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~990 Ma peak granulitic metamorphism and amalgamation of Oaxaquia, Mexico: U–Pb zircon geochronological and common Pb isotopic data

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ABSTRACT

The ~1 Ga Oaxacan Complex is the largest exposure of the Oaxaquia terrane of Mexico. Microprobe analysis of co-existing ortho- and clino-pyroxene, garnet, ilmenite and plagioclase indicate metamorphic conditions of 735 ± 5° C and 7.7 ± 0.1 kbar in the granulite facies, which was dated between ~998 and ~979 Ma using U–Pb isotopic analyses of zircon populations. Peak temperatures >800° C and isobaric cooling are indicated by the sporadic presence of sapphirine in metasediment and garnet–cordierite–sillimanite–K-feldspar–rutile assemblages, and garnet coronas around pyroxene and titanomagnetite. Common Pb isotopes from acid-leached whole-rock samples define an errorchron that intersects the Stacey/Kramers crustal growth curve at 1,187 ± 63 Ma indicating a major crust-forming event at this time, and suggesting that the granulite facies metamorphism did not significantly reset the common Pb isotopic composition. These data are comparable with those from other ~1 Ga inliers in Mexico and is consistent with a single Oaxaquia terrane by ~1 Ga.

Key words: U-Pb geochronology, Oaxaquia, Grenvillian Orogeny, granulite metamorphism.

RESUMEN

El Complejo Oaxaqueño constituye el afloramiento de rocas grenvillianas más extenso de México, y el núcleo del microcontinente Oaxaquia. Las condiciones de P y T, obtenidas por análisis de microsonda en la asociación orto y clinopiroxeno, granate, ilmenita, plagioclasa y cuarzo, se han calculado en 735 ± 5° C y 7.7 ± 0.1 kbar, y las texturas sugieren que estos valores representen un enfriamiento isobárico desde un pico de más alta temperatura estimado en más de 800° C. Los fechamientos de U–Pb en poblaciones de zircones indican que el metamorfismo granulítico tuvo lugar en un arco de tiempo comprendido entre ~979 y ~ 998 Ma. Análisis isotópicos de Pb común en polvos de roca total, previamente lixiviados con ácidos, definen una isocrona de 1,187 ± 63 Ma que indica un evento mayor de formación cortical, y sugiere que los isótopos de Pb común no fueron rehomogeneizados durante el metamorfismo granulítico. Los datos de Pb común del Complejo Oaxaqueño son comparables con datos similares de otros macizos grenvillianos en Oaxaquia sugiriendo la existencia de un único terreno a ~ 1 Ga.

Palabras clave: geocronología de U-Pb, Oaxaquia, orógeno Grenville, facies de granulita.

INTRODUCTION

The Oaxacan Complex of southern Mexico represents the most extensive outcrop (~10,000 square km) of ~1 Ga basement that is inferred to underlie the backbone of Mexico, which was grouped into a terrane called Oaxaquia by Ortega-Gutiérrez *et al.* (1995) (Figure 1A). Similar rocks appear to extend beneath the Chortis block of Guatemala, Honduras, El Salvador and northern Nicaragua (Donnelly *et al.*, 1990; Manton, 1996; Nelson *et al.*, 1997). Recent papers (Keppie *et al.*, 2001, 2003, and in press; Solari *et al.*, 2003; Cameron *et al.*, in press) presented structural, metamorphic, geochemical and isotopic data, which suggest Oaxaquia–Chortis lay adjacent to Amazonia at ~1 Ga.

To shed further light on Oaxaquia and its Precambrian tectonics, we present additional U–Pb and common Pb isotopic and geochemical data obtained at the Laboratorio Universitario de Geoquímica Isotópica (LUGIS), Universidad Nacional Autónoma de Mexico (UNAM). Our results allow a better definition of the high-grade metamorphism in the northern Oaxacan Complex and indicate that U–Pb and common Pb isotopic analyses performed at LUGIS are comparable to those from other laboratories.

GEOLOGICAL SETTING

The Oaxacan Complex north of Oaxaca, in southern Mexico (Figure 1), is composed of two major thrust slices: (1) the lower Huitzo unit composed of anorthosite, mangerite, garnet-bearing charnockite, gabbros (AMCG), and Fe-diorite with U–Pb protolith ages of ~1,012 Ma (Keppie *et al.*, 2003; Solari *et al.*, 2003) that intrudes the El Catrín migmatitic gneiss unit with U–Pb protolith ages older than ~1,350 Ma (Solari *et al.*, 2003); and (2) the upper El Marquez unit made up of paragneiss intruded by intraplate charnockite, syenite and gabbro with U–Pb protolith ages between ~1,130 and ~1,230 Ma (Keppie *et al.*, 2003). The El Catrín unit was migmatized at ~1,100 Ma in the Olmecan teconothermal event, and subsequently all the rocks were affected by the dominant granulite facies, Zapotecan tectonothermal event that involved at least two phases of folding (Solari *et al.*, 2003). This was followed by low amphibolite to greenschist facies shearing and several other phases of folding, which culminated with NW-trending, low grade late folds (Solari *et al.*, in press). The Zapotecan event was constrained between the U–Pb igneous age of ~1,012 Ma of the AMCG suite and the ~978 Ma, crystallization age of late tectonic pegmatites, not affected by granulite-facies metamorphism (Solari, 2001; Solari *et al.*, 2003; Keppie *et al.*, 2003). Keppie *et al.* (2003) also argued that the U–Pb upper intercepts age of 999 ± 9 Ma on an anorthosite sample and concordant U–Pb ages on a metagabbronorite and garnet-bearing charnockite of 990 ± 4 and ~1,001–1,004 Ma, respectively, may date the granulite metamorphism. Other published age data in the northern Oaxacan Complex are a garnet–whole rock, Sm–Nd age of ~963 Ma (Patchett and Ruiz, 1987), an U–Pb titanite age of ~968 (Keppie *et al.*, in press), and cooling ages provided by $^{40}Ar/^{39}Ar$ ages of 977 ± 12 Ma and 945 ± 10 Ma on hornblende and micas and hornblende, respectively (Keppie *et al.*, in press; and Solari *et al.*, 2003), and K–Ar analyses ranging between ~927 and 775 Ma on hornblende, mica and K-feldspar (Fries *et al.*, 1962; Fries and Rincón-Orta, 1965).

Mora and Valley (1985) and Mora *et al.* (1986), by using feldspar thermometers and the garnet–plagioclase barometer, calculated peak temperatures and pressures of 710 ± 50º C at 7–8 kb for the granulite-facies metamorphism in the same area studied in this work. Granulite metamorphism in the northern Oaxacan Complex is characterized by an assemblage of anhydrous minerals such as pyroxenes, garnet, quartz, perthitic potassic feldspar, plagioclase and ores, which are often accompanied by high temperature hydrous phases such as brown hornblende and titaniferous biotite. Hydrated phases are generally secondary, produced during retrogression and cooling, with formation of amphibole pseudomorphs that replace pyroxenes, generally tremolite–actinolite–anthophyllite–cummingtonite, green hornblende and micas such as fine-grained biotite, rare muscovite and sericite. Titanite–uralite–biotite coronas are also common around ilmenite and titanomagnetite.

