

New insights into the history and origin of the southern Maya block, SE México: U–Pb–SHRIMP zircon geochronology from metamorphic rocks of the Chiapas massif

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Abstract The histories of the pre-Mesozoic landmasses in southern México and their connections with Laurentia, Gondwana, and among themselves are crucial for the understanding of the Late Paleozoic assembly of Pangea. The Permian igneous and metamorphic rocks from the Chiapas massif as part of the southern Maya block, México, were dated by U–Pb

zircon geochronology employing the SHRIMP (sensitive high resolution ion microprobe) facility at Stanford University. The Chiapas massif is composed of deformed granitoids and orthogneisses with inliers of metasedimentary rocks. SHRIMP data from an anatectic orthogneiss demonstrate that the Chiapas massif was part of a Permian (~ 272 Ma) active continental margin established on the Pacific margin of Gondwana after the Ouachita orogeny. Latest Permian (252–254 Ma) medium- to high-grade metamorphism and deformation affected the entire Chiapas massif, resulting in anatexis and intrusion of syntectonic granitoids. This unique orogenic event is interpreted as the result of compression due to flat subduction and accretionary tectonics. SHRIMP data of zircon cores from a metapelite from the NE Chiapas massif yielded a single Grenvillian source for sediments. The majority of the zircon cores from a para-amphibolite from the SE part of the massif yielded either 1.0–1.2 or 1.4–1.5 Ga sources, indicating provenance from South American Sunsás and Rondonian-San Ignacio provinces.

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Introduction

The scarcity of basement outcrops together with differences in age, metamorphic history, and deformation style of the crystalline rocks are factors that complicate the interpretation of the geology of southern México. Most of southern México, therefore, has been interpreted to be a collage of tectonostratigraphic terranes

(e.g., Campa and Coney 1983; Coney and Campa 1987; Ortega-Gutiérrez et al. 1990; Dickinson and Lawton 2001) whose correlation to each other and to North America is uncertain (Fig. 1a). The pre-Mesozoic locations of these land masses either as peri-Gondwanan blocks between Gondwana and Laurentia or as outboard terranes in the Pacific margin (e.g., Elías-Herrera and Ortega-Gutiérrez 2002; Centeno-García and Keppie 1999) are important for the understanding of the Late Paleozoic assembly of Pangea.

Permo-Triassic intrusive rocks occur along the entire length of México from the states of Chihuahua and Coahuila in the north, to the states of Puebla, Oaxaca, and Chiapas in the south, and extend probably into northwestern South America (Centeno-García and Keppie 1999). An apparent linear arrangement of most of these granitoid plutons has lead to the conclusion that this granitoid belt (Fig. 1) is the result of an active continental margin which was established after the final amalgamation of Pangea during the Ouachita orogeny in the Late Carboniferous to Early Permian

(Torres et al. 1999). The Ouachita-Marathon orogenic belt was established by diachronous closure of an oceanic basin between Laurentia and Gondwana during the formation of Pangea (e.g., Viele and Thomas 1989; Hatcher 2002). This process concluded by the closure of the Marathon suture in west Texas at ~ 281 Ma (Ross 1986) and was followed by the Early Permian magmatic arc that was active along the Gondwana flank of Pangea (e.g., Dickinson and Lawton 2001). Many of the suspect tectonostratigraphic terranes of southern México (Fig. 1a), therefore, must have been in contact at least since the Permian assemblage of Pangea.

Isotopic ages reported from the entire granitoid belt range from 287 to 232 Ma (Torres et al. 1999 and references therein; Iriando et al. 2003). The intrusions are mostly undeformed and scarcely affected by regional metamorphism. Most of these ages are K–Ar cooling ages of biotite, which can be affected by secondary Ar-loss. More reliable U–Pb zircon ages are mostly in the range of 287–270 Ma (Yañez et al. 1991; Solari et al.

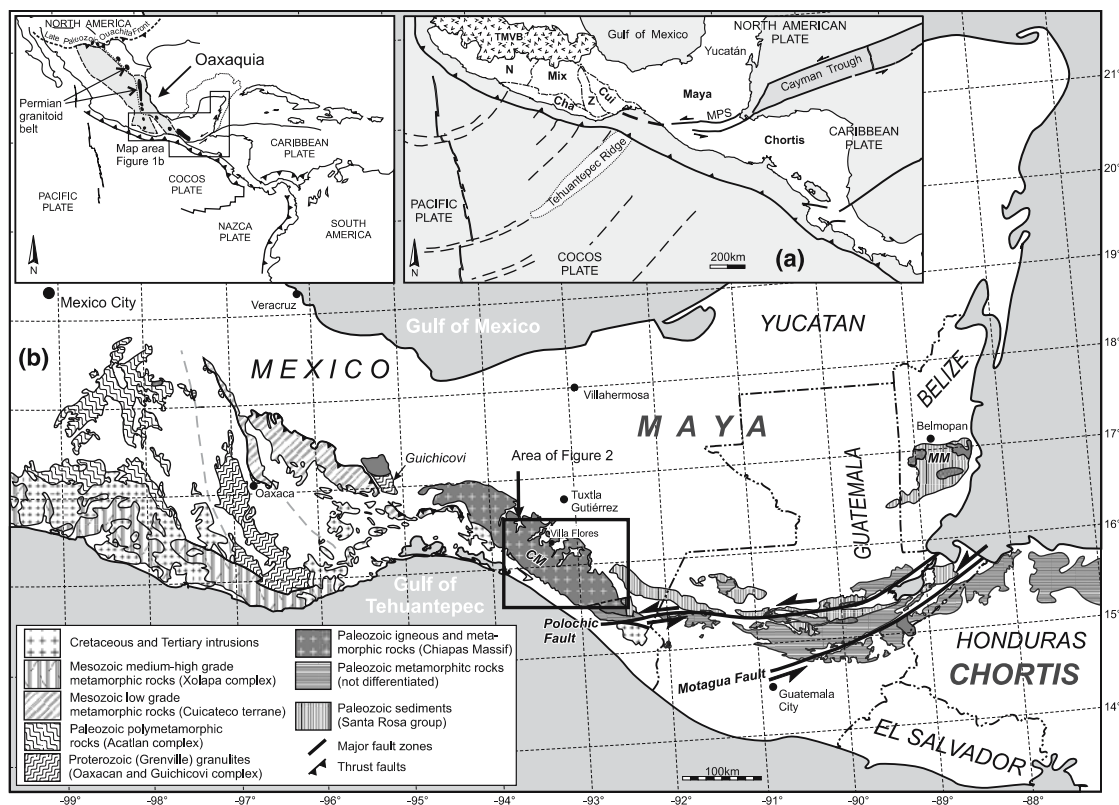


Fig. 1 **a** Terrane map of southern México and Central America (modified after Ortega-Gutiérrez et al. 1990). MPS Motagua–Polochic fault system, TMVB Trans Mexican Volcanic Belt, Terranes— *Cha* Chatino, Chortis, *Cui* Cuicateco, Maya, *Mix* Mixteco, *N* Nahuatl, *Z* Zapoteco. **b** Geologic map showing igneous and metamorphic rocks exposed in southern México and

central America and the Paleozoic sedimentary rocks of the Santa Rosa Formation (modified after Ortega-Gutiérrez et al. 1992; French and Schenk 1997). *CM* Chiapas massif, *MM* Maya Mountains. *Inset* (upper left) Possible extension of Oaxaquia modified after Ortega-Gutiérrez et al. (1995) and main localities of Permian granitoids after Torres et al. (1999)

2001; Elías-Herrera and Ortega-Gutiérrez 2002; Ducea et al. 2004).

In the Chiapas massif, conventional U–Pb age dating of zircons from gneisses have shown that the crustal precursors of the Chiapas massif are of Grenvillian age and that the most important tectonothermal event is of Late Permian age (Weber et al. 2005). However, with conventional zircon dating techniques, it was impossible to distinguish between igneous and metamorphic events in the Chiapas massif. In this paper, we present the results of our new approach to resolve this dating problem by SHRIMP-RG (sensitive high resolution ion microprobe-reverse geometry) analysis of zircons from the Chiapas massif. We also present SHRIMP data of zircon cores from metasediments giving new insights on the provenance of those sediments.

Geologic frame

The Maya block

The Maya block (Dengo 1985) is the southeasternmost of the Mexican terranes and it includes the Yucatan peninsula, parts of the coastal plain of the Gulf of México, and southeastern México from the Tehuantepec isthmus to Guatemala, where the Motagua–Polochic Fault system separates the North American Plate from the Caribbean plate and the Chortis block of Central America, respectively (Fig. 1a).

The Yucatan peninsula is mainly composed of a Cretaceous carbonate platform with an important 200-km-wide impact structure, the Chicxulub crater, which was formed at the K–T boundary (e.g., Blum et al. 1993). Shocked zircons from samples of a drill core south of Chucxulub have been dated at about 545 Ma (Krogh et al. 1993). This age indicates that rocks which were formed during the Brazilian–Pan African orogenic cycle underlie the Yucatan peninsula. The Brazilian–Pan African orogeny led to the formation of Gondwana and comprises multiple collisions of smaller continental masses starting ~ 750 Ma ago, but developing mainly between 650 and 530 Ma (e.g., Cordani et al. 2000).

Pre-Mesozoic basement rocks crop out only in the southern part of the Maya block. West of the Tehuantepec isthmus 1-Ga-old granulites form the Guichicovi complex (Fig. 1b) indicate that at least this piece of the Maya block is similar to the Oaxacan complex (Fig. 1b) and part of the Oaxaquia microcontinent (Murillo-Muñetón 1994; Murillo-Muñetón et al. 1994; Weber and Köhler 1999; Weber and Hecht 2003), which is supposed to be the backbone of México

that underlies most of the central and southern Mexican terranes (Fig. 1 inset; Ortega-Gutiérrez et al. 1995). Oaxaquia is part of the Grenville orogenic belts, which are indicated as being responsible for the formation of the Late Proterozoic supercontinent Rodinia 1.1–1.0 Ga ago (e.g., Hoffman 1991; Dalziel 1997). The Guichicovi complex is intruded by undeformed Late Permian and Jurassic granitoids and gabbros of the Mixtequita batholith (Damon et al. 1981; Murillo-Muñetón 1994; Weber 1998).

East of the Chiapas massif, the pre-Mesozoic basement is represented by undeformed and unmetamorphosed sandstones and limestones of the Pennsylvanian to Permian Santa Rosa Formation (Fig. 1b). These sediments are widespread in northern Guatemala and Belize where they are folded and metamorphosed under low-grade conditions (Bateson and Hall 1977). The Santa Rosa Formation covers the metamorphic Chuacús Group in Guatemala and Silurian plutons of the Maya Mountains in Belize (Fig. 1b; Steiner and Walker 1996).

