Tectonometamorphic evolution of the Acatlan Complex eclogites (southern Mexico)

Diana Meza-Figueroa, Joaquin Ruiz, Oscar Talavera-Mendoza, Fernando Ortega-Gutierrez

Abstract: The Acatlan Complex of southern Mexico is linked to the evolution of the Appalachian–Caledonian chains and records events related to the Taconian, Acadian, and Alleghanian orogenies of northeastern North America. Mafic eclogites and garnet amphibolites from two selected localities are used to partially reconstruct the tectonometamorphic evolution of this complex. Eclogites contain garnet (almandine) + Ca–Na pyroxene + phengitic mica + zoisite–clinozoisite + quartz ± Ca–Na amphibole (barroisite, katophorite) ± albitic plagioclase ± rutile. Phase and textural relationships, thermobarometric determinations, and available radiometric ages indicate that eclogite-facies metamorphism took place during the Ordovician at temperatures around 560 ± 60°C and pressures between 11 and 15 kbar (1 kbar = 100 MPa). Eclogites underwent widespread retrogression to epidote-amphibolite then greenschist facies during exhumation, most probably during Devonian times. Epidote–amphibolite facies include the critical assemblage calcic pyroxene + calcic amphibole (magnesiohornblende and pargasite) + muscovite + garnet + plagioclase + epidote ± quartz, whereas greenschist facies is defined by the assemblage actinolite + albitic plagioclase + epidote + chlorite. Thermobarometric data suggest that retrogression occurred at temperatures between 510 ± 20°C and 300 ± 25°C and pressures ranging from 6 to 3.5 kbar. The obtained *P–T* (pressure–temperature) path suggest that the Acatlan Complex evolved in a more complex continental collisional setting, including intraoceanic arcs, than shown in previously proposed models.

Résumé : Le complexe Acatlan du sud du Mexique est lié à l'évolution des chaînes de montagnes des Appalaches–Calédonie; on y retrouve des évidences reliées aux orogènes taconien, acadien et alléghanien du nord-est de l'Amérique du Nord. Les éclogites mafiques et les amphibolites à grenat de deux localités choisies sont utilisées pour reconstruire en partie l'évolution tectonométamorphique de ce complexe. Les éclogites comprennent du grenat (almandin) + pyroxène Ca–Na, + mica phengite + zoïsite–clinozoïsite + quartz ± amphibole Ca–Na (barroïsite, katophorite) ± plagioclase albite ± rutile. Les relations de phase et de texture, les déterminations thermobarométriques et les âges radiométriques disponibles indiquent que le métamorphisme au faciès des éclogites a eu lieu au cours de l'Ordovicien à des températures d'environ 560 ± 60 ºC et à des pressions entre 11 et 15 kbar (1 kbar = 100 MPa). Les éclogites ont subi une rétrogression étendue au faciès des épidotes-amphibolites ensuite au faciès des schistes verts durant une exhumation, probablement au Dévonien. Le faciès des épidotes-amphibolites comprend l'assemblage critique pyroxène calcique + amphibole calcique (hornblende magnésienne et pargasite) + muscovite + grenat +plagioclase + épidote ± quartz alors que le faciès des schistes verts est défini par l'assemblage actinolite + plagioclase albite + épidote + chlorite. Selon les données thermobarométriques la rétrogression aurait eu lieu à des températures entre 510 ± 20 ºC et 300 ± 25 ºC et à des pressions variant entre 6 et 3,5 kbar. Selon le diagramme *P–T* obtenu le complexe Acatlan aurait évolué dans un environnement de collisions continentales plus complexe, qui comprenait des arcs intraocéaniques, que les modèles proposés antérieurement.

[Traduit par la Rédaction]

Introduction

The Appalachian–Caledonian chains of northeastern North America resulted from complex tectonics involving accretion of island arcs and continental microblocks followed by continental collision during the formation of Pangea throughout much of the Paleozoic (Dalziel et al. 1994; Keppie et al. 1996). The rifting process associated with the breakup of

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D. Meza-Figueroa.[1] Departamento de Geologia, Universidad de Sonora, Rosales y Blvd. Encinas, Hermosillo, Son, Mexico, 83000.
J. Ruiz. Department of Geosciences, The University of Arizona, Tucson, AZ 85721, U.S.A.
O. Talavera-Mendoza. Escuela Regional de Ciencias de la Tierra, Universidad Autónoma de Guerrero, AP 197, Taxco, Guerrero, Mexico.
F. Ortega-Gutierrez. Instituto de Geologia, Universidad Nacional Autonoma de Mexico, Cd. Universitaria 04510, Coyoacan D.F., Mexico.

[1]Corresponding author (e-mail: d6meza@hotmail.com).

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Pangea produced the displacement of several blocks from the cratons of North America, South America, and Africa, as well as pieces of the Appalachian–Caledonian orogenic system. Such displaced crustal blocks constituted the metamorphic basements for eastern and southern Mexico and Central America during Mesozoic time. The Paleozoic Acatlan Complex represents the basement of the Mixteco terrane, and with the Grenvillian Oaxacan Complex on the east, they define a continental block about which younger Paleozoic and Mesozoic terranes accreted to southern Mexico (Fig. 1).