The Oaxacan Complex is unconformably overlain by Tremadocian and late Paleozoic sedimentary rocks, which were deformed at sub-greenschist facies (Centeno-García and Keppie, 1999) before deposition of Jurassic red beds, Cretaceous limestones and Tertiary lacustrine deposits and continental volcanic rocks. Several Late Paleozoic intrusions are also present (Solari *et al.*, 2001).

ANALYTICAL TECHNIQUES

One of the aims of this work was test the capabilities of the LUGIS to date small multi-grain zircon populations and perform common Pb whole-rock analyses. U–Pb zircon analytical methods used at LUGIS have been described by Solari *et al.* (2001) and will not be repeated here. At the time of this work, LUGIS could not obtain a mixed $^{205}Pb/^{235}U$ spike (acquired in 2003), so a mixed $^{208}Pb/^{235}U$ spike was used instead. This does not allow very small amounts of material to be analyzed, however three of the samples (6498, OC9810 and 66C98) were previously analyzed by Keppie *et al.* (2003) and Solari *et al.* (2003). The results obtained here allow comparison of the $^{205}Pb/^{235}U$ and $^{208}Pb/^{235}U$ methods. Furthermore, two fractions of the zircon standard (Ontario 91500: Wiedenbeck *et al.*, 1995) were also analyzed to check our laboratory reproducibility. Our analyses of the zircon standard 91500 are in good agreement with data published by Wiedenbeck *et al.* (1995) in both U and Pb concentrations, as well in the obtained U/Pb ratios and ages (Table 1, Figure 2). Repeated analyses allow us to

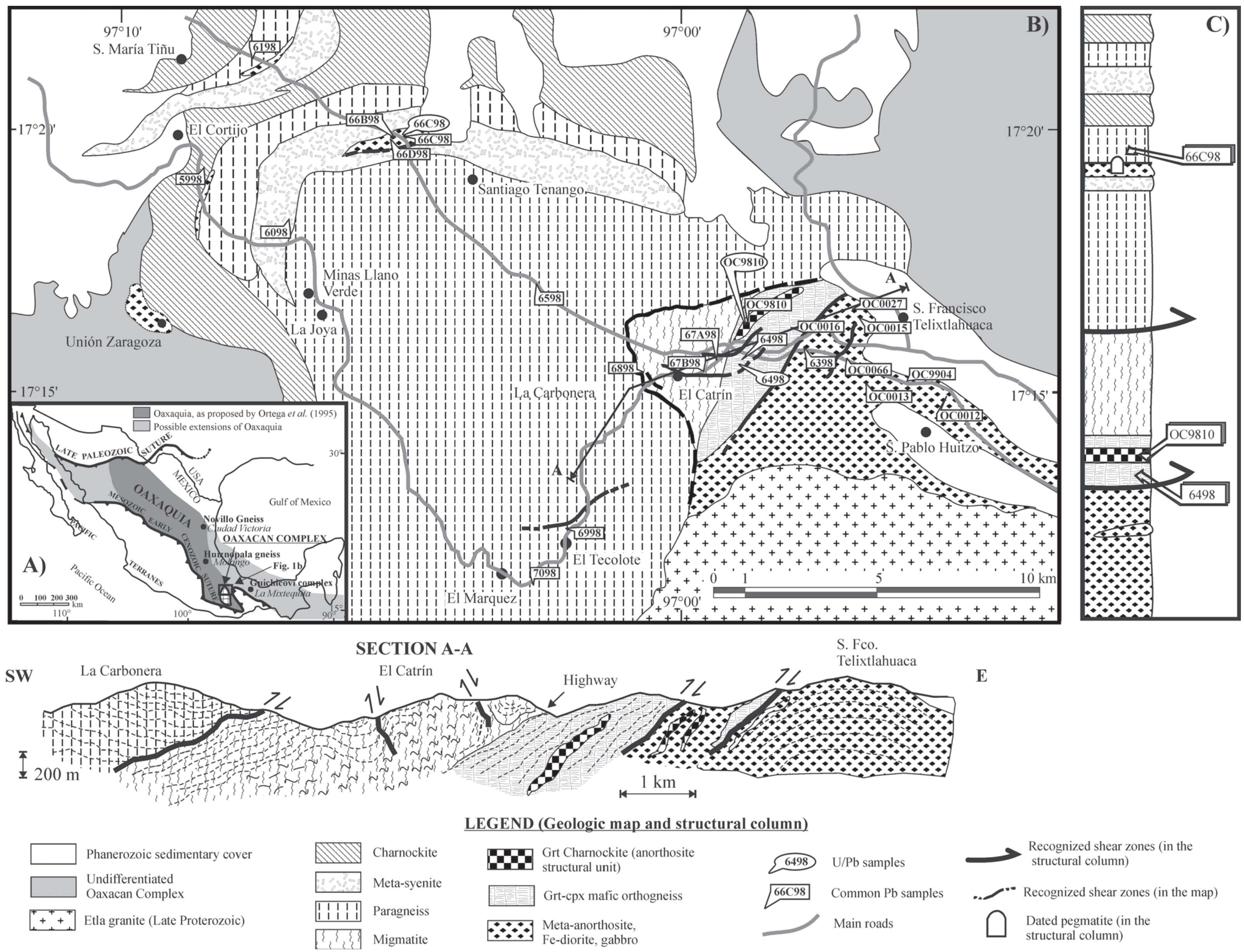

Figure 1. Locality map of Mexico. A: Oaxaquia, as proposed by Ortega-Gutiérrez *et al.* (1995), shaded in dark gray, with the possible extensions in light gray, including the Guichicovi gneiss; B: Geological map of the studied area in the northern Oaxacan Complex; C: Structural column of the Northern Oaxacan Complex showing the sample localities and the main thrust faults. Modified from Solari *et al.* (2003).

Table 1. U–Pb zircon analyses of samples from the northern Oaxacan Complex, southern Mexico, and for the standard Zircon 91500.

	Weight §	**U §**	**Total Pb §**	**Com. Pb**	**^{206}Pb/^{204}Pb**	**^{207}Pb/^{206}Pb**	**^{208}Pb/^{206}Pb**	**^{206}Pb/^{238}U**	**^{207}Pb/^{235}U**	**^{207}Pb/^{206}Pb**	**^{206}Pb/^{238}U**	**^{207}Pb/^{235}U**	**^{207}Pb/^{206}Pb**	**% Disc.**
Fraction‡	mg	ppm	ppm	pg	Observed Ratios†			Atomic Ratios††			Age (Ma)†††			
6498 Grt-px bearing orthogneiss														
A, ov-stby, 100-200 mm	0.5	48	8	150	876	0.0859	0.1151	0.16555	1.64779	0.07219	988	989	991 ± 4	0.4
B, rnd, > 200 mm	0.3	59	10	0.5	2354	0.0722	0.0699	0.16561	1.64503	0.07204	988	988	987 ± 8	-0.1
OC9810 Charnockite														
A, sh prsm, 100-200 mm	0.6	99	4	170	383	0.1048	0.1628	0.16688	1.66657	0.07243	995	996	998 ± 9	0.3
66C98 Late-tectonic pegmatite														
66C98, brk xls, > 300 mm	1.8	210	4	81	2695	0.0770	0.1781	0.16350	1.61762	0.07176	976	977	979 ± 3	0.3
Standard zircon 91500, Ontario, Canada *														
91500-208A	1.392	68	13	290	2055	0.0818	0.1244	0.17840	1.84543	0.07502	1058	1062	1069 ± 2.3	1.0
91500-208B	3.35	69	13	62	2389	0.0802	0.1222	0.17917	1.85278	0.07500	1062	1064	1068 ± 1.4	0.5
Published simple mean	–	81.2	15	–	–	–	–	0.17917	1.85021	0.07488	1062 ± 0.4	–	1065.4 ± 0.3	–