Dickinson and Lawton (2001) assumed that the Maya block behaved as an internally rigid microplate during the Mesozoic breakup of Pangea. Paleomagnetic data (Molina-Garza et al. 1992) indicate that this microplate rotated ~ 42° anticlockwise during the seafloor spreading in the Gulf of México (Bazard and Butler 1992; Marton and Buffler 1994). An additional 18° of rotation by intracontinental extension prior to the opening of the Gulf of México must be added for a 60° rotation, necessary to achieve a close intra-Pangea fit between continental crust of northern South America and the Maya block (Dewey and Pindell 1985; Pindell et al. 2000; Dickinson and Lawton 2001).

The Chiapas massif

The Chiapas massif, which extends over an area of more than 20,000 km² parallel to the Pacific coast in southeastern México, is the most voluminous of the Permian batholithic complexes in México and it forms the southern part of the Maya block (Fig. 1b). Due to the apparent predominance of granitoid rocks in the Chiapas massif, it was named Chiapas batholith by former authors (e.g., Damon et al. 1981; Morán-Zenteno 1984). The main difference between the Chiapas massif and all other Permo-Triassic intrusions in México is the fact that most of the granitoid rocks of the Chiapas massif are foliated, strongly deformed, and metamorphosed. More recent work has shown that the Chiapas massif is formed of gneisses, frequently augengneisses and migmatites of mostly granodioritic to tonalitic chemical and mineralogical composition,

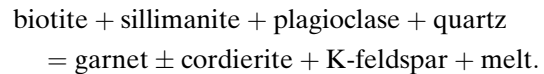
which have been interpreted as orthogneisses due to their homogeneous composition at outcrop scale. The orthogneisses are intruded by calcalkaline plutonic rocks varying from granite to gabbro in composition. These plutonic rocks are themselves affected by subsequent ductile deformation before uplift and cooling in the Early Triassic (Schaaf et al. 2002). The metaigneous rocks are intercalated by metasedimentary rocks (Weber et al. 2002), which provide information about the prebatholithic history of the southern Maya block. Our reconnaissance work yielded to the discovery of several different sequences of metasedimentary rocks in the Chiapas massif (Fig. 2) briefly described in the following.

The Sepultura unit (type locality)

Metasedimentary rocks in the Chiapas massif were first described from an area ~ 50 km west of the city of Villa Flores in the valley of the El Tablón River near the village of Los Angeles and named the Sepultura unit (Fig. 2; Weber et al. 2002).

At its type locality, the sequence is mainly composed of metapelites and metapsammites with intercalated marbles, calcsilicates, and metagreywackes. The calcsilicate rocks have two assemblages: (1) garnet, clinopyroxene, and/or wollastonite and (2) feldspar, clinozoisite, clinopyroxene, and quartz. Calcite marbles locally contain olivine or garnet

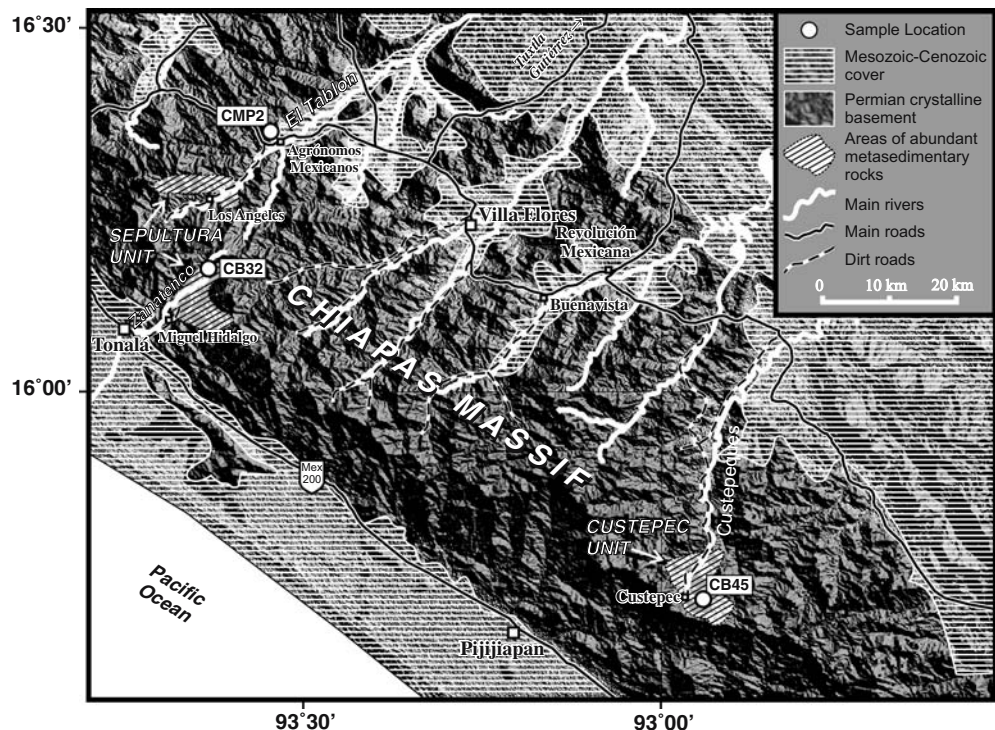
lumps. Metagreywacke, composed of quartz, feldspar, pyroxene (partly amphibolized), is associated with the calcsilicate rocks. The metapelites show medium- to high-grade metamorphic conditions with partial anatexis. The mineral assemblage is feldspar, reddish biotite, garnet, sillimanite, and sometimes cordierite. White mica occurs as secondary mineral, i.e., replacing sillimanite. The migmatites often show a neosome with garnet surrounded by K-feldspar and quartz that indicates a melt building reaction like:



Garnet–biotite thermometry and garnet–alumosilicate–plagioclase barometry yielded peak metamorphic conditions at 730–780°C and ~ 5.8 kbar, whereas retrograde assemblages revealed thermal overprinting at ~ 540°C and ~ 4.5 kbar (Hiller et al. 2004). Retrograde conditions are further indicated by chloritization of biotite and garnet.

The strike of mostly subvertical foliation in the Sepultura unit is E–W and therewith oblique to the NW–SE elongated Chiapas massif and the Pacific coast. Isoclinal folding indicates compressive tectonics following or contemporaneous with anatexis. Pervasive E–W striking foliation is younger than the latter event, as most of the intrusive rocks are affected by this deformation.

Fig. 2 SRTM (NASA Shuttle Radar Topography Mission) image of the central Chiapas massif, showing sample locations and areas of abundant metasedimentary rocks from the Sepultura and the Custepec units (see text for details). Chiapas massif boundary modified after Ortega-Gutiérrez et al. (1992)



The Sepultura unit (southern part)

Metapelitic rocks are common in about 15 km south of the type locality of the Sepultura unit and south of the main escarpment of the Chiapas massif, ENE of the city of Tonalá (Fig. 2). They are surrounded by orthogneisses that have possibly reached granulite facies conditions (Weber et al. 2005). The peak metamorphic assemblage of the metapelites is sillimanite + Ti-biotite + garnet + plagioclase $\pm$ K-feldspar. Muscovite is a major constituent and present in several generations: (1) Relic muscovite (with sillimanite inclusions) of the prograde path, indicating incomplete muscovite decomposition at the limit of high-grade metamorphism, (2) muscovite crystallized from the melt in the neosome, and (3) secondary muscovite at the expense of K-feldspar and/or sillimanite during retrogression. Andalusite is replacing sillimanite and indicates a low-pressure retrograde PT path of these rocks. The metapelitic rocks display partial anatexis, represented by garnet-bearing quartz–feldspar neosomes. These neosomes are present in cm-wide, mostly folded layers or they are consuming the rock entirely at several m-scale defining these rocks as diatexites. Secondary, retrogressive metamorphism to low grade affected the metapelites by forming secondary muscovite and chlorite at the expense of sillimanite and garnet. In adjacent hornblende–biotite gneisses, hornblende is replaced by actinolite and the latter by chlorite at a later stage.

The Custepec unit

In the southeastern part of the Chiapas massif, gneisses that are probably metasedimentary rocks are exposed in the Custepec area (Fig. 2). In this area, the gneisses are mostly hornblende-rich and are thought to be derived from mafic protoliths. They differ clearly from the metasediments of the Sepultura unit and we define them here as the Custepec unit. We interpret the rocks from the Custepec unit as being of sedimentary or volcanosedimentary origin because of their intense compositional layering and the presence of marble and/or calcsilicate layers or pebbles. The gneisses are anatectic with cm- to dm-wide neosome composed of quartz, plagioclase, and/or K-feldspar, titanite, and garnet, the latter sometimes several cm in size. The paleosome is rich in hornblende, plagioclase, and reddish biotite. The peak metamorphic assemblage is brownish Ti-hornblende, Ti-biotite, garnet (partly with sillimanite inclusions), plagioclase, titanite, and quartz. The presence of Ti-rich hornblende indicates metamorphic conditions at the amphibolite–granulite facies transition above 800°C (Spear 1981). Furthermore, the

observed assemblage of garnet + hornblende + plagioclase indicates a PT-path with higher relative pressures compared to the low-pressure PT-path of the Sepultura unit. The peak metamorphic assemblages are mostly retrogressed with green after brown hornblende, chlorite after garnet and/or hornblende, epidote and sericite in plagioclase. Marbles and calcsilicate rocks are sometimes boudinaged and they occur as dm-wide layers, m-size blocks, and at one location, marbles are exposed over several tens of meters. On the basis of our reconnaissance field work, we assume that the anatectic garnet–hornblende paragneisses (para-amphibolites) of the Custepec unit are exposed within an area of tens of km².

Analytical methods

Zircons were separated by standard procedures at the Geology Department, Centro de Investigación Científica y de Educación Superior de Ensenada (CICESE; samples CB42 and CB45) and at the Institute of Geology, Universidad Nacional Autónoma de México (UNAM; sample CMP2) by using a Wilfley® table, Frantz® isodynamic separator, heavy liquids, and handpicking techniques.

Sensitive high resolution ion microprobe procedures used in this study are similar to those reported in Nourse et al. (2005). Zircons handpicked from a total sample population and chips of zircon standard R33 (419 Ma; from monzodiorite, Braintree Complex, Vermont; R. Mundil, Berkeley Geochronology Center, personal communication; S.L. Kamo, Jack Satterly Geochronology Laboratory, Royal Ontario Museum, personal communication; J.N. Aleinikoff, personal communication) were mounted in epoxy, ground to nearly half their thickness using 1,500-grit, wet–dry sandpaper, and polished with 6- and 1- μ m-grit diamond suspension abrasive. Transmitted- and reflected-light photos were taken of all mounted grains. In addition, CL (cathodo-luminescence) images of all zircons were prepared on a JEOL 5600 SEM prior to analysis and used to reveal internal zoning related to chemical composition in order to avoid possible problematic areas within grains. The mounts were cleaned in 1 N HCl and gold-coated for maximum surface conductivity.

