas (Grenvillian or older; De Brito Neves and Cordani, 1991) and position of the later mobile belt, which formed during the Early Paleozoic and completely reworked the southern part of the hypothetical Arequipa–Antofalla craton (AAC). Shallow marine and off-shelf Early Cambrian sediments age were deposited in the Puncoviscana basin (Durand and Aceñolaza, 1990; Omarini et al., 1999). A subduction zone and oceanic crust north of the Pampean Terrane cannot be identified and are therefore hypothetically placed west of the present coast line. Other cratonic areas are: AMC, Amazonian; RAC, Rio Apa; SFC, San Francisco; RPC, Rio de La Plata.

'Pampean craton'), because the difference to the area west of the proposed terrane boundaries is an overprint of the ca. 500 Ma metamorphism by a ca. 440 Ma metamorphism of similar metamorphic conditions. This Silurian overprint may actually be more widely distributed, because it is also indicated in the basement of the Puna (Hombre Muerto and Salar de Diablillos), where mafic dikes crosscut the foliation of migmatites, but are also ductilely deformed. It might be also the cause of nearly identical K–Ar ages in hornblende and biotite at some locations in the Puna (Table 2). Extent and uplift history of the 440 Ma metamorphism need further investigations, but the thermal overprint is explained easiest with the widespread Famatinian magmatism. Also the correlation of the very low-grade metamorphism in the Puncoviscana Formation (Fig. 5) with the Silurian metamorphism, and if these rocks represent the unmetamorphosed equivalents of the high-grade rocks in their continuation to the south, needs more data (see Omarini et al., 1999, for their concept of the Puncoviscana formation).

Another argument against the terrane concept and for a mobile belt are the conditions of metamorphism. High *P*/low *T* metamorphism indicative for subduction–accretion regimes is absent, and the whole area shows rather uniform low- to intermediate *P*/high *T* conditions. This widespread type of metamorphism of the Pampean to Famatinian Cycle is important for any speculations about the geodynamic setting. Such a metamorphism at mid-crustal depth must be linked to a prominent thermal anomaly of considerable extent. Assuming a similar history in a collision zone as outlined for the Sierras de Córdoba by Rapela et al. (1998a,b), this zone must have been at least ca. 600 km wide with unknow extent to the west and to the east. The thermal gradient, which produces 700°C at 15–20 km depth must have been very high considering that temperatures at 20 km depth estimated from continental heat flow data are between 220°C for shield areas to a maximum of 500–580°C for rifting situations (Chapman, 1986). A conductive transport of heat which causes the 700°C would require an unrealistically high geotherm, where large parts of the lower crust and the mantle are already molten. Advective heat transfer by intrusion of melts is a more realistic model for high *T* mid-crustal metamorphism (e.g. Wells, 1980; Barton and Hanson, 1989). In the investigated area the widespread, long standing magmatic activity, derived from crustal melts, as indicated by the isotopic composition of pre-, syn- and post-deformational intrusions of Cambrian to Devonian age in NW Argentina (Rapela et al., 1990, 1992, 1998a,b; Galliski et al., 1990; Damm et al., 1990, 1991, 1994) points to its important contribution to the mid-crustal heat budget. A common model to explain the mid-crustal low *P*/high *T* metamorphism, migmatization and the generation of crustal melts is magmatic underplating of mantle derived melts at the base of the lower crust (e.g. Sandiford and Powell, 1986; Bohlen, 1987; Harley, 1989).

In a possible geodynamic situation in the Cambrian (Fig. 6) we assume a continental margin with a subduction zone with the formation of a broad belt of thickened crust and magmatic activity, the latter mainly recycling the pre-existing crust. The crustal thickening at this active margin would follow the incomplete rifting, which formed the Puncoviscana Basin. The Puncoviscana Basin comprises passive margin sediments in the south, which implies the existence of oceanic crust (Rapela et al., 1998b) and deep marine sediments continuing to the north into south Boliva grading into platform sediments further to the north (paleogeographic reconstruction after Durand and Aceñolaza, 1990). The N–S extension of the deep marine sediments mirrors the extension of the Pampean mobile belt (Fig. 6). The isostatic exhumation of such a broad belt of homogeneously thickened crust could explain the similarities of the exposed *P*–*T*-t section of the Early Paleozoic orogen. Our concept would not contradict the accretion of allochthonous or parauthochthonous terranes in the south, where ultramafic rocks of an ophiolitic sequence in the Sierras de Córdoba were described from the Pampean Cycle (Pankhurst and Rapela, 1998, and references therein).

Also, this concept is in line with the information about sedimentation in the Ordovician: the development of the sedimentary basins in NW Argentina and Bolivia (e.g. Moya, 1988; Bahlburg and Breitkreuz, 1991; Bahlburg and Hervé, 1997) is contemporaneous to the cooling of the metamorphic basement in N. Chile and and the Argentine Puna (Fig. 4). The thick infill points to prominent erosion in the source region. Sm–Nd t_{DM} ages from the sediments (Bock et al., 1998; NW-Argentina) are identical to those from the metamorphic basement. Ordovician sediments are rare in N. Chile (Bahlburg and Breitkreuz, 1991) and principal directions of transport in the basins are from the west and south to the north. A morphological high in the area of N. Chile, which forms the western limit of these Ordovician basins (Fig. 5) is likely and assumed in reconstructions of sedimentary basins (e.g. in Tankard et al., 1995).

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Appendix A. Sample preparation and analytical techniques

Sm–Nd analyses. Minerals were separated using a magnetic separator followed by hand picking. Hornblendes and pyroxenes have been leached in 2.5 M HCl and garnets in 6 M HCl for 20 min prior to dissolution. Dissolution of the samples in HF and HNO_3 was carried out in screw top Savillex® beakers at 180°C. After conversion to nitrates, complete dissolution was checked by centrifugation in 10% HNO_3. Nd isotope and REE analyses were carried out using the VG 354 multicollector mass spectrometer at the University of London Radiogenic Isotope Laboratory at Royal Holloway. An aliquot of the solution was analyzed for REE concentrations by the mass spectrometric isotope dilution technique of Thirlwall (1982) modified for the multicollector. The reproducability is better than 1% for the Sm and Nd contents and 0.1% for the Sm/Nd ratio. Nd was separated by conventional ion-exchange techniques and the isotope ratios were measured as Nd oxide following the mass spectrometric procedure of Thirlwall (1991). The laboratory standard gave $^{143}Nd/^{144}Nd$ of 0.511423 ± 7 (2σ; $n = 69$; samples #3/379, #4/164, #4/327, #6/120) and 0.511419 ± 9 (2σ; $n = 12$) during the periods of measurement, equivalent to La Jolla standard value of 0.511857. Values of $^{143}Nd/^{144}Nd$ were normalized to $^{42}Nd/^{144}Nd = 1.4187$, equivalent to $^{146}Nd/^{144}Nd = 0.72190$.Regression lines of isochrons (Fig. 4) were determined according to the method of York (1969). For MSWD >1 errors were multiplied by the square root of the MSWD. A decay constant of $6.54 \times 10^{-12}\ a^{-1}$ for Sm is used.

Whole rock samples analyzed for Nd isotopes at Zentrallabor für Geochronologie, Universität Münster and GFZ Potsdam: analytical method see Lucassen et al. (1999c) and Romer and Heinrich (1998), respectively.

K–Ar dating. Minerals were separated magneticaly and by handpicking under the binocular microscope. Purified biotite was ground in pure alcohol to remove alterated rims that might have suffered 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 Herrmann (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 ^{40}Ar was determined by isotope dilution method using a highly enriched ^{38}Ar 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 2).

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