Zircon sample dissolution and ion exchange chemistry modified after Krogh (1973) and Mattinson (1987) in Parrish (1987)-type microcapsules.
‡ All diamagnetic fractions at 2.0 Amp. rnd: round; sh prsm: short prismatic to stubby grains; ov-stby: ovoid to stubby grains; brk xls: broken crystals.
Numbers refer to the micrometric size of the fraction chosen for analysis.
§ Concentrations are known at ± 30%, due to the weight error.
† Observed isotopic ratios are corrected for mass fractionation of 0.12% for ^{208}Pb spiked fraction.
Two sigma uncertainties on the ^{207}Pb/^{206}Pb and ^{208}Pb/^{206}Pb ratios are < 0.8%, generally better than 0.1%; uncertainties in the ^{206}Pb/^{204}Pb ratio vary from 0.1% to 2.4%.
†† Decay constants used: ^{238}U = 1.55125x 10^{-10}; ^{235}U= 9.48485x 10^{-10}; ^{238}U/^{235}U= 137.88.
Uncertainities on the U/Pb ratio is 0.5%.
††† ^{207}Pb/^{206}Pb age uncertainties are 2 sigma and from the data reduction program PBDAT of K. Ludwig (1991). Total processing Pb blank were <100 pg.
* Zircon standard 91500 published simple mean is from analyses of Wiedenbeck *et al.* (1995). The published ^{207}Pb*/^{206}Pb* apparent ages range between 1,064 ± 2 Ma and 1,067 ± 1 Ma, and the ^{206}Pb*/^{238}U ages range between 1,060 ± 1.6 Ma and 1,065 ± 1.4 Ma. All errors on the LUGIS data are 2σ. Published uncertainties for simple mean of Wiedenbeck *et al.* (1995) are 1σ.
Initial Pb composition are from isotopic analyses of feldspar separates.
Isotopic data were measured on a Finnigan MAT 262 mass Spectrometer with SEM Ion Counting at UNAM, Mexico City.

reproduce, during this study, U/Pb ratios with an error of ± 0.4 to 0.5 %. We adopt, conservatively, the higher value of 0.5%.

Microprobe analysis of one sample (OC9810) was carried out on a Cameca SX 100 microprobe at the American Museum of Natural History, New York, with current setting of 15 kV and 10 nA and counting times ranging between 20 and 60 s (Table 2). Glasses and natural mineral were used as standards to monitoring the analyses.

Seven whole-rock samples were analyzed for their Pb isotopic compositions at LUGIS (Table 3) with the methods described by Schaaf *et al.* (submitted). Other 15 whole-rock common Pb analyses of samples of the northern Oaxacan Complex and three from the Novillo Gneiss (Ciudad Victoria, Tams., Figure 1A) were performed at UCSC using methods descibed by Lopez *et al.* (2001).

ANALYTICAL RESULTS

U–Pb geochronology

Sample 6498 from the Huitzo unit is an ore-enriched mangerite composed of hypersthene, ferroaugite, garnet, perthitic alkalic feldspar, quartz, abundant zircon, and cumulates of ilmenite, magnetite and apatite. Geochemical analysis of this sample shows it to have an intraplate character (Keppie *et al.*, 2003). Zircons extracted from this rock are rounded to stubby–short prismatic (2:1 to 3:1 aspect ratio). Fraction A is composed of 35 light pink, gem quality, ovoid to stubby grains (Figure 3). Fraction B is composed of 25 rounded and flat grains (Figure 3). Cathodoluminescence (CL) observations reveal the absence of a clear internal oscillatory zoning, typical of igneous growing, for both ovoid-stubby and rounded zircons (Figure 4). It is thus inferred they crystallized during granulite metamorphism. This is consistent with the rounded terminations and the shapes of these zircons under binocular microscope (Figure 3), and the low U content (Table 1), which are typical attributes of granulite-facies zircons crystallized in a mafic rock (Doig, 1991; Aleinikoff *et al.*, 1996). Consequently, both fractions yielded low $^{206}Pb/^{204}Pb$ ratios, and are almost concordant with $^{207}Pb/^{206}Pb$ ages of 991 ± 4 and 987 ± 8 Ma (Table 1, Figure 5) that are interpreted as dating the granulite-facies metamorphism. Previous results using single grains or a few grains yielded slightly discordant results with an upper intercept age of 1,012 ± 12 Ma that was interpreted as dating the time of intrusion of the mafic gneiss protolith (Keppie *et al.*, 2003).

Sample OC9810 from the Huitzo unit is a garnet-bearing charnockite that consists of perthitic feldspar, oligoclase, quartz, hypersthene, ferroaugite, garnet (20% grossular, 74% almandine), and accessory ore minerals such as ilmenite and magnetite, and zircon. Geochemical analysis of this sample indicates an intraplate character (Keppie *et al.*, 2003). Zircons separated from this sample are clear, colorless to light pink, short prismatic with an aspect ratio of up to 3:1 (Figure 3). Under CL they show uniform, low luminescence and an absence of inherited cores or oscillatory zoning, and thus appear to have crystallized during metamorphism. The isotopic analysis of the 35 selected zircon population yielded a concordant age with a $^{207}Pb/^{206}Pb$ age of 998 ± 9 Ma (Figure 5). Previous results obtained from single grains gave concordant ages of 1,004 ± 3 Ma and 1,001 ± 8 Ma, which were interpreted as dating the granulite facies metamorphism (Keppie *et al.*, 2003).

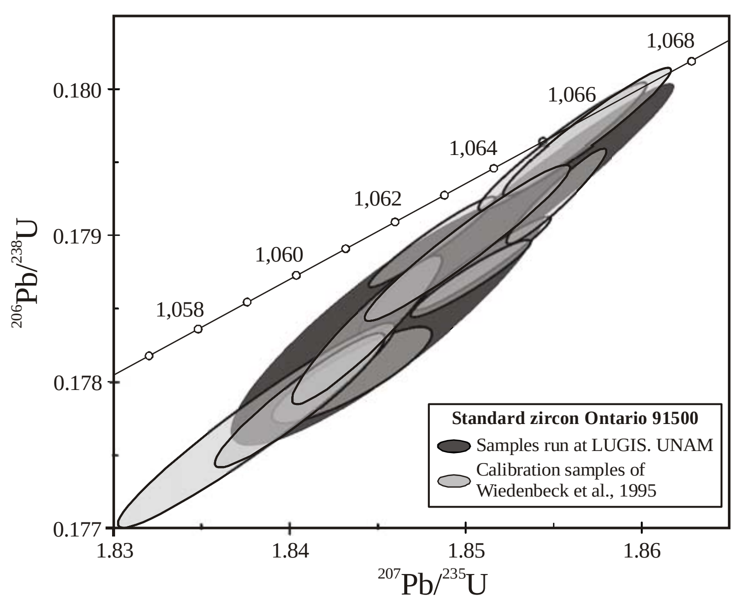

Figure 2. Concordia diagram with the analyses performed on the 91500 Ontario standard zircon in this work, compared with results obtained by Wiedenbeck *et al.* (1995).