The U–Th–Pb analyses were made using the SHRIMP-RG housed in Green Hall at Stanford University, California and co-owned by the U.S. Geological Survey. The primary oxygen ion beam, operated at about 2–4 nA and excavated an area of about 25–30 μ m in diameter to a depth of about 1 μ m, with sensitivity ranged from 5 to 30 cps per ppm Pb.

Data for each spot were collected in sets of either five or six scans through the mass range. The reduced $^{206}\text{Pb}/^{238}\text{U}$ ratios were normalized to zircon standard R33 and either SL13 (238 ppm U) or CZ3 (550 ppm U) depending on the mount; these standard values based on conventional U–Pb dating of replicate isotope dilution analyses of milligram-sized fragments. Samples and standard were analyzed in succession for the closest control of Pb/U ratios. U and Pb concentrations are accurate to about 10–20%. SHRIMP isotopic data were reduced and plotted using the Squid and IsoplotEx programs of Ludwig (2001a, b).

Results

We have chosen zircons from three samples from the Chiapas massif for the SHRIMP-RG U–Th–Pb isotopic analysis. The isotope data and apparent ages of a total of 107 individual spots are listed in Table 1.

Migmatic orthogneiss, CMP2

This orthogneiss sample is from an outcrop on the Arriaga–Villa Flores road close to the village of Agrónomos Mexicanos (Fig. 2). It is a foliated and banded orthogneiss of granodioritic composition with alternating biotite-rich (melanosome) and quartz-feldspar-rich (leucosome) layers, interpreted as migmatite texture (Weis 2000; Schaaf et al. 2002). Zircons from this sample have already been analyzed by TIMS (thermal ionization mass spectrometry), yielding discordant isotope ratios with upper- and lower-intercept ages of $1,148 \pm 260$ and 246 ± 21 Ma, respectively (Weber et al. 2005). The low precision of both upper- and lower-intercept ages of this sample indicates that the zircons are complex and more than two events participated in the formation and history of these zircons.

As it seemed obvious from several orthogneisses analyzed by TIMS (Weber et al. 2005) that the inherited cores are Precambrian (Grenville) in age, we focused the SHRIMP study from zircons of this orthogneiss on the determination of (1) the age of the igneous protolith and (2) the age of metamorphism.

Figure 3a shows CL images of some typical zircons from sample CMP2 and the location of our SHRIMP analyses. We tried to analyze only areas of the zircon that were free of fractures and inclusions to avoid discordance as much as possible. Besides inherited cores, which are not shown at detail here, most of the zircons display magmatic zoning and metamorphic overgrowth. Accordingly, we analyzed areas either on

metamorphic tips and rims or within magmatic zoning (Fig. 3a).

Analyses from metamorphic tips and rims of four zircon grains yielded a concordia age of 254.0 ± 2.3 Ma (2σ ; Fig. 4a). The low $^{232}\text{Th}/^{238}\text{U}$ ratios of 0.1 or less (Table 1) indicate their metamorphic origin (e.g., Hoskin and Black 2000). The isotope data from ten spots on magmatic zones yielded a concordia age of 271.9 ± 2.7 Ma (95% confidence level; Fig. 4a). All except one of those spots yielded $^{232}\text{Th}/^{238}\text{U}$ ratios >0.4 (Table 1), confirming their igneous origin. Figure 4b shows the apparent $^{206}\text{Pb}/^{238}\text{U}$ ages of all analyzed spots from zircons of sample CMP2. Some of the spots were omitted from concordia age calculations (open bars in Fig. 4b) because of either high analytical errors or high discordance. Nonetheless, some spots have mixed ages (transition), probably because the oxygen beam touched zircon matter from both, igneous and metamorphic areas. Although, the age difference is only about 18 Ma, the two calculated ages are clearly discernible and do not overlap within errors ($\sim 1\%$ each) and hence, they define separate geologic events. Consequently, we interpret the ~ 272 -Ma age as the crystallization age of the granodioritic protolith, which was affected by a medium- to high-grade metamorphic event at ~ 254 Ma ago.

Paragneiss anatexite, CB32

This sample is from an outcrop on the southern part of the Sepultura unit ENE of the city of Tonalá (Fig. 2). Quartz, K-feldspar, plagioclase, garnet, Ti-biotite, and muscovite are the major constituents together with minor sillimanite, which is often retrogressed to muscovite. By its mineralogical composition and its intense compositional layering, this sample can be interpreted as a metasedimentary rock with a protolith of probably pelitic composition. The mineral assemblage indicates high-grade metamorphic conditions which led to anatexis and the formation of folded garnet-bearing leucosomes at cm–dm-scale, penetrating these rocks through the entire outcrop.

Cathodo-luminescence images of some of the zircon grains with inherited cores analyzed from sample CB32 are shown in Fig. 3b. The zircons have large, mostly prismatic cores, commonly with typical zoning observed from magmatic environments. The cores measure between about 40 and $>90\%$ of the grain size. Often, they represent partly resorbed fragments of formerly much larger grains. The cores are overgrown by zircon rims which, in most cases, display also magmatic zoning. This observation may be surprising, taking into consideration that the sample is a

Table 1 Sensitive high resolution ion microprobe–reverse geometry U–Th–Pb data for zircons from the Chiapas massif

Grain-spot (1)	²⁰⁶ Pb _c % (2)	U ppm	Th ppm	²³² Th/ ²³⁸ U	²⁰⁶ Pb* ppm (3)	²⁰⁷ Pb*/ ²⁰⁶ Pb*±1σ % (4)	²⁰⁷ Pb*/ ²³⁵ U±1σ % (4)	²⁰⁶ Pb*/ ²³⁸ U±1σ % (4)	% Disc (5)	Err corr	Apparent ages		
											²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁸ Pb/ ²³² Th
											(Ma) ± 1σ(6)	(Ma) ± 1σ(6)	(Ma) ± 1σ(6)
Orthogneiss migmatite, CMP2, Villa Flores–Arriaga highway, N 16.33917°, W 93.54383°													
CMP2-01	1.16	162	112	0.72	6.2	0.0451±9.1	0.274±9.2	0.044±1.3		0.146	277.6±3.7	–50±220	261±13
CMP2-02	2.79	184	130	0.73	7.46	0.048±13	0.303±13	0.04585±1.4		0.106	289±4	100±310	278±19
CMP2-03	3.39	69	47	0.71	2.55	0.036±29	0.204±29	0.04152±2.1		0.072	262.2±5.3	–654±790	232±31
CMP2-04	0.10	2,549	257	0.10	89.6	0.05275±1.1	0.2974±1.4	0.0409±0.8		0.614	258.4±2.1	318±24	286±7
CMP2-05	2.15	115	79	0.71	4.33	0.0482±14	0.285±14	0.04288±1.6		0.116	270.6±4.3	109±330	249±21
CMP2-06	1.25	94	42	0.46	3.54	0.0601±8.9	0.357±9	0.0431±1.6		0.183	272±4.4	609±190	305±28
CMP2-07	1.81	86	39	0.46	3.19	0.0435±14	0.253±15	0.0423±1.8		0.121	267.1±4.6	–142±360	263±30
CMP2-08	0.00	111	48	0.45	4.24	0.0595±3.7	0.364±4	0.04444±1.5		0.363	280.3±4	585±81	301±10
CMP2-09	2.17	139	68	0.51	5.18	0.0528±16	0.309±16	0.04244±1.7		0.104	267.9±4.4	319±360	263±38
CMP2-10	1.18	112	52	0.48	4.03	0.0531±14	0.303±14	0.04145±1.7		0.123	261.8±4.3	332±310	265±32
CMP2-11	1.59	87	58	0.69	3.25	0.0433±13	0.256±13	0.04289±1.7		0.130	270.7±4.5	–150±320	274±18
CMP2-12	4.10	64	27	0.44	2.28	0.037±34	0.205±34	0.0401±2.3		0.067	253.5±5.7	–548±920	162±56
CMP2-13	0.41	492	41	0.09	17.2	0.0522±3.4	0.293±3.6	0.04062±1		0.278	256.7±2.5	295±78	339±54
CMP2-14	4.62	1,822	148	0.08	68.6	0.053±28	0.306±28	0.04183±1.6		0.058	264.1±4.3	333±640	343±300
CMP2-15	0.28	739	61	0.09	26.4	0.0498±2.4	0.2847±2.6	0.04149±0.9		0.358	262.1±2.4	184±56	212±23
CMP2-16	0.13	815	850	1.08	30.7	0.0515±2.1	0.3113±2.3	0.04383±0.9		0.387	276.5±2.4	263±49	281±10
CMP2-17	0.17	1,165	110	0.10	40.7	0.05188±1.6	0.2903±1.8	0.04058±0.9		0.473	256.4±2.2	280±37	253±12
CMP2-18	17.78	670	58	0.09	26.8			0.0383±4.8			242.3±11.3	244.7±12	
CMP2-19	0.19	851	59	0.07	29.1	0.05084±2	0.2784±2.2	0.03971±0.9		0.415	251±2.2	234±45	227±20
CMP2-20	0.26	811	73	0.09	27.9	0.0507±2.3	0.279±2.5	0.03993±0.9		0.366	252.4±2.3	226±54	237±20
CMP2-21	0.10	1,439	99	0.07	52.7	0.05165±1.3	0.3032±1.6	0.04258±0.9		0.547	268.8±2.3	270±31	260±12
CMP2-22	0.76	1,372	548	0.41	50.9	0.0507±2.8	0.2999±2.9	0.04291±0.9		0.299	270.8±2.3	227±64	257±9
CMP2-23	1.04	268	165	0.64	9.92	0.0489±7	0.288±7.1	0.04268±1.2		0.165	269.4±3.1	144±160	269±11
Two mica—sillimatite—garnet paragneiss anatexite, CB32, Costa Rica, north of Tonalá, N 16.15261°, W 93.63598°													
CB32-01	–	403	19	0.05	13.5	0.0519±3.1	0.2786±3.3	0.03892±1		0.313	246.1±2.5	282±71	327±51
CB32-02	0.10	524	23	0.05	17.6	0.0503±2	0.2706±2.2	0.03901±1		0.428	246.7±2.3	209±47	222±24
CB32-03	–	562	131	0.24	19	0.0517±2.1	0.2815±2.3	0.03952±1		0.422	249.9±2.3	270±47	267±22
CB32-04	1.82	330	39	0.12	11.1	0.0408±12	0.217±13	0.03861±1.3		0.105	244.2±3.2	–299±320	
CB32-05	0.37	96	48	0.51	10.7	0.0682±3	1.211±3.3	0.1287±1.3	12	0.392	780.6±9.4	876±62	780±29
CB32-06	–	159	47	0.30	16.7	0.0732±2.2	1.23±2.4	0.1219±1.1	37	0.460	741.7±7.8	1,019±44	805±29
CB32-07	0.22	315	26	0.09	11	0.0501±3.2	0.2791±3.4	0.04038±1.1		0.318	255.2±2.7	201±75	213±26
CB32-08	0.37	382	32	0.09	13.4	0.0496±4	0.279±4.1	0.04075±1.1		0.255	257.5±2.7	178±93	203±40
CB32-09	0.22	619	153	0.26	21.6	0.0504±2.6	0.2818±2.8	0.04053±1		0.340	256.1±2.4	215±61	246±9
CB32-10	0.00	402	79	0.20	14.1	0.0513±2.2	0.2877±2.5	0.04066±1		0.417	256.9±2.6	255±51	245±7
CB32-11	–	373	36	0.10	12.9	0.0506±2.4	0.281±2.7	0.04025±1		0.392	254.4±2.6	224±56	250±13
CB32-12	1.08	917	252	0.28	32.1	0.0488±4	0.272±4.1	0.04032±1		0.232	254.8±2.4	140±93	251±16
CB32-13	0.37	275	72	0.27	9.42	0.0495±3.8	0.271±4	0.03966±1.1		0.278	250.7±2.7	172±90	240±12
CB32-14	1.70	62	67	1.11	2.22	0.0365±22	0.206±22	0.04092±2		0.093	258.5±5.2	–597±590	240±16
CB32-15	2.27	47	42	0.93	1.56	0.0414±21	0.218±21	0.03823±2.3		0.110	241.9±5.5	–264±530	216±20
CB32-16	0.51	464	70	0.16	16.1	0.048±3.8	0.267±3.9	0.04026±1		0.257	254.4±2.5	102±90	230±21
CB32-17	0.00	400	18	0.05	13.7	0.0519±2.2	0.2857±2.4	0.03994±1		0.420	252.5±2.5	280±51	272±15
CB32-18	0.17	602	101	0.17	21	0.0489±2.5	0.2732±2.6	0.04052±1		0.359	256.1±2.4	143±58	252±9.8
CB32-19	0.00	319	28	0.09	10.9	0.0521±2.5	0.2872±2.7	0.03998±1.1		0.419	252.7±2.8	290±56	273±12
CB32-20	–	373	41	0.11	13	0.0545±2.4	0.3049±2.7	0.04054±1.1		0.396	256.2±2.6	393±55	253±14
CB32-21	–	211	49	0.24	31.3	0.0747±1.5	1.778±1.8	0.1727±1.1	3	0.578	1,027±10	1,059±30	1,052±27
CB32-22	0.28	656	205	0.32	22.6	0.0491±2.5	0.2713±2.7	0.04009±1		0.350	253.4±2.3	152±59	245±8
CB32-23	0.24	176	52	0.31	15.8	0.0701±2.4	1.006±2.6	0.1042±1.1	46	0.419	638.7±6.7	931±49	614±28
CB32-24	0.23	291	33	0.12	9.76	0.05±3.3	0.2684±3.6	0.03896±1.5		0.425	246.4±3.7	193±76	226±20
CB32-25	0.08	1,179	457	0.40	41.2	0.05235±1.4	0.2931±1.6	0.04061±0.9		0.531	256.6±2.2	301±32	253±4
CB32-26	0.05	342	59	0.18	36.1	0.0692±1.9	1.172±2	0.12288±0.6	21	0.292	747.1±4.2	903±40	714±13
CB32-27	0.01	154	67	0.45	22.8	0.07063±1.3	1.679±1.6	0.1724±0.8	–8	0.528	1,025.3±7.9	947±27	1,024±18
CB32-28	–	50	21	0.44	8.43	0.075±1.9	2.029±2.3	0.1962±1.2	–7	0.544	1,155±13	1,069±38	1,159±30
CB32-29	0.23	90	28	0.32	12.6	0.0702±3.4	1.58±3.9	0.1632±1.9	–4	0.492	975±17	934±70	933±66