A pre-late Paleozoic age for the tectonic juxtaposition of the Oaxacan and Acatlan complexes is supported by the regional stratigraphic relationships of continental and marine deposits of Early Mississippian – middle Permian age, which uncomformably cover Devonian granitoids stitching the contact between the Acatlan and Oaxacan complexes.

These data, together with the presence of eclogitized continental and ophiolitic rocks in the Acatlan Complex and a major thrust nappe that implies long range tectonic transport, led Ortega-Gutierrez (1993) to propose that an early to middle Paleozoic collisional orogeny occurred in southern Mexico. This orogeny, named Acatecan, was related to the closure of the Iapetus Ocean and terrane transfer between Gondwana and Laurentia, and the Acatlan Complex represented the suture (Ortega-Gutierrez et al. 1999).

However, the position of the Acatlan Complex (Paleozoic) west of the Oaxacan Complex (Grenvillian) is opposite to the Appalachian relationship of eastern North America. Ruiz et al. (1988) proposed two models for a correlation between southern Mexico and North to South America. The “Cordilleran” model considers the crustal complexes as terranes. In this model, the Acatlan Complex would be transferred from the Colombian Andes (Gondwana) to Laurentia during a late Paleozoic orogeny. The “Appalachian–Caledonian” model assigns a common Paleozoic orogeny to the Acatlan Complex, the Oaxacan Complex, the Appalachian–Caledonian system, and the Grenville belt of North America. A direct connection between those orogenic belts is difficult because of their spatial arrangement and because the field relationships with North America are obscured by the presence of younger rock cover (Fig. 1). Therefore, the model envisages the Appalachian suture as having cut across the Grenville belt in northern Mexico, such that the Oaxaca Complex lay on the opposite side of the Paleozoic ocean to most of North America.

The Acatlan complex has been thought to record much of the evolution of the Taconian, Acadian and Alleghanian orogenies of northeastern North America (Yañez et al. 1991; Ortega-Gutierrez et al. 1999). Ramírez-Espinosa (2001) correlates Late Ordovician – Silurian U–Pb zircon isotopic ages in granitic rocks from the Acatlan Complex (Ortega-Gutierrez et al. 1999) with ages widely reported in the northern Appalachian system as a consequence of a Silurian orogenic pulse (called the Salinian or early Acadian by Dunning et al. 1990 and Hibbard 1994, respectively). This orogenic pulse was a result of the final collision of the Avalonian microplates against Laurentia and is correlative to the Caledonian orogeny in Europe. Ramírez-Espinosa (2001) based on geochronological data and lithological similarities, proposes a possible correlation of the Acatlan Complex with rocks currently sited in northeastern North America.

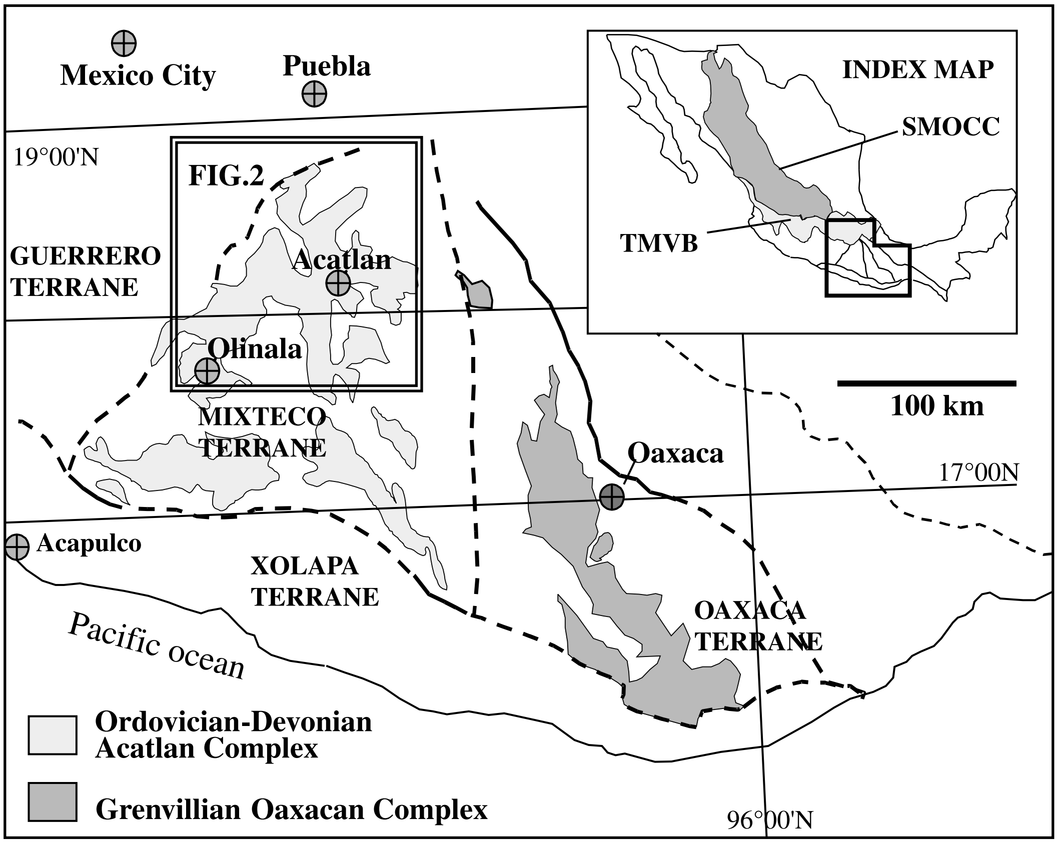
The Acatlan Complex of southern Mexico is a lower Paleozoic polymetamorphic complex; it represents the basement of the Mixteco terrane, which is surrounded by the Oaxaca, Xolapa, and Guerrero terranes and covered by the overlapping Trans-Mexican volcanic belt (Campa and Coney 1983; Fig. 1). The Mixteco terrane is boundered to the east by the Oaxaca terrane, which consists of weakly deformed to nondeformed Paleozoic and Mesozoic rocks that overlie a granulite basement of Grenville age. The Oaxaca and Mixteco terranes are overlapped by Pennsylvanian–Permian sedimentary rocks. South of the Mixteco terrane lies the Xolapa terrane, which is made up of poorly dated, possibly Mesozoic and Tertiary metamorphic and plutonic rocks. The Guerrero terrane lies to the west; this is an assemblage of Mesozoic island-arc rocks that was thrust eastward over the Mixteco terrane (Campa and Coney 1983). The northern limit of the Mixteco terrane is obscured by the Trans-Mexico volcanic belt, and it is therefore not clear whether its northern boundary is the Guerrero terrane or the Sierra Madre terrane, which is made up mainly of folded and thrust-stacked upper Mesozoic sedimentary rocks resting on sequences of deformed strata as old as early Paleozoic in age (Fig. 1).