Table 2. Microprobe analyses for the metamorphic paragenesis of the charnockite OC9810. Cationic recalculation performed with the program CPX (Berman, 1991).

	Opx	**Garnet**	**Cpx**	**Ilmenite**	**Plagioclase**
SiO_2	46.15	36.69	48.23	0.11	61.59
TiO_2	0.11	0.07	0.15	49.97	0.00
Al_2O_3	0.29	20.00	1.01	0.00	23.45
FeO	46.95	33.58	25.73	49.03	0.18
MnO	0.53	1.22	0.27	0.19	0.00
MgO	4.12	0.63	3.56	0.76	0.00
CaO	0.71	7.34	19.57	0.61	5.24
Na_2O	0.04	0.03	0.56	0.00	8.51
K_2O	0.01	0.02	0.01	0.00	0.30
Total	98.907	99.602	99.075	100.67	99.297
	[X_{Mg}] 0.1347	[Gr] 0.2075	[Mg M1] 0.1876	[X_{Mg}] 0.0282	[An] 0.2495
	[X_{Fe}] 0.8616	[Py] 0.0247	[Fe M1] 0.7618	[X_{Fe}] 0.9678	[Ab] 0.7337
	[X_{Al2}] 0.0037	[Alm] 0.7406	[Al M1] 0.046	[X_{Ti}] 0.9396	[Or] 0.0168
		[Sp] 0.0272	[Ti M1] 0.0045		
			[Ca M2] 0.8502		
			[Mg M2] 0.0191		
			[Fe M2] 0.0777		
			[Na M2] 0.0438		

Zircons separated from a late tectonic pegmatite (sample 66C98) are large, pink to red to dark amber, and generally consist of angular fragments of broken crystals (Figure 3). The dated fraction was composed of 15 clear fragments, about 500 µm in size, without any visible inclusions, cracks or alteration. These zircon fragments show under CL thin, oscillatory zoning, that, combined with higher U contents (Table 1), suggest formation during magmatic crystallization. The concordant analysis gave a $^{207}Pb/^{206}Pb$ age of 979 ± 3 Ma (Figure 5), which is interpreted as the crystallization age. This age is undistinguishable from a previously reported age of 983 +11/-5.5 Ma (U–Pb, discordia upper-intercept on zircons; Solari *et al.*, 2003).

Comparison of U–Pb geochronology using different Pb spikes indicates that they produce comparable results for the granulite facies metamorphism in the Grenvillian rocks of southern Mexico, and for intrusions that post-date the peak metamorphism, such as the pegmatite 66C98. However, because the use of a ^{208}Pb spike requires a large zircon population, it does not allow discrimination between the earlier intrusive event and the granulite facies metamorphic event. ^{208}Pb spike is thus adequate for determination of igneous or metamorphic events that are not overprinted by a later, high-grade tectonothermal event that recrystallized zircon.

Microprobe analyses relating to metamorphism

One of the dated samples (OC9810) was also analyzed by microprobe to determine the composition of mineral paragenesis (Table 2).This sample was selected because it is a fresh charnockite with low lost on ignition (LOI = 0.37%; Keppie *et al.*, 2003). Garnet has a uniform composition both in the large, unzoned crystals (≤1 mm), as well as in the thin coronas (50–250 µm) around titanomagnenite (Figures 6A and B). The clinopyroxene and/or orthopyroxene (Figures 6C and D) generally occur between the latter minerals and plagioclase and also forms symplectites with quartz (*e.g.*, Figure 6A) or clinopyroxene (*e.g.*, Figure 6C and D). These textures indicate: (1) qualitative metamorphic reactions such as:

Orthopyroxene + Plagioclase
= Garnet + Quartz

Titanitomagnetite + Plagioclase (+ Quartz)
= Garnet + Ilmenite

Orthopyroxene + Plagioclase
= Garnet + Clinopyroxene

and (2) an isobaric cooling path for the granulites of the northern Oaxacan Complex during high-grade metamorphism. With the mineral compositions (Table 2) for the assemblage garnet–plagioclase–ilmenite–clino-pyroxene–orthopyroxene–quartz and the TWQ program, version 2.02, of Berman (1991) values for equilibrium temperature and pressure of 735 ± 5º C and 7.7 ± 0.1 kbar, respectively, were calculated.

Common Pb

Present-day Pb isotopic compositions of samples from the northern Oaxacan Complex and Novillo Gneiss fall into the three lithologic groups: igneous rocks, metasediments, and pegmatites (Table 3, Figure 7). The igneous rocks of the Huitzo and El Marquez units, the Novillo Gneiss, and the El Catrín leucosome and igneous melanosome show a

Table 3. Whole-rock common Pb isotopic compositions of selected samples from the northern Oaxacan Complex, southern Mexico, and the Novillo Gneiss, Ciudad Victoria, northeastern Mexico.

Sample	Rock description	U†	Th†	Pb†	WR measured ratios (present-day ratios)			Age (Ma) (ref.) ††	
					$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$		
Northern Oaxacan Complex									
Huitzo AMCG unit									
OC9904	Ferronorite	ND	ND	ND	17.7517	15.5341	36.9911	1,010	(1)
OC0012	Garnet charnockite	ND	ND	ND	17.6230	15.5217	36.7623	1,010	(1)
OC0015	Granitic orthogneiss	ND	ND	ND	17.2528	15.4667	36.5134	1,010	(1)
6398	Ferronorite	ND	ND	ND	18.3800	15.6140	37.6096	1,010	(1)
OC0027	Gabbro	ND	ND	ND	17.3167	15.4873	36.6655	1,010	(1)
OC0066	Anorthosite	ND	ND	ND	17.4317	15.5141	36.8071	1,010	(1)
6498	Garnet mangerite	0.76	0.98	3.03	18.9941	15.6964	37.6510	1,010	(1)
OC9810	Garnet charnockite	0.42	0.58	9.26	17.5545	15.5387	36.6513	1,010	(1)
OC0016	Charnockite	ND	ND	ND	17.9470	15.6750	37.7032	1,010	(1)
El Catrín Migmatite unit									
67A98	El Catrín melanosome	0.67	8.05	21.50	17.4399	15.5102	37.6259	1,350	(3)
67B98	El Catrín leucosome	0.11	0.09	5.63	17.4820	15.5260	36.6616	1,100	(3)
6898	Pegmatite	ND	ND	ND	17.2991	15.5028	36.5534	1,100	(3)
El Marquez unit									
6598	Calcsilicate	6.09	13.13	7.25	35.3009	16.9341	43.9717	970	(2)
OC0013	Mafic metasediment	ND	ND	ND	21.2774	15.9143	38.7800	>1,100	
6998	Pegmatite	3.84	1.8	12.82	27.4080	16.2260	37.9500	980	(3)
7098	Paragneiss	1.15	1.83	4.57	22.9116	15.9712	39.6513	>1,100	
66B98	Pegmatite	0.95	0.39	17.4	17.3164	15.5464	36.9352	1,125	(3)
66C98	Pegmatite	3.37	49.57	17.26	21.1037	15.8251	42.0623	980	(3)
66D98	Pegmatite	0.89	4.3	17.63	18.5420	15.6330	38.5130	980	(3)
6198	Garnet paragneiss	1.73	2.1	18.59	17.3260	15.5485	36.6265	980	(3)
5998	Charnockite	0.57	1.18	19.62	17.2011	15.5087	36.5448	1,150	(1)
6098	Syenite orthogneiss	0.23	0.89	18.26	17.1154	15.5000	36.6570	1,150	(1)
Novillo Gneiss									
4177	Metagranitoid	0.28	2.63	12.99	17.9740	15.5730	37.6610	–	
4178	Metagranitoid	0.17	5.55	17.60	17.5590	15.5360	37.4490	–	
4183	Garnet granulite	0.06	0.34	2.39	17.2710	15.4890	36.7930	–	