Table 1 continued

Grain-spot (1)	$^{206}\text{Pb}_c$ % (2)	U ppm	Th ppm	$^{232}\text{Th}/^{238}\text{U}$	$^{206}\text{Pb}^*$ ppm (3)	$^{207}\text{Pb}^*/^{206}\text{Pb}^* \pm 1\sigma$ % (4)	$^{207}\text{Pb}^*/^{235}\text{U} \pm 1\sigma$ % (4)	$^{206}\text{Pb}^*/^{238}\text{U} \pm 1\sigma$ % (4)	% Disc	Err corr	Apparent ages		
											$^{206}\text{Pb}/^{238}\text{U}$ (Ma) $\pm 1\sigma(6)$	$^{207}\text{Pb}/^{206}\text{Pb}$ (Ma) $\pm 1\sigma(6)$	$^{208}\text{Pb}/^{232}\text{Th}$ (Ma) $\pm 1\sigma(6)$
CB32-30	1.05	64	54	0.87	2.16	0.0436 $\pm$ 11	0.235 $\pm$ 11	0.03901 $\pm$ 1.7		0.149	246.7 $\pm$ 4.1	–132 $\pm$ 280	245 $\pm$ 13
CB32-31	0.23	502	67	0.14	17.5	0.0492 $\pm$ 2.4	0.2752 $\pm$ 2.5	0.04055 $\pm$ 0.6		0.238	256.3 $\pm$ 1.5	159 $\pm$ 56	228 $\pm$ 15
CB32-32	0.35	173	48	0.29	17.2	0.0687 $\pm$ 2.1	1.096 $\pm$ 2.2	0.11572 $\pm$ 0.8	26	0.332	705.9 $\pm$ 5	890 $\pm$ 44	705 $\pm$ 27
CB32-33	2.25	722	76	0.11	24.2	0.0533 $\pm$ 5.1	0.281 $\pm$ 5.1	0.03822 $\pm$ 0.6		0.108	241.8 $\pm$ 1.3	342 $\pm$ 110	259 $\pm$ 45
CB32-34	0.09	804	62	0.08	31.6	0.05523 $\pm$ 1.3	0.3485 $\pm$ 1.4	0.04576 $\pm$ 0.4		0.319	288.4 $\pm$ 1.2	422 $\pm$ 29	319 $\pm$ 15
CB32-35	0.47	39	10	0.27	5.75	0.0724 $\pm$ 3.6	1.7 $\pm$ 3.9	0.1704 $\pm$ 1.4	–2	0.368	1,014 $\pm$ 13	996 $\pm$ 73	923 $\pm$ 66
CB32-36	0.30	95	25	0.27	13	0.0721 $\pm$ 2.2	1.584 $\pm$ 2.4	0.1593 $\pm$ 1	4	0.396	952.8 $\pm$ 8.4	989 $\pm$ 45	964 $\pm$ 40
CB32-37	–	101	54	0.55	12.1	0.0719 $\pm$ 2.3	1.382 $\pm$ 2.5	0.1393 $\pm$ 0.9	17	0.367	840.9 $\pm$ 7.3	984 $\pm$ 48	916 $\pm$ 23
CB32-38	0.03	101	29	0.29	13.6	0.0768 $\pm$ 1.6	1.654 $\pm$ 1.8	0.1562 $\pm$ 0.9	19	0.492	935.7 $\pm$ 7.7	1,116 $\pm$ 31	998 $\pm$ 29
CB32-39	–	43	14	0.33	3.67	0.0723 $\pm$ 3.2	1.001 $\pm$ 3.6	0.1004 $\pm$ 1.5	61	0.429	616.8 $\pm$ 9	995 $\pm$ 66	708 $\pm$ 41
CB32-40	0.31	128	34	0.27	19.3	0.0707 $\pm$ 2.4	1.701 $\pm$ 2.5	0.1745 $\pm$ 0.8	–9	0.308	1,037 $\pm$ 7.4	948 $\pm$ 49	983 $\pm$ 39
CB32-41	0.21	57	22	0.39	8.06	0.0733 $\pm$ 2.2	1.651 $\pm$ 2.5	0.1634 $\pm$ 1.2	5	0.459	976 $\pm$ 10	1,021 $\pm$ 45	1,006 $\pm$ 35
<i>Garnet-bearing anatectic para-amphibolite, CB45, south of Custepec, N 15.71475°, W 92.95560°</i>													
CB45-01	0.40	130	136	1.08	18.9	0.0695 $\pm$ 2.7	1.621 $\pm$ 4.8	0.1692 $\pm$ 3.9	–9	0.821	1,008 $\pm$ 37	914 $\pm$ 57	992 $\pm$ 46
CB45-02	0.00	683	46	0.07	111	0.0901 $\pm$ 1.8	2.356 $\pm$ 2.4	0.1897 $\pm$ 1.6	27	0.666	1,120 $\pm$ 16	1,427 $\pm$ 34	1,143 $\pm$ 29
CB45-03	–	198	74	0.38	27.4	0.06919 $\pm$ 1.2	1.537 $\pm$ 2.1	0.1611 $\pm$ 1.7	–6	0.806	963 $\pm$ 15	904 $\pm$ 25	981 $\pm$ 21
CB45-04	1.83	108	114	1.09	3.71	0.0474 $\pm$ 13	0.257 $\pm$ 13	0.03929 $\pm$ 2.1		0.164	248.4 $\pm$ 5.2	68 $\pm$ 310	265 $\pm$ 13
CB45-05	1.14	78	54	0.72	2.74	0.0452 $\pm$ 9.7	0.253 $\pm$ 10	0.04066 $\pm$ 2.2		0.222	256.9 $\pm$ 5.6	–45 $\pm$ 240	244 $\pm$ 15
CB45-06	–	330	57	0.18	78.4	0.09292 $\pm$ 0.6	3.543 $\pm$ 1.7	0.2766 $\pm$ 1.6	–6	0.939	1,574 $\pm$ 23	1,486 $\pm$ 11	1,501 $\pm$ 53
CB45-07	1.12	103	129	1.29	3.59	0.044 $\pm$ 9.6	0.243 $\pm$ 9.8	0.04001 $\pm$ 2.1		0.211	252.9 $\pm$ 5.1	–109 $\pm$ 240	247 $\pm$ 9
CB45-08	0.09	152	89	0.60	22.7	0.082 $\pm$ 1.3	1.961 $\pm$ 2.6	0.1736 $\pm$ 2.2	21	0.860	1,032 $\pm$ 21	1,245 $\pm$ 26	1,043 $\pm$ 30
CB45-09	0.03	197	55	0.29	71.9	0.1219 $\pm$ 1.8	7.16 $\pm$ 2.5	0.4258 $\pm$ 1.7	–13	0.679	2,287 $\pm$ 32	1,984 $\pm$ 32	2,263 $\pm$ 63
CB45-10	0.00	632	32	0.05	124	0.0893 $\pm$ 4.7	2.81 $\pm$ 5	0.2281 $\pm$ 1.6	7	0.318	1,325 $\pm$ 19	1,411 $\pm$ 91	1,223 $\pm$ 39
CB45-11	0.23	131	121	0.95	26	0.0886 $\pm$ 1.8	2.804 $\pm$ 2.5	0.2294 $\pm$ 1.8	5	0.691	1,331 $\pm$ 21	1,397 $\pm$ 35	1,288 $\pm$ 27
CB45-12	0.02	1,084	106	0.10	132	0.0835 $\pm$ 1.7	1.632 $\pm$ 2.3	0.1418 $\pm$ 1.6	50	0.678	855 $\pm$ 13	1,281 $\pm$ 33	889 $\pm$ 21
CB45-13	0.01	668	82	0.13	102	0.0811 $\pm$ 2.8	1.995 $\pm$ 3.8	0.1784 $\pm$ 2.5	16	0.657	1,058 $\pm$ 24	1,224 $\pm$ 56	922 $\pm$ 36
CB45-14	0.26	325	227	0.72	11.4	0.0507 $\pm$ 2.8	0.2842 $\pm$ 3.3	0.04068 $\pm$ 1.7		0.513	257.1 $\pm$ 4.3	226 $\pm$ 65	261 $\pm$ 7.3
CB45-15	1.29	110	17	0.16	3.86	0.0409 $\pm$ 11	0.227 $\pm$ 11	0.04025 $\pm$ 2		0.187	254.4 $\pm$ 5.1	–291 $\pm$ 270	143 $\pm$ 53
CB45-16	–	1,658	43	0.03	283	0.0888 $\pm$ 1.2	2.434 $\pm$ 2	0.1988 $\pm$ 1.6	20	0.797	1,169 $\pm$ 17	1,400 $\pm$ 23	1,190 $\pm$ 33
CB45-17	–	1,217	73	0.06	224	0.092 $\pm$ 1.4	2.725 $\pm$ 2.1	0.2148 $\pm$ 1.6	17	0.744	1,254 $\pm$ 18	1,468 $\pm$ 26	1,325 $\pm$ 29
CB45-18	0.04	263	112	0.44	33.9	0.0702 $\pm$ 1.9	1.45 $\pm$ 2.7	0.1498 $\pm$ 2	4	0.726	900 $\pm$ 17	935 $\pm$ 38	957 $\pm$ 94
CB45-19	0.03	367	73	0.20	70	0.0921 $\pm$ 1.1	2.825 $\pm$ 2	0.2224 $\pm$ 1.6	14	0.830	1,295 $\pm$ 19	1,470 $\pm$ 21	1,439 $\pm$ 29