Eventhough the Acatlan Complex is an important piece of the world’s Paleozoic orogens, containing an almost complete record of pre-Mississippian Paleozoic ocean closure and its consequent continental interactions, its precise location during the Paleozoic evolution remain uncertain. Eclogitized rocks are considered important pieces within the Acatlan Complex, and they could provide a detailed petrogenetic record of convergence during the Acatecan collisional orogeny (Late Ordovician – Early Silurian). Although the presence of this eclogitized sequence in the region has been recognized since it was presented by Ortega-Gutierrez (1974), its geology, structure, composition, and metamorphism remain essentially unknown. This paper summarizes and discusses the mineralogy, textural relationships and pressure–temperature (P–T) conditions of mafic eclogites in an attempt to better understand the petrotectonic history of this poorly known complex. These data together with available radiometric ages are used to deduce the prograde P–T path followed during their formation and exhumation.

Acatlan complex

According to Ortega-Gutierrez et al. (1999) the Acatlan Complex can be subdivided into two principal tectonic units (Petlalcingo and Piaxtla groups) separated by a major thrust overlapped by a weakly metamorphosed and strongly deformed Devonian volcanosedimentary sequence named the Tecomate Formation. The lower plate is known as the Petlalcingo Group, and it consists of a thick package of metasedimentary rocks which includes migmatite, biotite schist, and phyllite and quartzite. From bottom to top, the Petlalcingo Group is subdivided into three units. The Magdalena migmatite consists of alternating, mainly granitic neosomes and biotite-rich paleosomes from pelitic and psammitic sediments. Ages of protolith and migmatite remain uncertain. Reported ages of the migmatite are Sm/Nd garnet whole-rock and Rb/Sr muscovite whole-rock ages of 204 ± 6 and 163 ± 2 Ma, respectively, (Yañez et al. 1991); the Sm–Nd age (204 ± 6 Ma) agrees with a Rb–Sr muscovite whole-rock age for the

Fig. 1. Tectonostratigraphic setting of Acatlan Complex of southern Mexico. Guerrero and Xolapa terranes are Mesozoic in age. Oligocene and Pliocene–Quaternary volcanic rocks of Sierra Madre Occidental (SMOCC) and Trans-Mexican volcanic belt (TMVB) are included in the index map.