† Concentrations in ppm determined by ICP-MS at UCSC, following Lopez *et al.* (2001). ND: not determined.
†† Crystallization ages are calculated by U-Pb, in the following works (references in parenthesis): (1) Keppie *et al.* (2003); (2) Solari (2001); (3) Solari *et al.* (2003).
Ages of the metasediments are inferred minimum ages, whereas for sample 67B98 is the age of migmatization. For the Novillo Gneiss, no crystallization ages for the selected samples are available. 2-sigma errors on isotopic ratios are <0.02%, generally better than 0.015% ($^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$), and <0.05% ($^{207}Pb/^{204}Pb$).

limited range of isotopic values (Table 3), which implies a common magmatic origin. On the other hand, the metasediments show extremely high and variable values of common Pb isotopes (Table 3) probably due to the large variation in age of the rocks in the source area. The common Pb isotopes in the pegmatites are also high and variable probably due to their high U/Pb ratios (Table 3).

In the $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram (Figure 7B) the northern Oaxacan Complex and Novillo samples fall in a coherent array, which scatter around an errorchron of 1,155 ± 110 Ma that intersects the Stacey and Kramers (1975) average crustal-growth curve at 1,138 Ma. This may be interpreted as the age of the main crust-formation event. It is similar to the first magmatic event dated at ~ 1,130 – ~1,230 Ma in the northern Oaxacan Complex (Keppie *et al.*, 2003).

DISCUSSION AND CONCLUSIONS

Comparisons with other ~1 Ga inliers in Mexico

Our new geochronological data indicate that granulite-facies metamorphism in the northern Oaxacan Complex occurred at about 998–988 Ma confirming earlier ages of 1,004–990 Ma (Keppie *et al.* 2003). These ages are similar to those recorded elsewhere for the ~1 Ga granulites of Mexico: (1) 988 ± 5 Ma in the southern Oaxacan Complex

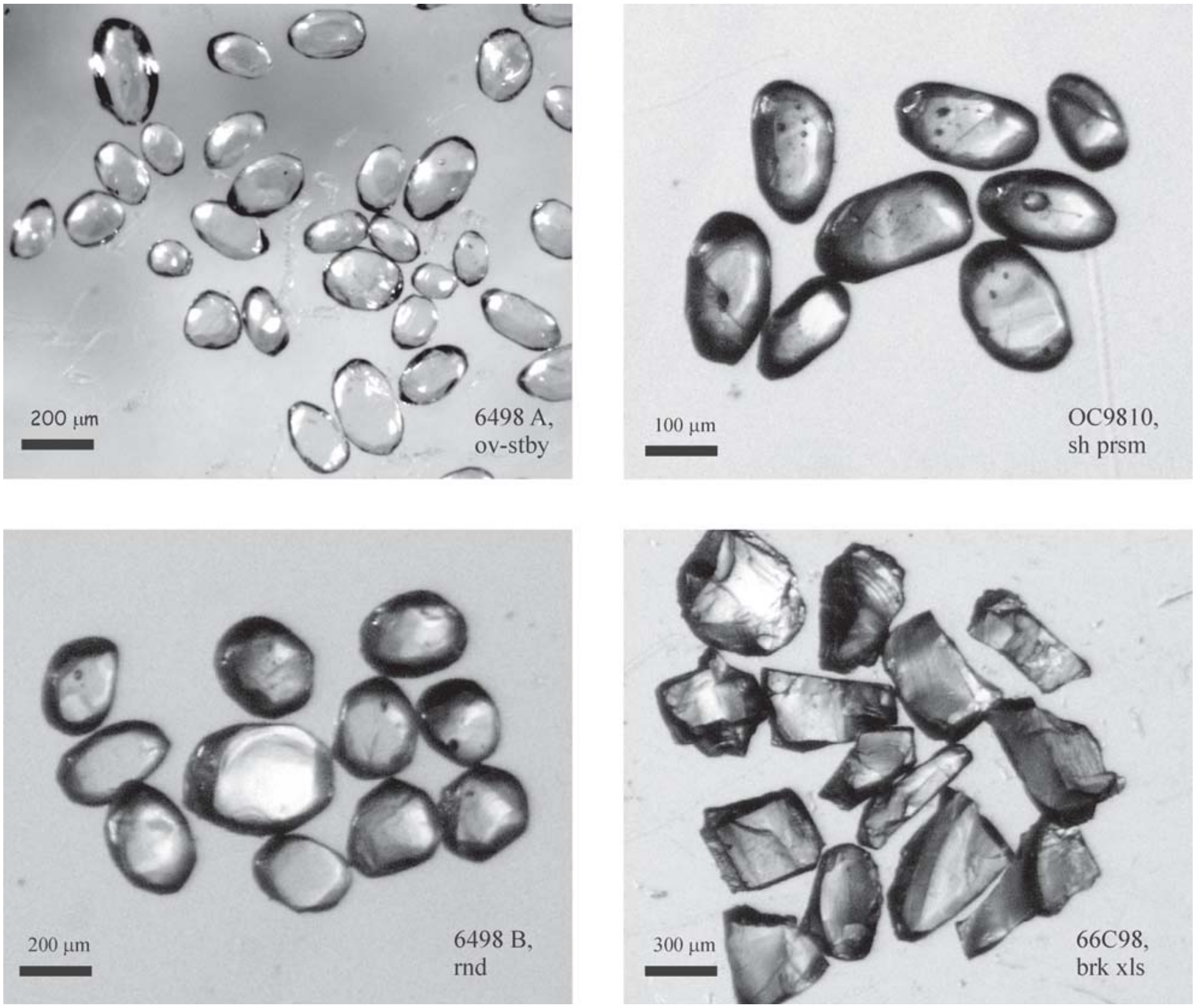

Figure 3. Photomicrographs of the zircon populations selected for U–Pb analyses. Pictures taken under stereomicroscope, with both reflected and transmitted light. Scale bar provided on each image. Ov-stby: oval to stubby grains; sh prsm: short prismatic grains; rnd: rounded grains; brk xls: broken fragments of larger crystals.

(Keppie *et al.*, 2001); (2) 986 ± 4 Ma (Ruiz *et al.*, 1999) and 975 ± 36 Ma (U–Pb lower-intercept age; Weber and Köhler, 1999) in the Guichicovi Complex (Figure 1A); (3) ~985 Ma in the Novillo Gneiss (Cameron *et al.*, in press).