CB45-20	0.17	235	108	0.47	28.6	0.071 $\pm$ 1.7	1.38 $\pm$ 2.4	0.1409 $\pm$ 1.7	13	0.712	850 $\pm$ 13	957 $\pm$ 34	902 $\pm$ 20
CB45-21	–	73	30	0.43	2.46	0.0619 $\pm$ 9.2	0.339 $\pm$ 9.5	0.03976 $\pm$ 2.2		0.235	251.3 $\pm$ 5.5	669 $\pm$ 200	318 $\pm$ 30
CB45-22	0.07	71	87	1.26	2.51	0.0586 $\pm$ 8.4	0.331 $\pm$ 8.6	0.04092 $\pm$ 2.2		0.255	258.5 $\pm$ 5.6	552 $\pm$ 180	269 $\pm$ 11
CB45-23	0.06	699	23	0.03	84.3	0.07112 $\pm$ 0.7	1.375 $\pm$ 1.7	0.1402 $\pm$ 1.6	14	0.919	846 $\pm$ 12	961 $\pm$ 14	751 $\pm$ 51
CB45-24	–	107	124	1.21	10.2	0.0694 $\pm$ 2.6	1.065 $\pm$ 3.2	0.1113 $\pm$ 1.9	34	0.581	680 $\pm$ 12	912 $\pm$ 54	676 $\pm$ 19
CB45-25	0.00	349	151	0.45	47	0.07143 $\pm$ 10	1.546 $\pm$ 1.9	0.1569 $\pm$ 1.6	3	0.853	940 $\pm$ 14	970 $\pm$ 20	935 $\pm$ 18
CB45-26	0.00	179	100	0.58	25.4	0.0722 $\pm$ 2.2	1.641 $\pm$ 2.8	0.1648 $\pm$ 1.7	1	0.614	983 $\pm$ 16	991 $\pm$ 45	1,053 $\pm$ 23
CB45-27	0.00	531	291	0.57	119	0.09709 $\pm$ 0.5	3.486 $\pm$ 1.7	0.2604 $\pm$ 1.6	5	0.955	1,492 $\pm$ 21	1,569 $\pm$ 9	1,535 $\pm$ 30
CB45-28	0.06	286	134	0.48	54.1	0.08034 $\pm$ 0.8	2.44 $\pm$ 1.8	0.2203 $\pm$ 1.7	–6	0.902	1,283 $\pm$ 19	1,205 $\pm$ 16	1,279 $\pm$ 28
CB45-29	0.98	75	78	1.07	2.53	0.0525 $\pm$ 8	0.281 $\pm$ 8.3	0.03885 $\pm$ 2.2		0.262	245.7 $\pm$ 5.2	308 $\pm$ 180	249 $\pm$ 11
CB45-30	0.73	95	37	0.40	3.2	0.0456 $\pm$ 13	0.246 $\pm$ 13	0.0391 $\pm$ 2.2		0.163	247.2 $\pm$ 5.3	–24 $\pm$ 320	260 $\pm$ 31
CB45-31	1.38	103	34	0.34	3.57	0.042 $\pm$ 12	0.23 $\pm$ 12	0.03968 $\pm$ 2.1		0.172	250.8 $\pm$ 5.1	–225 $\pm$ 300	221 $\pm$ 30
CB45-32	2.38	39	61	1.62	1.34	0.036 $\pm$ 29	0.192 $\pm$ 30	0.039 $\pm$ 3		0.100	246.7 $\pm$ 7.2	–652 $\pm$ 810	239 $\pm$ 16
CB45-33	–	632	60	0.10	96.9	0.0864 $\pm$ 1.6	2.126 $\pm$ 2.3	0.1784 $\pm$ 1.6	27	0.695	1,058 $\pm$ 16	1,348 $\pm$ 32	998 $\pm$ 21
CB45-34	–	1,936	35	0.02	316	0.091 $\pm$ 0.4	2.385 $\pm$ 1.6	0.1901 $\pm$ 1.5	29	0.972	1,122 $\pm$ 16	1,447 $\pm$ 7	1,036 $\pm$ 24
CB45-35	0.79	78	29	0.39	2.65	0.048 $\pm$ 15	0.26 $\pm$ 15	0.03923 $\pm$ 2.3		0.153	248.1 $\pm$ 5.6	98 $\pm$ 350	234 $\pm$ 37
CB45-36	0.07	255	84	0.34	37.9	0.089 $\pm$ 1.7	2.128 $\pm$ 2.4	0.1733 $\pm$ 1.7	36	0.709	1,030 $\pm$ 16	1,405 $\pm$ 32	1,014 $\pm$ 25
CB45-37	0.30	97	97	1.04	15.9	0.0703 $\pm$ 4.4	1.851 $\pm$ 4.8	0.1909 $\pm$ 1.9	–17	0.390	1,126 $\pm$ 19	938 $\pm$ 91	1,107 $\pm$ 30
CB45-38	0.25	94	115	1.27	13.2	0.0685 $\pm$ 3.6	1.55 $\pm$ 4	0.1641 $\pm$ 1.9	–10	0.467	979 $\pm$ 17	885 $\pm$ 74	990 $\pm$ 42
CB45-39	2.19	81	45	0.57	2.86	0.0387 $\pm$ 20	0.213 $\pm$ 20	0.04 $\pm$ 2.3		0.113	252.8 $\pm$ 5.7	–439 $\pm$ 530	216 $\pm$ 28
CB45-40	1.57	88	37	0.44	3.06	0.0395 $\pm$ 15	0.216 $\pm$ 16	0.03965 $\pm$ 2.2		0.141	250.7 $\pm$ 5.4	–382 $\pm$ 400	224 $\pm$ 28

Table 1 continued

Grain-spot (1)	²⁰⁶ Pb _c % (2)	U ppm	Th ppm	²³² Th/ ²³⁸ U	²⁰⁶ Pb* ppm (3)	²⁰⁷ Pb*/ ²⁰⁶ Pb*±1σ	²⁰⁷ Pb*/ ²³⁵ U±1σ	²⁰⁶ Pb*/ ²³⁸ U±1σ	% Disc (5)	Err corr	Apparent ages		
						% (4)	% (4)	% (4)					
						²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁸ Pb/ ²³² Th					
						(Ma) ± 1σ(6)	(Ma) ± 1σ(6)	(Ma) ± 1σ(6)					
CB45-41	0.01	499	14	0.03	80.7	0.08546±0.7	2.218±1.7	0.1882±1.6	19	0.920	1,112±16	1,326 ±13	1,012±100
CB45-42	1.16	83	36	0.45	2.82	0.0393±13	0.212±13	0.03912±2.2		0.163	247.4 ±5.3	−399±340	207±23
CB45-43	–	109	97	0.92	3.74	0.0591±5.3	0.327±5.7	0.04016±2		0.358	253.8 ±5	572±110	277±10

(1) Sample identifier–spot number. (2) Contribution of common ^{206}Pb to total ^{206}Pb in %. (3) Concentration of radiogenic (*) lead ^{206}Pb . (4) Isotope ratios corrected for common Pb using measured ^{204}Pb for correction. Individual errors are given as 1σ standard deviation. (5) Deviation of $^{206}\text{Pb}/^{238}\text{U}$ age relative to $^{207}\text{Pb}/^{206}\text{Pb}$ age is given only if $^{207}\text{Pb}/^{206}\text{Pb}$ age is considered as most reliable apparent age. Positive values are for normal discordance, negative values for inverse discordance. (6) Apparent ages and all other corrections are calculated with SQUID1.11 (adapted from Ludwig 2001b). Most reliable apparent ages are in bold letters. *Note:* If the average of apparent ages is Precambrian, then $^{207}\text{Pb}/^{206}\text{Pb}$ ages are considered as most reliable apparent ages, for Phanerozoic values $^{206}\text{Pb}/^{238}\text{U}$ ages are used

high-grade metamorphic rock with a pelitic protolith. However, anatexis played an important role in the history of these rocks and, therefore, zircon growth might have been controlled by partial melting and recrystallization rather than by metamorphic zircon growth.