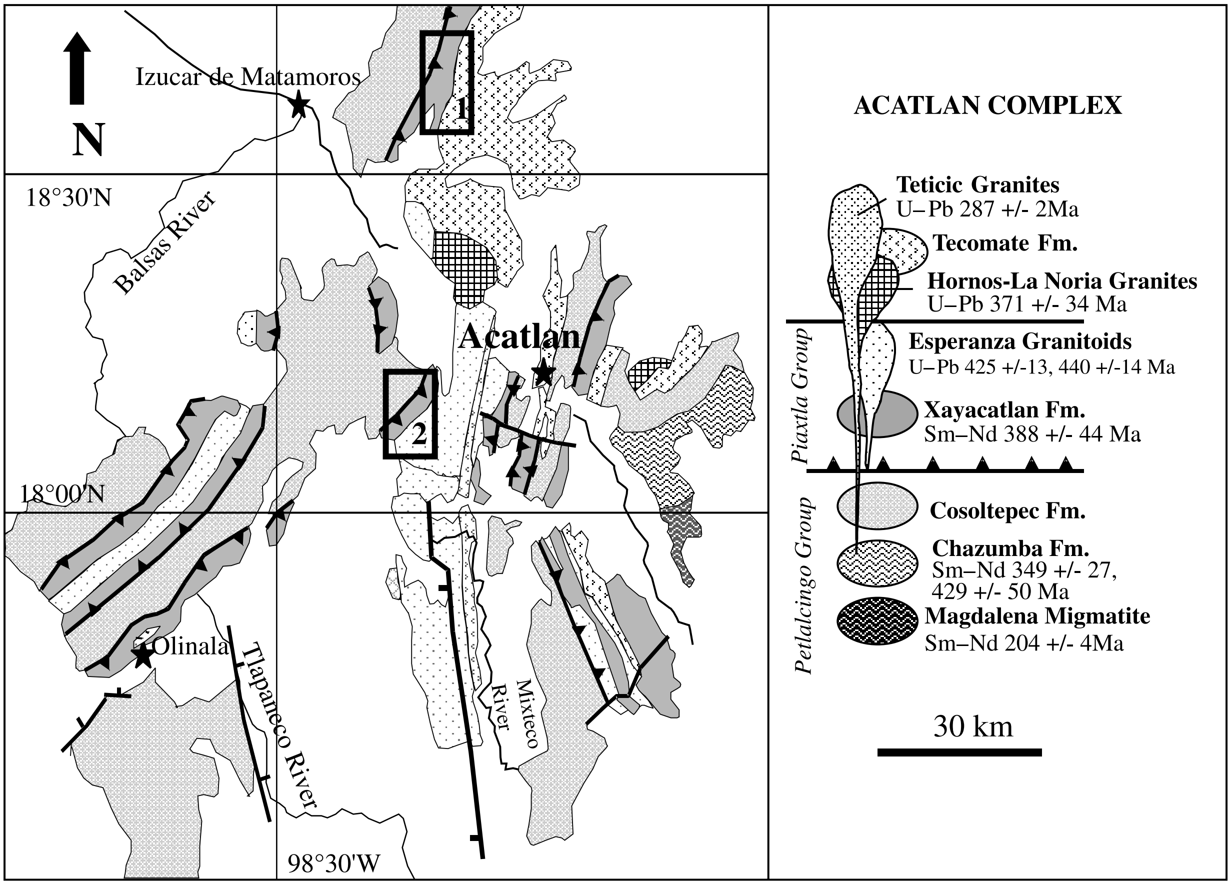


posttectonic San Miguel intrusion of 207 ± 9 Ma obtained by Ruiz-Castellanos (1979). Deformational and compositional similarities in the granitoids of the San Miguel and Magdalena units suggest that they may be the result of a single tectonothermal event (Yañez et al. 1991). However, if these data represent reset ages remain a subject of debate. The younger biotite whole-rock age (163 ± 2 Ma) for the Magdalena may be the result of the lower blocking temperature, about 300°C, of biotite (Dodson 1973). The overlying Chazumba Formation is made up mostly of biotite-rich psammitic and pelitic schist, minor quartzite, and scarce mafic and ultramafic rocks affected by an amphibolite-facies metamorphism (Ortega-Gutierrez 1974). The metamorphism of the Chazumba Formation has been dated as 429 Ma (Sm/Nd single-garnet whole-rock pair isochrone, Yañez et al. 1991). The widespread Cosoltepec Formation represents almost 70% of the Acatlan Complex outcrops. The Cosoltepec Formation is either tectonically overthrust by the Piaxtla Group or unconformably overlain by the posttectonic Tecomate Formation. The Cosoltepec Formation is made up of black slate, phyllite, and fine-grained quartzite. In the eastern part of the Acatlan Complex, the Cosoltepec Formation displays a prograde metamorphism developing chlorite–biotite–garnet metamorphic zones, whereas in the western part, it is only characterized by the development of chlorite. Massive and pillowed basaltic rocks are included as tectonic slivers within the Cosoltepec Formation sharing the same deformation and metamorphism. Reported isotopic ages performed in pillowed basaltic rocks yielded 288 ± 13 Ma (Ar^{39}/Ar^{40} whole-rock, Campa and Lopez 2000) and 452 ± 22 Ma (Rb/Sr whole-rock analysis; Ortega-Gutierrez et al. 1999). Campa and Lopez (2000) also reported the presence of argon excess, which could be due to the greenschist metamorphism, therefore the crystallization age of these rocks is not well defined. The Cosoltepec Formation is tectonically overlain by the Piaxtla Group of Late Ordovician – Early Silurian, as well as by the unconformable Tecomate Formation of probable Devonian age (Fig. 2). Based on this, Ramírez-Espinosa (2001) reported a best estimated of the deposition of the Cosoltepec Formation as Cambrian(?)–Ordovician.