Our results also indicate that peak metamorphism reached pressures and temperatures of 7.7 ± 0.1 kbar and 735 ± 5º C. These values confirm earlier results from the northern Oaxacan Complex: 7.0 ± 1 kbar and 770 ± 55º C using the garnet–plagioclase barometer, and 815 ± 85º C, and 710 ± 40º C using the independent feldspar thermometers (Mora and Valley, 1985; Mora *et al.*, 1986). Higher temperatures (>800º C) are indicated by the sporadic presence of sapphirine in metasediments and garnet–cordierite–sillimianite–K-feldspar–rutile assemblages in metapelites. On the other hand, the lack of kyanite suggests a similar pressure. These observations indicate isobaric cooling within the granulite facies.

The P–T estimates are similar to those reported from other ~1 Ga granulite facies rocks in Mexico: (1) ~850º C and 7.4 ± 0.1 kbar in the Guichicovi granulites (Murillo-Muñetón and Anderson, 1994); (2) 7.2 ± 0.5 kbar and 725 ± 50º C, and similar isobaric cooling textures in the Huiznopala gneiss (Lawlor *et al.*, 1999); (3) 8.9–9.7 kbar and 729–773º C in garnet cores and 642 ± 33º C and 7.9 ± 0.5 kbar in garnet rims in the Novillo Gneiss (Orozco-Esquivel, 1991).

Common Pb isotopes have sometimes been used to distinguish terranes of Laurentian and Amazonian origin (Tosdal, 1996), and this led Ruiz *et al.* (1999) to draw a line along the Trans-Mexican Volcanic Belt separating Novillo and Huiznopala of Laurentian affinity from Oaxaca and Guichicovi of Amazonian affinity. However, this distinction is unclear because, although the Huiznopala data are less radiogenic, the Novillo data overlap those from the

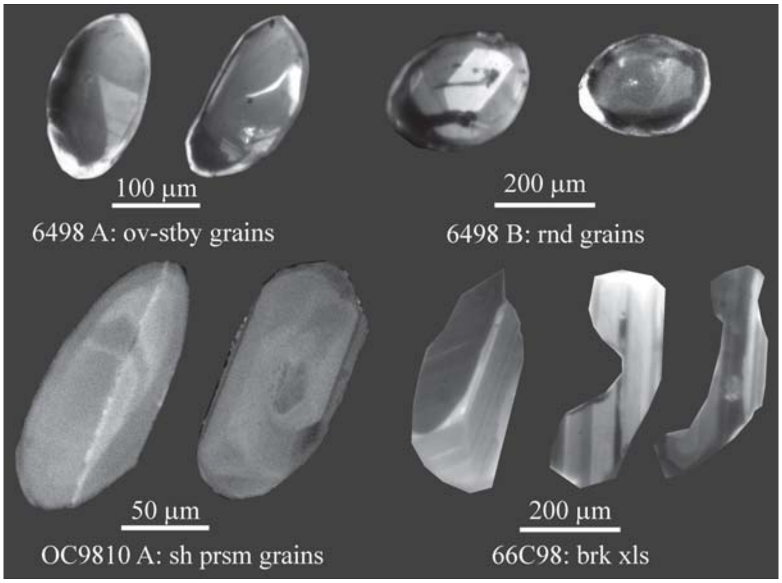

Figure 4. Cathodoluminescence images of zircons selected from the four samples described in the text. Scale bar provided on each image.

northern Oaxaca Complex and Guichicovi (Figure 8). An explanation for such variation was provided by Cameron *et al.* (in press), who showed that common Pb isotopic compositions of feldspars from igneous protoliths in these Mexican ~1 Ga inliers define a linear array between ~1,200 Ma and ~1,000 Ma, which is interpreted by those authors as the result of mixing of two different isotopic reservoirs. This implies that the magmas had access to both source regions during their genesis, and the differences in the common Pb isotope compositions reflect variable amounts of the end components in the magmas rather than different terranes.

The reference errorchron at 1,155 ± 110 Ma (Figure 7B) for the northern Oaxacan Complex is similar to the 1,250 ± 50 Ma reference errorchron for the Guichicovi Complex (Ruiz *et al.*, 1999). Combining all the data for Oaxaquia yields a reference isochron of 1,187 ± 63 Ma that may roughly date the major crust-forming event. These data are in good agreement with the magmatic ages obtained so far in the Huiznopala gneiss (1,150–1,200 Ma; Lawlor *et al.*, 1999), in the northern Oaxacan Complex (1,130–1,230 Ma; Keppie *et al.*, 2003), in the southern Oaxacan Complex (>1,117 Ma; Keppie *et al.*, 2001), and in the Guichicovi (upper intercept age of 1,231 ± 43 Ma; Weber

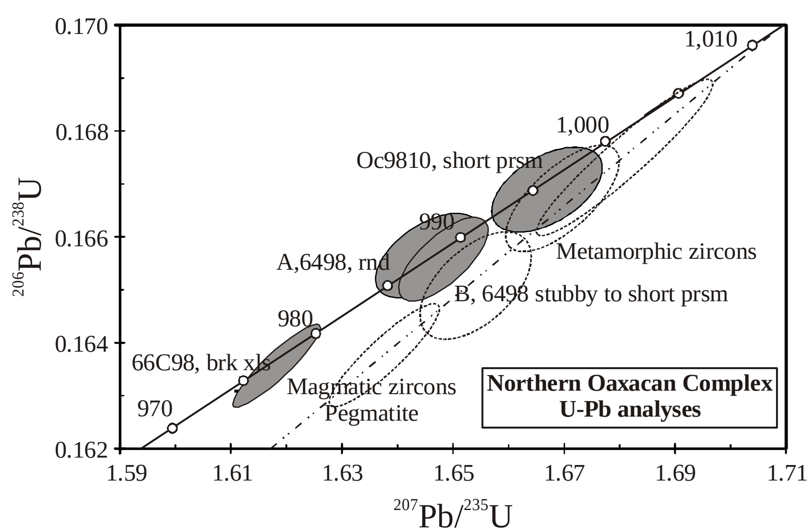

Figure 5. U–Pb concordia plot for the three analyzed samples in the northern Oaxacan Complex. Also showed for comparison are the discordant analyses (dashed ellipses) on sample 6498, with the upper intercept of ~1,012 Ma interpreted as igneous ages by Keppie *et al.* (2003). See text for further details.