Most of the analyzed spots have apparent $^{206}\text{Pb}/^{238}\text{U}$ ages between 243 and 259 Ma (Table 1). The 19 most reliable isotope ratios of this group of spots define a concordia age of 253.9 $\pm$ 1.6 Ma (95% confidence limit; Fig. 5a). We interpret this age as the time of high-grade metamorphism and anatexis. This is in perfect agreement with the age of the metamorphic tips from orthogneiss sample CMP2 (254.0 $\pm$ 2.3 Ma, see above).

The apparent $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the cores range from 876 to 1,116 Ma but half of them (8 out of 16) are more than 10% discordant. (Note: We consider apparent $^{207}\text{Pb}/^{206}\text{Pb}$ ages as a more reliable approximation of individual measurements than $^{206}\text{Pb}/^{238}\text{U}$ ages in case of Precambrian zircon cores, especially when they are discordant.) Figure 5b shows the isotope data of the spots on the cores in a concordia plot. The discordant spots plot on a regression line with an upper-intercept age at 1,019 $\pm$ 43 Ma. Several spots (7) are almost concordant at this age. These results indicate that the discordant cores with apparent $^{206}\text{Pb}/^{238}\text{U}$ ages below the upper-intercept age have the same $\sim$ 1-Ga age as the concordant cores and that they have undergone lead loss during the $\sim$ 254 Ma metamorphic event. Therefore, all analyzed zircon cores of this metasediment probably came from one source of Grenville ($\sim$ 1-Ga) age and younger source regions are not involved. The maximum age of deposition is therefore roughly 950 Ma. As these zircon cores are

mostly large, euhedral, and only slightly rounded; they were not transported far, indicating a nearby source region for these sediments.

Anatectic para-amphibolite, CB45

The outcrop of this sample is located about 100 km southeast of the other two samples at the type locality of the Custepec unit (Fig. 2). The sample is a garnet-bearing anatectic para-amphibolite as described above. Quartz–feldspar leucosomes contain garnet of cm-size, indicating its in-situ formation. This sample contains rounded zircon grains (Fig. 3c) with different types of cores; for example, euhedral prismatic, rounded, fragments of formerly larger grains, either with or without complex internal structure. These cores are overgrown by mostly zoned zircon and they have sometimes unzoned rims, indicating at least two phases of zircon growth. However, the sample also contains long prismatic zircons that display magmatic zoning and unzoned rims (Fig. 3c). As we did not separate the neosome from the paleosome, the long prismatic grains, which do not show any effect of transportation, must be considered as magmatic zircons from the neosome that grew during anatexis.

Isotope data from 11 spots on long prismatic grains, magmatic overgrowth, and metamorphic rims yielded a concordia age of 251.8 $\pm$ 3.8 Ma (95% confidence; Fig. 6a, b). We interpret this age as the time of high-grade metamorphism and anatexis. It is not possible to distinguish between the igneous and the metamorphic zircon growth phases visible on CL images (e.g., spot no. 7 and spot no. 35 in Fig. 3c). Apparently, high-grade metamorphism and anatexis happened

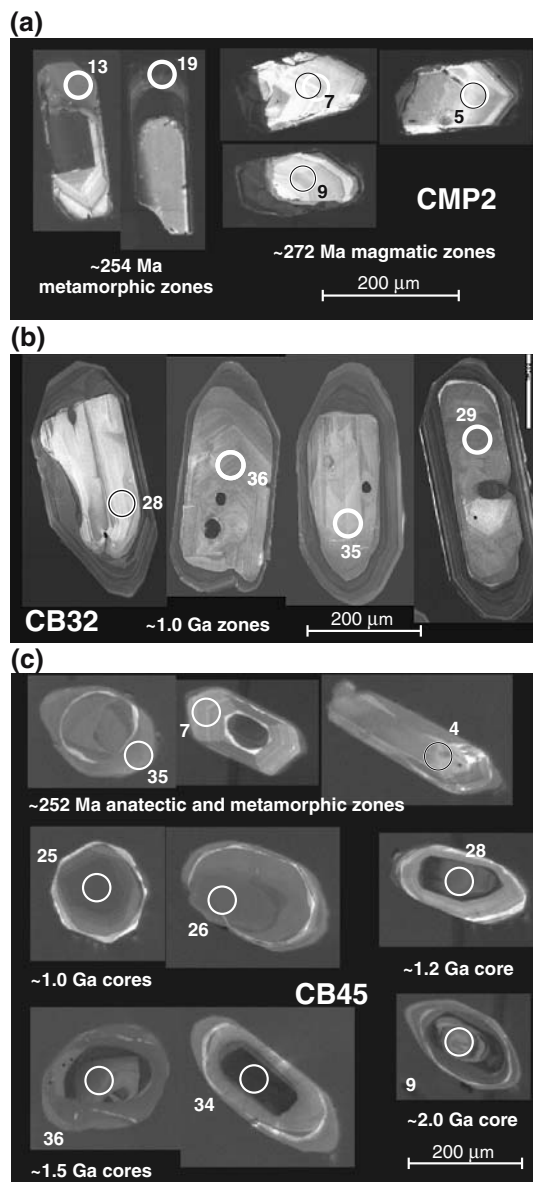


Fig. 3 Cathodo-luminescence images of selected zircons from **a** orthogneiss sample CMP2; **b** paragneiss anatexite sample CB32; and **c** anatectic para-amphibolite sample CB45. White ring number indicates spot position/grain spot number on inherited cores

simultaneously and metamorphic rims grew shortly after crystallization of the neosomes during a time span within the error of the concordia age.

The concordia plot in Fig. 6c shows the isotope data of all spots analyzed from sample CB45. Besides the anatectic and metamorphic zircons, spots were done on zircon cores in order to elucidate the provenance of the sediment. The $^{207}\text{Pb}/^{206}\text{Pb}$ apparent ages of the zircon cores range from 885 to 1,984 Ma but most of the data are discordant probably due to lead loss during anatexis and metamorphism at ~ 250 Ma. Figure 6d shows

a relative probability plot of $^{207}\text{Pb}/^{206}\text{Pb}$ apparent ages of the zircon cores. Two important zircon populations can be observed, one at 961 Ma and another at 1,447 Ma. Secondary peaks are between 1.2 and 1.3 Ga, 1.6 Ga, and one 2 b.y.-old grain. We calculated regression lines (Fig. 6c) for zircon cores that belong to the two most pronounced populations. Isotope data of most of the zircon cores (23 out of 27) plot on one of these two regression lines, which have upper-intercept ages of 968 ± 23 Ma (population 1) and $1,495 \pm 39$ Ma (population 2), respectively. The intermediate three cores correspond to a ~ 1.2-Ga-old source. The ~ 1-Ga-old population 1 has mostly zircons of igneous origin (9 out of 10) with $^{232}\text{Th}/^{238}\text{U}$ ratios greater than 0.3 (Fig. 6e). The 1.5-Ga-old population 2, in contrast, is mostly (10 out of 13) represented by zircons with $^{232}\text{Th}/^{238}\text{U}$ ratios smaller than 0.3, indicating the likelihood of a metamorphic origin (Fig. 6e; Hoskin and Black 2000).

Discussion

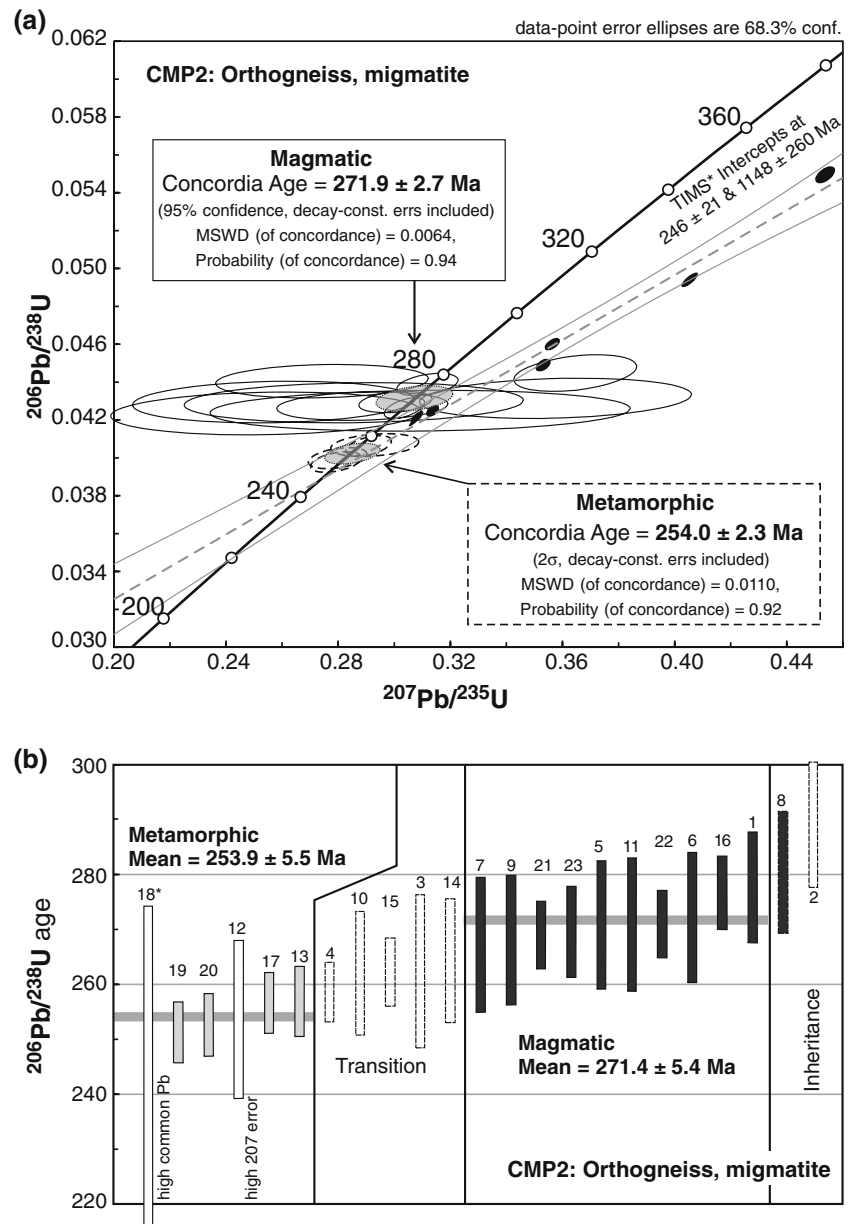
The Permian igneous and metamorphic history

The SHRIMP-RG analyses on all three samples yielded an age for a high-grade thermal event within a narrow time range from 254.0 ± 2.3 (2 σ) to 251.8 ± 3.8 Ma (95% confidence level). This is the most prominent event in the entire Chiapas massif, leading to partial anatexis in pelitic and psammitic metasediments and also locally in the orthogneiss. There is no significant time difference of this event between samples from the northwestern part of the Chiapas massif with low-pressure/high-temperature metamorphism and the sample from the Custepec area ~ 100 km further south that shows medium-pressure/high-temperature metamorphism and paragneisses of relatively mafic composition. Supported by our field observation, we suggest that this high-grade metamorphism is contemporaneous with compressive ductile deformation and folding. East–west trending foliation is the most common structural feature in the Chiapas massif in both the metamorphic basement and the plutonic rocks (Schaaf et al. 2002). Together with intense tight to isoclinal folds, the structures roughly indicate N–S directed compression. We interpret this Late Permian metamorphic event as the result of stacking in an orogenic wedge of which the Sepultura unit is the upper part and the Custepec unit is the lower part.