The upper plate is made up of the Piaxtla Group (Ramírez-Espinosa 2001), which is formed by eclogitized mafic and ultramafic rocks and garnet amphibolites interlayered with pelitic and siliceous metasedimentary rocks (Xayacatlan Formation), structurally overlain by tabular-shaped bodies of variable dimension, composition, and fabrics varying from megacryst K-feldspar augen gneiss to fine-grained augen gneiss and affected by high-pressure metamorphism (Esperanza Granitoids; Fig. 2).

Eclogites of the Xayacatlan Formation yielded a Devonian metamorphic age of 388 ± 44 Ma (Sm/Nd garnet whole-rock isochron, Yañez et al. 1991). Geochemistry of the eclogites shows that they represent oceanic material intermixed with continental rocks during subduction. Major and trace element analyses performed on mafic eclogites from the areas of Mimilulco and Piaxtla consistently point to two geochemically coherent groups: protoliths consistent with mid-ocean ridge basalts (MORB) and ocean-island basalts OIB (Piaxtla) and

Fig. 2. Geologic map of the Acatlan Complex, after Ortega-Gutierrez et al. (1999). Studied areas are shown within squares 1: Mimilulco and 2: Piaxtla.



island-arc basalts (Mimilulco) (Meza-Figueroa 1998), both localities are considered as part of the same metamorphic belt.

The Esperanza Granitoids have been associated with the Xayacatlan Formation due to similar high-pressure metamorphism but erroneously considered of Devonian age (U/Pb zircon ages of 371 ± 34 Ma, Yañez et al. 1991). Late Ordovician – Early Silurian U/Pb zircon ages (440 ± 14 Ma, Ortega-Gutierrez et al. 1999) have been obtained from two different localities of the Esperanza Granitoids. These data suggest that there were different granitoids within the Acatlan Complex sharing similar mylonitic deformation but different metamorphism and ages.

The Piaxtla Group tectonically overrides the Petlalcingo Group following a westward emplacement (Ortega-Gutierrez 1993). The major thrust that separates the plates was subsequently folded twice along northeast- to northwest-trending recumbent and upright structures. The two tectonically superposed units were exhumed and covered by a sequence that included basic volcanics, sandstones, and phyllites currently affected by greenschist metamorphism (Tecomate Formation) and intruded by plutons (La Noria: 371 ± 34 Ma, and Totoltepec: 287 ± 2 Ma; Yañez et al. 1991; Fig. 2).

Sampling and analytical techniques

Cross section encompassing eclogitic belts in the Piaxtla and Mimilulco regions, State of Puebla, allowed a detailed determination of their internal stratigraphy and structural characteristics. Key megascopic structures and mineralogical compositions were studied in further detail under the polarized light. In total, more than 45 samples were studied under the polarizing microscope, six of which were selected for detailed electron microprobe analysis. Location of probed samples are shown in Fig. 3.