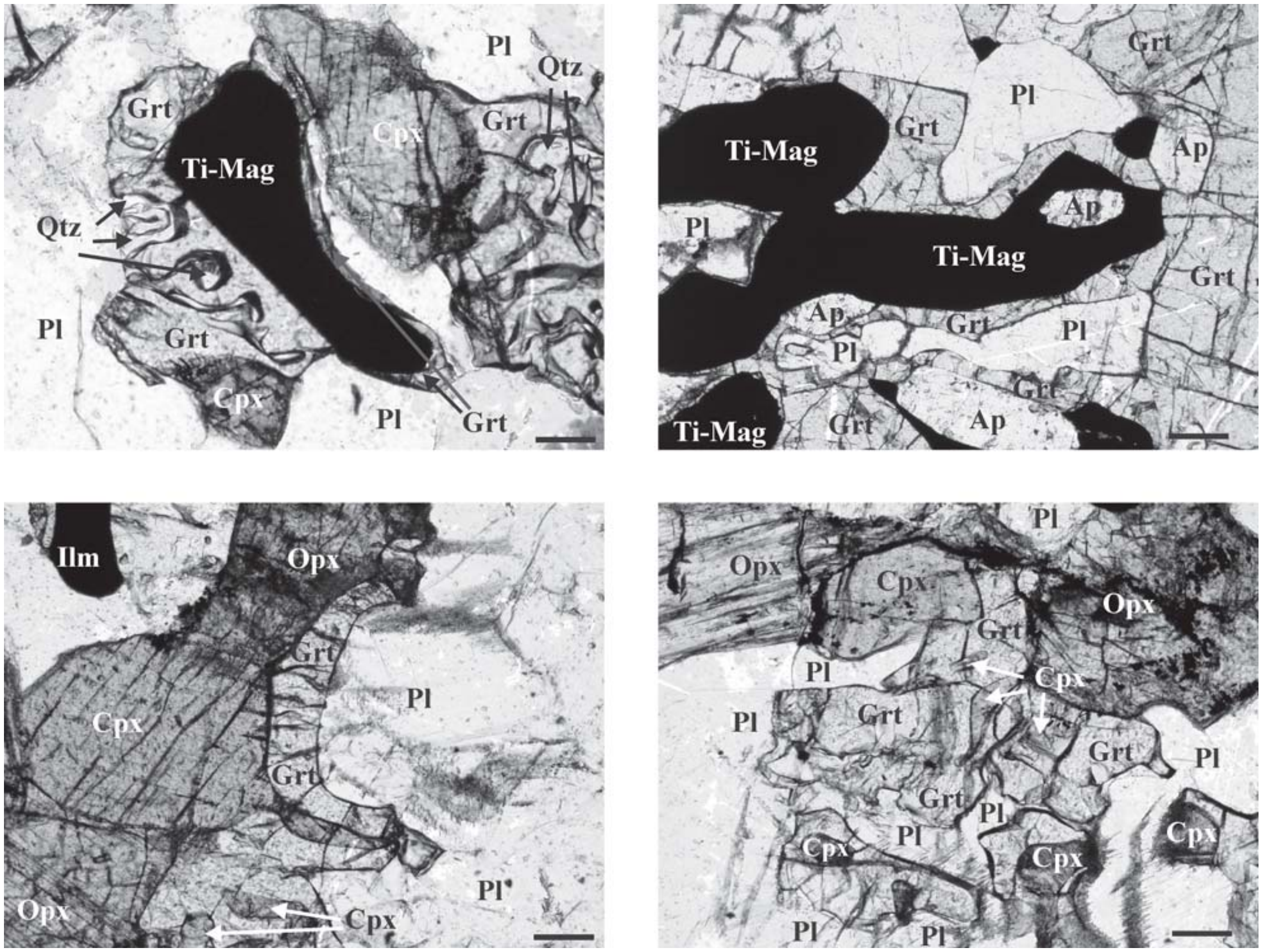

Figure 6. Photomicrographs, taken under plane polars, of the textural relationships in the sample OC9810, probed for P and T calculation. Mineral abbreviations are after Kretz (1983). The black bar in the lower right of each picture corresponds to a 100 µm scale. A+B: garnet–Ti magnetite–quartz–clinopyroxene–plagioclase association, with quartz symplectite in garnet. Garnet forms thin coronas around Ti-magnetite suggesting reaction with plagioclase. C: garnet coronas around plagioclase, suggesting reaction with clino- and orthopyroxene. D: symplectites of clinopyroxene in garnet, interpreted as reaction of the former with plagioclase.

and Kohler, 1999). On the other hand, Nd isotopic data from igneous rocks in the ~1 Ga inliers of Mexico have yielded T_{DM} model ages of ~1.4–1.6 Ga (Ruiz *et al.*, 1988; Weber and Köhler, 1999; Weber and Hecht, 2000) which suggest the presence of an older source, however it is uncertain whether this is basement or subducted sediments. Furthermore, sedimentary provenance from an older cratonic source is indicated by the common Pb isotopic signatures in the Oaxacan metasediments and by their older T_{DM} model ages (Ruiz *et al.*, 1988; Weber and Köhler, 1999).

The similarity of geochronological, P–T, and common Pb isotopic data suggest that the granulite metamorphism during to the Zapotecan orogeny (Solari *et al.*, 2003) was coeval throughout the ~1 Ga inliers of Mexico and that they probably formed a coherent block by at least ~990 Ma. This is consistent with the similar T–t cooling paths recorded in the ~1 Ga Mexican inliers (Keppie and Ortega-Gutiérrez, 1999). These data argue against subdividing the Mexican ~1 Ga inliers into Laurentian and Amazonian units across the Trans-Mexican Volcanic Belt. At the present time, Oaxaquia and the Maya terranes are separated by the Juarez terrane, a Mesozoic oceanic fragment (Figure 1A), which suggest dispersion during the breakup of Pangea and the opening of the Gulf of Mexico.

Comparisons between ~1 Ga rocks of Mexico and the other Grenvillian massifs

A comparison of the Pb isotopic composition of Mexican ~1 Ga inliers with other Grenvillian rocks indicates that the Mexican signatures generally coincide with those of the Andean Arequipa–Antofalla massif, the Colombian massifs, and the Appalachian high-µ group (Baltimore Gneiss, Blue Ridge, the State Farm Gneiss, and Pine Mountain) (Figure 9). On the other hand, they also partly

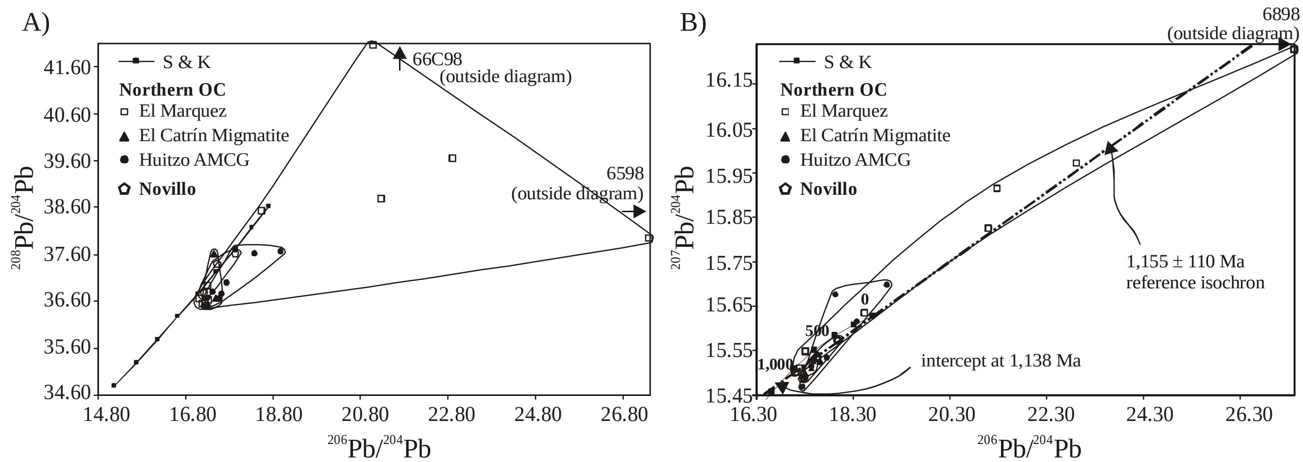

Figure 7. Whole-rock common Pb plot of the samples from the northern Oaxacan Complex analyzed in this work. A: $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$; B: $^{207}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$. S & K: average crustal growth curve of Stacey and Kramers (1975). See text for further explanations.

overlap the fields of Southern Grenville Belt and the Appalachian low-μ group (Honeybrook Uplands, Sauratown Mountains, Corbin and Tallulah Falls) (Figure 9). Relative to the Appalachian orogenic cycle, the Southern Grenville, and the Blue Ridge, Sauratown Mountains, Corbin, Tallulah Falls, and Pine Mountain massifs would be considered Laurentian basement, whereas the State Farm Gneiss and the Baltimore Gneiss have been attributed to a Gondwanan provenance (Hibbard *et al.*, 2002). Clearly, both high- and low-μ are present in each group. Thus, it would appear that the use of common Pb isotopes to distinguish Laurentian versus Gondwanan provenance fails. This is explicable in terms of the age of the source rocks, with higher radiogenic signatures reflecting an older source compared to lower radiogenic signatures reflecting a younger source. This is clear in the common Pb signatures in feldspar reported from the ~1 Ga Mexican inliers (Cameron *et al.*, in press).