The metasedimentary rocks, and also the orthogneisses of the Chiapas massif, are intruded by granitoids which are themselves deformed and foliated (Schaaf

Fig. 4 a Concordia diagram for zircons from orthogneiss sample CMP2. *Black filled error ellipses* are U–Pb isotope ratios measured by TIMS from multigrain fractions. * *Discordia line with error envelope* corresponds to TIMS data from Weber et al. (2005). *Open error ellipses* indicate individual spot analysis from igneous zones (*solid line*) and metamorphic tips (*dashed line*). *Gray error ellipses* correspond to the 2 σ and 95% confidence errors of the calculated Concordia ages.

b One σ error bars of $^{206}\text{Pb}/^{238}\text{U}$ apparent ages of individual spots on zircons from sample CMP2. Weighted mean age calculations of metamorphic tips (*gray bars*) and igneous zones (*black bars*). *Open bars* were omitted from Concordia age calculations



et al. 2002; Weber et al. 2002). These deformed granitoids represent the majority of rocks that constitute the Chiapas massif. Although, no detailed zircon geochronology is done on these plutons, the available Late Permian to Early Triassic ages (Damon et al. 1981; Schaaf et al. 2002; Weber et al. 2005) together with their structural features indicate that the deformed granitoids are syntectonic plutons contemporaneous with the Late Permian high-grade orogenic event in the Chiapas massif. It is not clear, however, if the plutonic rocks evolved directly from melts formed by anatexis of the para- and orthogneisses or if they evolved from a deeper level and intruded into the older basement rocks. Most probably, both processes are relevant.

Prior to the Late Permian metamorphic event, Mid Permian plutons intruded the Chiapas massif. This is documented by the 271.9 ± 2.7 -Ma age of magmatic portions of zircons from the orthogneiss sample (CMP2), which we interpret as the age of the igneous precursor. This age is similar to Early to Mid Permian ages reported from several plutonic rocks across México (Damon et al. 1981; Yañez et al. 1991; Solari et al. 2001; Elías-Herrera and Ortega-Gutiérrez 2002; Ducea et al. 2004), resulting from arc magmatism linked to the amalgamation of Pangea along the southern extension of the Ouachita-Marathon suture adjacent to east México (Fig. 7a; e.g., Torres et al. 1999). Our new SHRIMP data support the idea that

CB32: Paragneiss anatexite

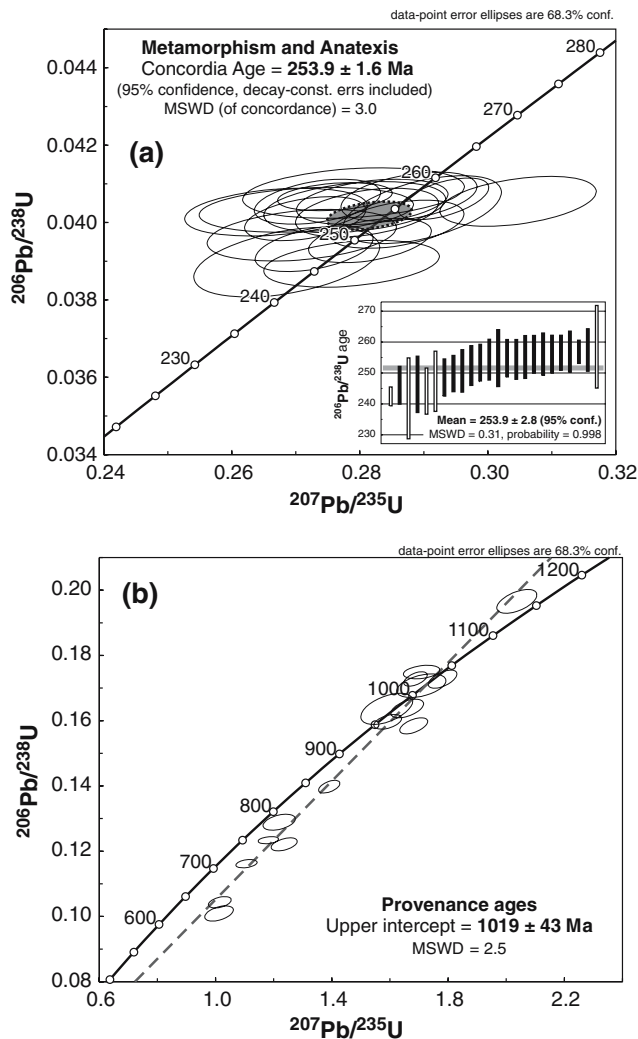


Fig. 5 Concordia diagrams for zircons from paragneiss anatexite sample CB32. **a** Open error ellipses are isotope ratios of individual grain spots on magmatic (anatectic) or metamorphic overgrowth. Gray error ellipse is the 95% confidence error of Concordia age. One σ error bars of $^{206}\text{Pb}/^{238}\text{U}$ apparent ages of individual spots and weighted mean. White bars were omitted from age calculation. **b** Open error ellipses are isotope ratios of individual grain spots on inherited cores. Concordia upper-intercept age calculated with a forced lower-intercept age of anatexis

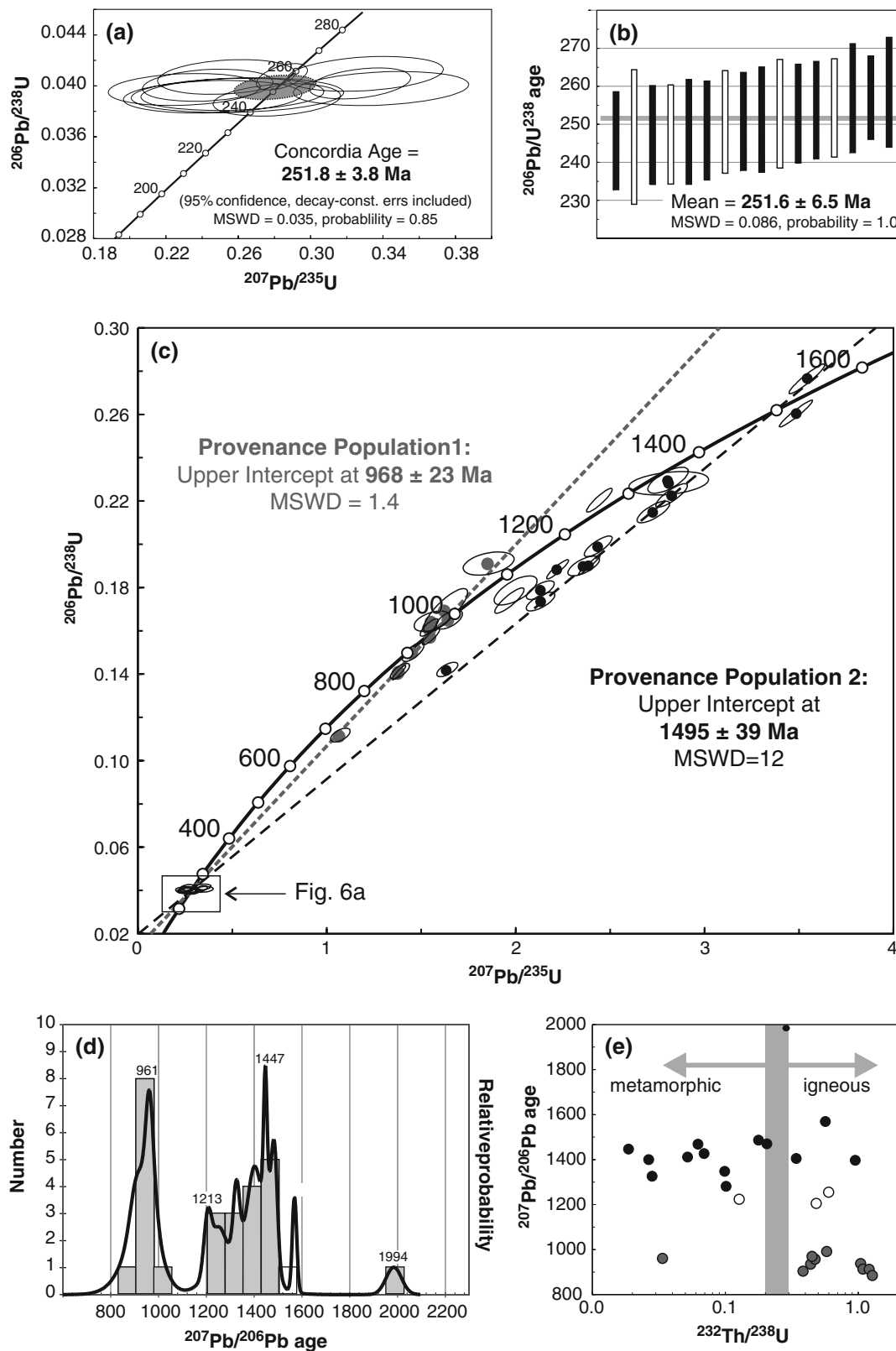
magmatism in the Chiapas massif was part of the Early to Mid Permian arc and that (at least part of) the orthogneisses of the Chiapas massif represent Mid Permian arc magmatism.