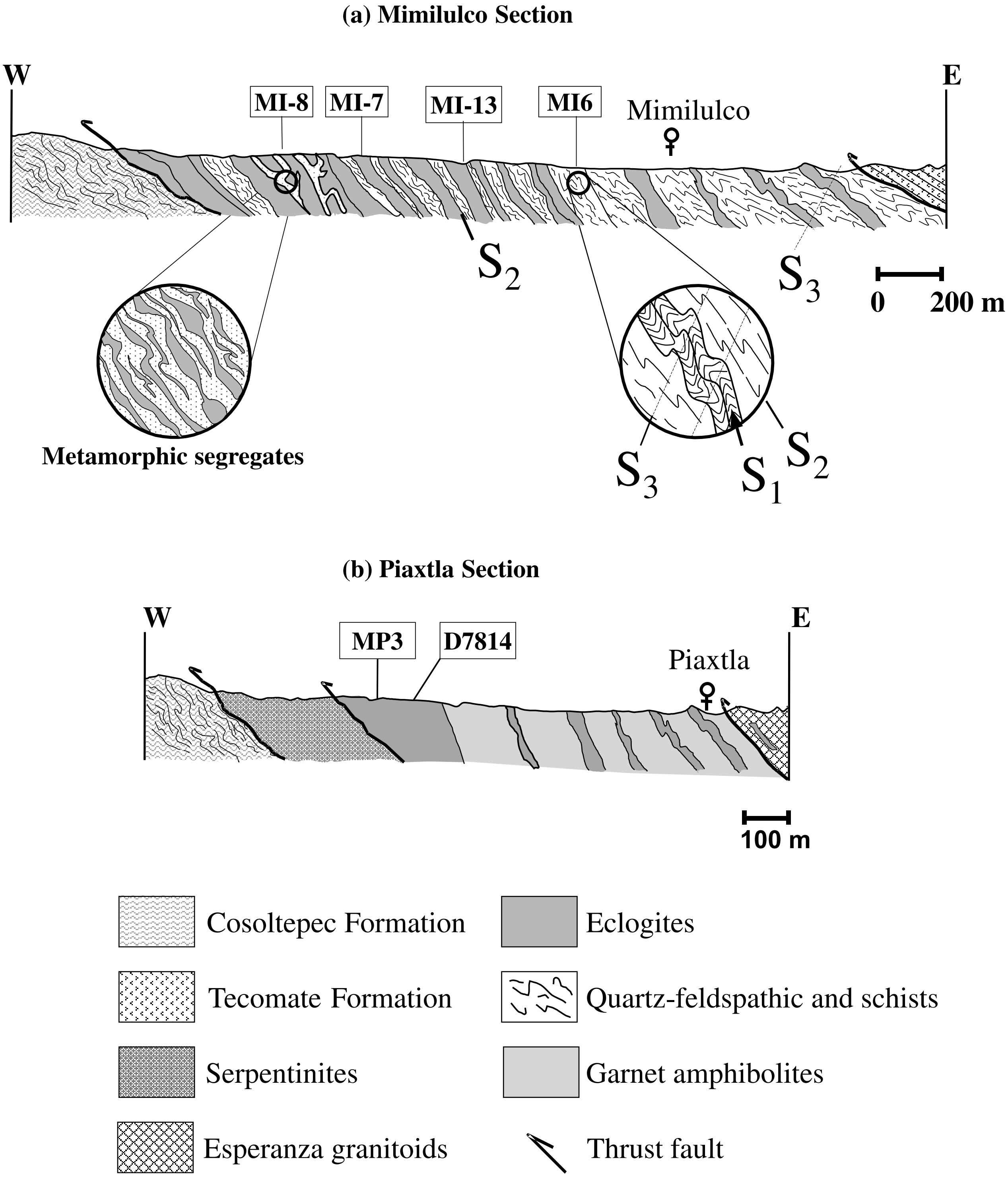
Mineral analyses were performed using a CAMECA SX-50 electron microprobe at the Lunar and Planetary Laboratory, Department of Planetary Sciences, University of Arizona, Tucson, Arizona, using the following conditions: beam current of 20 nA and an accelerating voltage of 15 kV. Counting time for all elements was 15 s. Natural minerals (albite: Na; K-feldspar: K; diopside: Si, Mg, and Ca; anorthite: Al; rhodonite: Mn; rutile: Ti; fayalite: Fe; chromite: Cr) were used as standards. Under these conditions, contents below 0.1% are considered below detection limits. Representative chemical analyses of minerals used for *P–T* estimates are shown in Table 1.

Petrography and mineral chemistry

In the Acatlan Complex, mafic eclogites and their retrogression products appear within numerous parallel, NE–SW-trending belts along the whole complex (Fig. 2). Invariably, these high-pressure belts rest tectonically above the low-grade metasedimentary rocks of the Cosoltepec Formation.

Although relics of high-pressure assemblages can be

Fig. 3. Schematic cross section of eclogitic belts in the (*a*) Mimilulco and (*b*) Piaxtla areas showing location of probed samples. Vertical profile not at scale.



recognized anywhere within the eclogitic belts, the best exposures recognized at present are undoubtly those of the Piaxtla and Mimilulco areas (Fig. 1). In both areas, eclogitized rocks and their retrogression products crop out continuously for more than 3 km, and their internal stratigraphy and structure can be clearly established.

Locally, garnet amphibolite interbedded with thin quartz-feldspathic and schist layers dominate at the lower structural levels (Mimilulco section, Fig. 3). In the middle structural levels, quartz-feldsphatic and schist layers appear as abundant as garnet amphibolite. In the Mimilulco area, eclogites of mafic composition appear as relics within the garnet amphibolite. Metamorphic segregates are conspicous in the middle part of the Mimilulco section, whereas serpentinitic bodies occur in the Piaxtla area. The upper structural levels are dominated by schist and garnet amphibolite, and eclogitic

Table 1. Representative microprobe analyses of minerals from metaeclogites from the Acatlan Complex, southern Mexico.