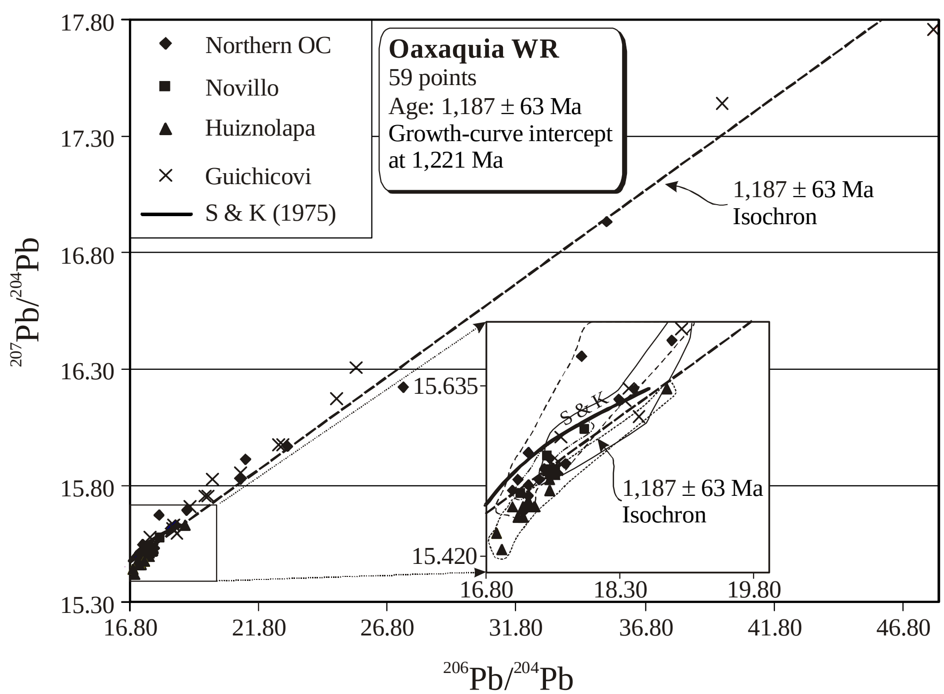

Figure 8. $^{207}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ acid-leached whole-rock common Pb plot of the samples analyzed in this work (northern Oaxacan Complex and Novillo Gneiss), as well as other samples from Oaxaquia. Guichicovi unleached whole-rock samples from Ruiz *et al.* (1999), Huiznopala are acid-leached whole-rock samples from Lawlor *et al.* (1999). S & K (1975) indicates the average crustal growth curve of Stacey and Kramers (1975). See text for further explanations.

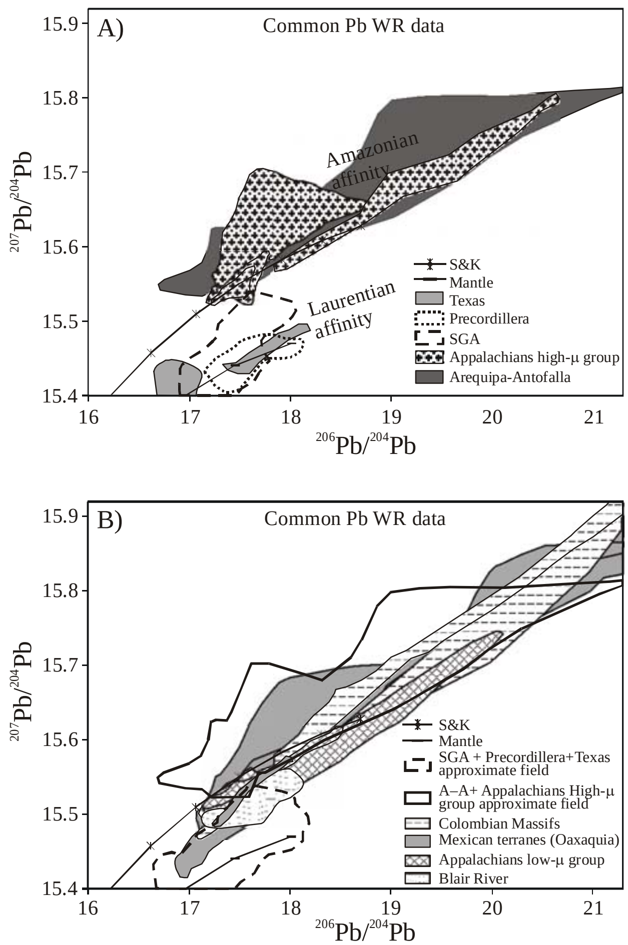

Figure 9. Common Pb data of Grenvillian massifs. A: ^{207}Pb/^{204}Pb *vs.* ^{206}Pb/^{204}Pb plot for the Laurentian-affinity Pb. SGA: field of feldspars from de Wolf and Mezger (1994); Precordillera: field of whole-rock samples from Kay *et al.* (1996); Texas fields are from Smith *et al.* (1997) and Cameron and Ward (1998). The Amazonian-affinity Pb are, respectively: Arequipa–Antofalla field from Tosdal (1996); Appalachians high-µ field from Sinha *et al.* (1996). B: ^{207}Pb/^{204}Pb *vs.* ^{206}Pb/^{204}Pb plot for the Grenville massifs: Colombian massifs (Garzón and Santa Marta) field is from Ruiz *et al.* (1999); the Appalachians low-µ field is from Sinha *et al.* (1996); the Blair River field is from Ayuso *et al.* (1996); the Mexican (Oaxaquia) field is as in Figure 8. S & K is the average crustal growth curve of Stacey and Kramers (1975). Mantle is the mantle-growth curve of Zartman and Haines (1988).

The metamorphic ages of the Zapotecan Orogeny in Mexico are similar to the U–Pb zircon age of 970 ± 23 Ma for the granulite facies metamorphism in the Arequipa massif of Peru (Wasteneys *et al.*, 1995), to the ages calculated by Aleinikoff *et al.* (1996), Miller *et al.* (1996), and Miller and Barr (2000) in Goochland and Blair River, respectively, and to those calculated along the Grenville Front by Krogh (1994). However, they are much younger than those of Texas (Roback, 1996; Rougvie *et al.*, 1996; Mosher, 1998), lending more credence to the lack of continuity between northern Oaxaquia and southern Laurentia.

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