During the Early to Mid Permian, the geological history of the Chiapas massif is consistent with the models that favor a continuous magmatic arc from northeastern México to northern South America (Fig. 7a; Torres et al. 1999; Centeno-García and Keppie 1999; Dickinson and Lawton 2001), which

Fig. 6 **a** Concordia diagram showing isotope ratios of individual spots (open ellipses) of igneous zones and metamorphic overgrowth of zircons from anatectic para-amphibolite sample CB45. Gray error ellipse is the 95% confidence error of Concordia age. **b** One σ error bars of $^{206}\text{Pb}/^{238}\text{U}$ apparent ages of individual spots and weighted mean. White bars were omitted from age calculation. **c** Concordia diagram showing isotope ratios of individual spots (open ellipses) of all zircons from CB45, including inherited cores. Regression lines (forced to a lower intercept of the Concordia age) were calculated for two groups of cores yielding either Grenville (gray dots) or 1.5-Ga (black dots) upper-intercept ages. **d** Relative probability and histogram plot of apparent $^{207}\text{Pb}/^{206}\text{Pb}$ ages of inherited cores from sample CB45. **e** Apparent $^{207}\text{Pb}/^{206}\text{Pb}$ age vs. $^{232}\text{Th}/^{238}\text{U}$ ratio plot of spot analysis on inherited cores. The majority of the ~1.5-Ga cores (population 2, black symbols) have low $^{232}\text{Th}/^{238}\text{U}$ ratios indicating metamorphic origin, whereas the Grenville zircons (population 1, gray symbols) are mostly of igneous origin

started its activity after the formation of Pangea at the western continental margin of former Gondwana, including the peri-Gondwanan terranes of central México. The Late Permian high-grade metamorphic event, which is accompanied by anatexis, syntectonic intrusion, and compressive deformation, is unique in México. Of course, many Late Permian and even Triassic ages have been reported from arc-related plutonic rocks outside the Chiapas massif in México (e.g., Damon et al. 1981; Torres et al. 1999). But these ages are mostly K–Ar and Rb–Sr cooling ages of minerals that postdate intrusion and crystallization. The only exception is a U–Pb crystallization age of zircons from the Mixtequita batholith (254 ± 7 Ma; Murillo-Muñetón 1994), which is located at the western side of the Tehuantepec isthmus and possibly genetically related to the Chiapas massif. However, the Mixtequita batholith is undeformed and the only metamorphism in the area is of Grenville age (Weber and Köhler 1999).

It can be ruled out that the Late Permian orogeny in the Chiapas massif is directly related to the Ouachita-Marathon orogenic event, because the latter is significantly older and ended by Early Permian times. Therefore, a different tectonic process must be responsible for the Late Permian compressive tectonism in the Chiapas massif. In order to explain high-grade metamorphism and crustal thickening, which followed arc magmatism within a time interval of only 18 Ma, a model of tectonic switching might be applicable to the situation in Chiapas (Collins 2002). In such a model, subduction first causes lithospheric extension and melting above the subducting slab, leading to the intrusion of the Mid Permian plutons. Ongoing crustal extension along the continental margin is followed by slab rollback. Such a scenario is susceptible to the

CB45: Anatectic Paraamphibolite

arrival of buoyant oceanic crust, like an oceanic plateau or an accretionary prism, which induces local, transient change from steep to flat subduction and contractional deformation in the upper plate (Collins 2002). The Late Permian contractional deformation in the Chiapas massif is accompanied by thickening of hot continental crust and hence, it rapidly reaches high-grade metamorphic conditions leading to anatexis and syntectonic magmatism. This can also explain the fairly short time interval between 248 and 256 Ma (including errors) for the peak metamorphism in the 20,000-km² mountain range of the Chiapas massif. The regional occurrence of this kind of contractional deformation is known from other modern active continental margins like the Andes (e.g., Gutscher et al. 2000) and in our opinion, it best explains the unique position of the Late Permian orogeny in Chiapas.

Our preliminary geotectonic reconstruction of western central Pangea from the Early Permian to the Early Triassic is based on the models of Elías-Herrera and Ortega-Gutiérrez (2002) and Dickinson and Lawton (2001) and is depicted in Fig. 7. In our model, the most likely position of the southern Maya block during Early to Mid Permian is the north-western margin of former Gondwana, attached to the Paleozoic Merida Andes of present day Venezuela (Aleman and Ramos 2000, and references therein), with the Colombian Andes located to the south (Fig. 7a). The collision of Gondwana and Laurentia culminated from the east to the west along the Ouachita-Marathon suture during the Early Permian and then, subduction initiated along the western margin of former Gondwana (e.g., Torres et al. 1999). Precambrian and Paleozoic basement units of México and Central America (namely Oaxaquia, Acatlán block, and Chortis block) were accommodated within this subduction complex by dextral faulting (Fig. 7a; Elías-Herrera and Ortega-Gutiérrez 2002). The Mesa Central subduction complex (Fig. 7b), which is a Late Permian serpentinite-bearing assemblage attached to northwestern Oaxaquia, has been interpreted by Dickinson and Lawton (2001) as a forearc basin associated with the Permian arc of eastern México. We suggest that flat subduction and accretion of this forearc basin together with re-accommodation of crustal blocks caused compression and crustal thickening in the Late Permian arc and consequently the orogenic process in the Chiapas massif. Probably, the Grenville basement of northern Oaxaquia was involved in this scenario, but deformation, metamorphism, and anatexis were concentrated in the supracrustal and plutonic rocks of the Chiapas massif.

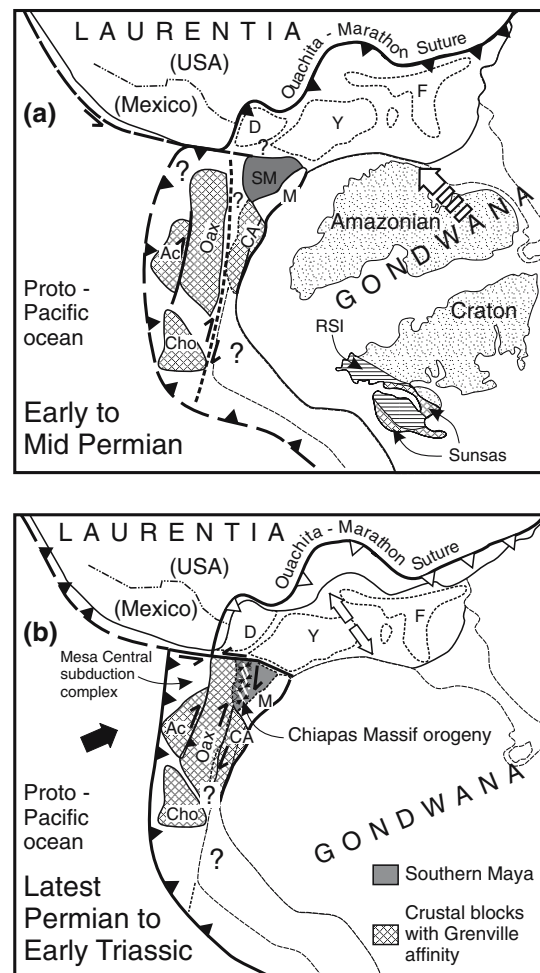


Fig. 7 Geotectonic reconstruction of west central Pangea by **a** Early to Mid Permian and **b** Late Permian to Early Triassic times modified after Elías-Herrera and Ortega-Gutiérrez (2002) and Dickinson and Lawton (2001), with emphasis on the evolution of the southern Maya block (SM). The present day outlines of the Amazonian craton are also shown in (a) together with Rondonian-San Ignacio Province (RSI) and Sunsás Province (after Cordani et al. 2000). Ac Acatlán Block, CA Colombian Andes, Cho Chortis block, D Delicias, F Florida, M Merida terrane, Oax Oaxaquia, Y Yucatán

Provenance of the basement

Concordia upper-intercept ages of zircons, analyzed by TIMS, from ortho- and paragneisses of the Chiapas massif yielded an average of ~ 1.06 Ga (Weber et al. 2005). These data, together with Sm–Nd depleted mantle model ages of similar rocks in the range from 1.0 to 1.2 Ga (Schaaf et al. 2002), already prove the existence of Precambrian, namely Grenville, crust beneath the Chiapas massif. However, these results gave only an approximation of the average age of the crust that participated in the formation of both igneous rocks and detritus. The igneous rocks, for example, can

be formed either by melting of ~ 1-Ga crust or they are a mixture of juvenile melts from the upper mantle and Grenville (or even older) crust. U–Pb TIMS analysis of multigrain zircon fractions or complex single zircon grains from the metasediments yielded upper-intercept ages from discordant isotope data that reflect only the average age of the material participating (in ortho- and in paragneisses). However, with respect to the Chiapas massif, it must be noticed that consistent results from several samples increase the probability that the average of the upper-intercept ages is the age of the provenance material.

Our metapelite sample from the southern Sepultura unit (CB32) confirms that the Chiapas massif has a basement that is Grenvillian in age, linking it with Oaxaquia or other crustal blocks of similar age (Fig. 7). All analyzed zircon cores of sample CB32 probably came from a nearby 1.0-Ga-old igneous source. This age can also be considered as a maximum age of sedimentation.

Zircon cores of the paragneiss sample from the Custepec unit (CB45) have two major provenance ages of ~ 1.0 and ~ 1.5 Ga. These results indicate that most of the sedimentary precursors of the Custepec unit come from a source region, which is older than the Grenvillian basement of southern México, namely Oaxaquia. The oldest zircons dated from Oaxaquia are from a ~ 1,350-Ma-old migmatite (Solari et al. 2003). Igneous crystallization ages of 1.2–1.25 Ga, which are also present in the Custepec sample (a minor population of three grains), are common in Oaxaquia and they have been interpreted as the age of arc magmatism (Weber and Köhler 1999; Lawlor et al. 1999; Lopez et al. 2001; Keppie et al. 2001). The ~ 1.5-Ga zircons cannot come from Oaxaquia if this microcontinent evolved as a juvenile island arc 1.35–1.2 Ga ago (Keppie et al. 2004). Even if this arc is 1.4–1.5 Ga (Weber and Köhler 1999; Weber and Hecht 2003), the source for the 1.5-Ga zircons must be different from Oaxaquia because most of the zircons of this age are of metamorphic origin. Such a metamorphic event is fairly unlikely in a juvenile arc environment.

Therefore, we propose that the southwestern Amazon craton at the Brazilian–Bolivian border is a possible source region for the sediments of the Custepec unit. In this area, the 1.0–1.2-Ga-old Sunsás orogenic belt is in contact with the 1.4–1.5-Ga Rondonia–San Ignacio province (Fig. 7a; Geraldès et al. 2001). The presence of an Early Proterozoic grain in our sample is consistent with our model, as these zircons can be transported from the Amazon craton (Fig. 7a). Keppie et al. (2003) proposed three different positions for Oaxaquia in his 1-Ga Rodinia reconstruction. One of

the possible positions for Oaxaquia at the margin of the Amazon craton is in front of the Sunsás orogen (Fig. 7a), which is consistent with our results from the Custepec unit. It is unclear, however, if this history applies to the entire Oaxaquia or if the Custepec unit is allochthonous with respect to Oaxaquia and even with respect to the northern part of the Chiapas massif. More provenance studies of the metasedimentary rocks from the Chiapas massif are required to understand the origin of these rocks and their relation to the Maya block, Oaxaquia, and South America.

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