	Garnet$_1$					Pyroxene				Amphibole				Phengite		
Sample:	MP3-23	MI6-86	MI8b-51	MI13-10	MP3-30	MP3-42	MP3-20	MI6a	M7	MP3-14	MI-13	MP3-15	D7814	MI8b-19	MP3-56	MP3
Analysis:	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2 (wt.%)	38.02	38.03	37.73	38.10	38.17	55.85	55.76	51.88	55.10	42.56	52.88	47.19	47.67	51.40	54.41	53.81
TiO_2	0.02	0.01	0.17	0.12	0.01	—	0.02	—	0.09	0.01	0.04	0.02	0.58	0.27	0.07	0.08
Al_2O_3	21.44	21.50	21.09	21.26	21.31	9.99	9.35	6.67	4.94	16.51	1.22	10.37	13.35	28.47	31.26	31.08
FeO	25.74	27.12	25.29	25.63	25.48	4.61	4.59	11.10	10.51	15.42	14.78	12.02	13.79	3.86	1.80	1.94
MnO	1.88	1.60	4.26	2.68	1.46	0.04	0.01	0.03	0.07	0.08	0.40	0.13	0.02	0.09	—	—
MgO	3.12	2.03	1.45	1.73	3.32	9.21	9.91	15.77	16.56	9.24	13.79	13.97	10.96	4.04	3.21	3.33
CaO	10.37	9.52	10.54	11.04	10.44	15.60	16.56	13.15	10.04	9.75	12.21	12.03	7.78	0.07	0.05	0.01
Na_2O	—	—	—	—	—	5.83	5.38	1.40	1.59	3.53	0.29	2.09	4.50	0.13	0.79	0.79
K_2O	—	—	—	—	—	—	—	—	—	—	0.05	—	0.58	8.58	4.44	4.52
Total	100.6	99.9	100.5	100.5	100.2	101.1	101.5	100.1	99.3	97.1	95.7	97.8	99.2	96.9	96.04	95.6
	24 oxygens					6 oxygens				23 oxygens				11 oxygens		
Si	5.969	6.026	5.987	6.012	5.997	1.979	1.972	1.899	2.004	6.318	7.865	6.841	6.819	6.709	6.873	6.843
Ti	0.003	0.001	0.021	0.014	0.002	0.001	0.001	—	0.002	0.001	0.005	0.002	0.062	0.027	0.032	0.008
Al	3.968	4.016	3.944	3.954	3.946	0.414	0.390	0.288	0.212	2.889	0.214	1.771	2.250	4.380	4.654	4.659
$Fe^{2}+$	3.319	3.594	3.311	3.352	3.296	0.298	0.101	0.327	0.320	1.914	1.838	1.457	1.649	0.421	0.190	0.207
$Fe^{3}+$	0.061	—	0.046	0.030	0.052	0.028	0.035	0.013	0.008	—	—	—	—	—	—	—
Mn	0.250	0.215	0.573	0.358	0.194	0.001	—	0.001	0.002	0.010	0.050	0.016	0.002	0.010	—	—
Mg	0.730	0.480	0.342	0.406	0.778	0.487	0.522	0.861	0.896	0.003	3.057	3.020	2.337	0.786	0.604	0.631
Ca	1.745	1.616	1.792	1.867	1.758	0.592	0.628	0.516	0.391	1.551	1.945	1.868	1.193	0.010	0.002	0.001
Na	—	—	—	—	—	0.401	0.369	0.100	0.112	1.015	0.082	0.589	1.248	0.033	0.194	0.195
K	—	—	—	—	—	—	—	—	—	—	0.009	—	1.105	1.432	0.717	0.735
Total	16.044	15.959	16.019	15.996	16.026	4.013	4.017	4.006	3.999	15.058	15.065	15.564	16.665	13.807	13.266	13.279

Note: act, actinolite; parg, pargasite; ede, edenite; MP3, Piaxtla basic metaeclogite; MI6, Mimilulco retrogressed eclogite-facies rock; MI8b, Mimilulco sample of a melanosome, MI13, MI8b, and M7, retrogressed basic eclogite from Mimilulco; D7814, retrogressed metabasic eclogite from Piaxtla.

rocks only appear as thin layers and lenses. In most places in the Acatlan Complex, the Esperanza Granitoids are separated from the eclogites and amphibolites (Xayacatlan Formation) by a thrust fold; however, in the Piaxtla area, eclogites are intruded in the uppermost structural levels by the Esperanza Granitoids and locally, lenses of well-preserved eclogites can be seen embedded into granitoids (Ortega-Gutierrez et al. 1999). In the Mimilulco area, eclogitic rocks are parallelly covered by mafic schists and metasediments of the Tecomate Formation, although the nature of the contact is clearly tectonic.

Two common eclogitic rocks are formed: (1) mafic eclogite consists of almandine (garnet$_1$) + omphacite + barroisite + phengite + rutile in textural equilibrium, and (2) eclogitic metasediment consists of almandine (garnet$_1$) ± omphacite ± diopside ± rutile + phengite; these phases represent high-pressure assemblages and they appear as relic, coarse-grained, granoblastic microstructures in extensively retrogressed rocks (Figs. 4*a*–4*c*). Garnet amphibolite is medium to coarse grained and is composed essentially of almandine (garnet$_2$) + calcic–sodic amphibole (magnesiokatophorite and barroisite) and plagioclase. Eclogitic metasediments are dominated by retrogressive phases, and the most frequent assemblage is quartz + albite + phengite + chlorite + epidote group and relic garnet, plus an amphibole that remained stable under retrograde conditions: barroisite.

Main fabric of the studied eclogites is dominated by a well-developed, continuous to spaced mylonitic foliation related to D$_2$, which accompanied retrogression. In most samples, previous, S$_1$ foliation planes containing premylonitic, high-pressure metamorphic phases are parallel to S$_2$ planes, and consequently prograde and retrograde phases appear forming the same planar structure. Except for omphacitic pyroxene and rutile, prograde and retrograde assemblages broadly contain the same mineral phases, and thus omphacite and rutile-bearing assemblages in textural equilibrium with other phases are the only microscopic criteria to identify high-pressure domains (Figs. 4*a*–4*c*).

Garnet$_1$ forms inclusion-rich porphyroblasts and invariably shows partial resorption by chlorite. Inclusions are dominated by quartz, but rutile and more rarely clinozoisite are also found. Many porphyroblasts form δ- and γ-type asymmetrical structures and often develop a symplectitic rim of Ca-pyroxene and (or) amphibole + plagioclase (Fig. 4*a*). Garnet$_2$ appears as idioblastic, unaltered and inclusion-free porphyroblasts. Recorded garnet compositions are shown in Fig. 5. Garnet$_1$ ($Py_{4.0-13.9}Al_{53.1-64.8}Gr_{24.2-32.1}Sp_{0.4-14.4}$) and garnet$_2$ ($Py_{5.0-12.8}Al_{47.5-58.1}Gr_{26.0-31.4}Sp_{1.1-15.4}$) show broadly the same compositional spectrum and compare well with garnets recorded in many C-type (subduction-related) eclogites and amphibolite terrains (Coleman et al. 1965). No significant differences exist between primary or secondary garnets of the two studied areas in spite of differences in protolith bulk compositions (Meza-Figueroa 1998).

Pyroxene occurs in two distinctive domains: (1) Ca–Na (omphacite) pyroxene appears as small, subidioblastic crystals associated with garnet$_1$, phengite, rutile, epidote, quartz, and paragonite; and, (2) Ca (diopside) pyroxene appears in symplectitic overgrows around garnet$_1$ and omphacite associated with plagioclase and Ca-amphibole, rarely as isolated crystals following foliation. Generally, omphacite is rimmed by blue-greenish, Ca–Na amphibole, whereas diopside is fresh or little altered to greenish, Ca-amphibole. Recorded pyroxene compositions are plotted in the Jd–Ac–Di join shown in Fig. 6. Omphacite from Piaxtla ranges from $Jd_{35.7}Ac_{2.7}Di_{58.0}$ to $Jd_{39.2}Ac_{3.4}Di_{60.8}$ whereas omphacite from Mimilulco is slightly poorer in jadeite (Jd) and richer in diopside (Di) end members ($Jd_{20.4-29.2}Ac_{0.0}Di_{70.8-79.6}$). Such compositional differences will be addressed further in discussion. Diopsidic pyroxene is more homogeneous in composition ($Wo_{26.0-30.0}Fs_{15.2-19.8}En_{44.7-56.2}$), and no significant differences among secondary pyroxene from the two localities was found.

Amphibole is, together with garnet, the most abundant phase in the mafic rocks. It appears either as discrete, blue-greenish prismatic crystals in apparent textural equilibrium with garnet$_1$, omphacite and phengite or as elongate, green to colorless crystals defining mylonitic foliation. It further appears as rims around omphacite or forming symplectites after garnet$_1$ together with Ca-pyroxene and plagioclase. Blue-greenish, prismatic amphibole, and amphibole rimming omphacite show Ca–Na amphibole compositions from barroisite, magnesiokatophorite to magnesiotaramite, whereas green to colorless amphiboles in foliations and symplectites show Ca-amphibole compositions from edenite, pargasite, magnesiohornblende to actinolite (Fig. 7).

Plagioclase has a homogeneous albitic ($Ab_{95.8-99.7}$) composition. It appears as poikiloblastic crystals associated with Ca-amphibole, epidote, and phengite defining mylonitic matrix or as symplectitic overgrowths around garnet$_1$, therefore, it postdates the high-pressure mineral assemblage.

Clinozoisite (Cz) appears as elongate, prismatic crystals defining foliation, rarely as inclusions in garnet$_1$ porphyroblasts. Chemical data indicate that clinozoisite is characterized by varying, but rather low contents of the pistachite (Ps) end member ($Ps_{3-19}Cz_{81-97}$) and compares well with epidotes recorded in eclogite and amphibolite rocks (Yokoyama et al. 1986). Microgranular epidote has been recorded associated with chlorite and actinolite in many samples, but no analytical data are available. However, its structure and high birefringence indicate Fe-rich epidote.

White mica occurs as subidioblastic crystals apparently in textural equilibrium with garnet$_1$, omphacite, and barroisite. White mica formulae are based on 11 oxygens and all iron is assumed to be ferrous. White mica are phengites with high celadonite substitution (Si ≈ 3.3–3.5) in most eclogites from the Piaxtla and Mimilulco areas (Table 1).

Rutile is a common inclusion mineral in garnet$_1$, but it also appears as discrete crystals in matrix. Titanite was recognized by microprobe analyses and mainly occurs in coronitic reaction zones around rutile. Stilpomelane is also present. Chlorite appears as the main late alteration phase replacing some of the primary and secondary phases such as garnet, pyroxene, and amphibole.

Pressure–temperature conditions

Mineral texture and composition in eclogites from the Xayacatlan Formation suggest that an early high-pressure assemblage, involving omphacite, garnet of the type found in "Group C" eclogites, barroisite, and Si-rich phengite, was overprinted by a later higher temperature and lower pressure assemblage. Garnet-clinopyroxene and garnet-phengite

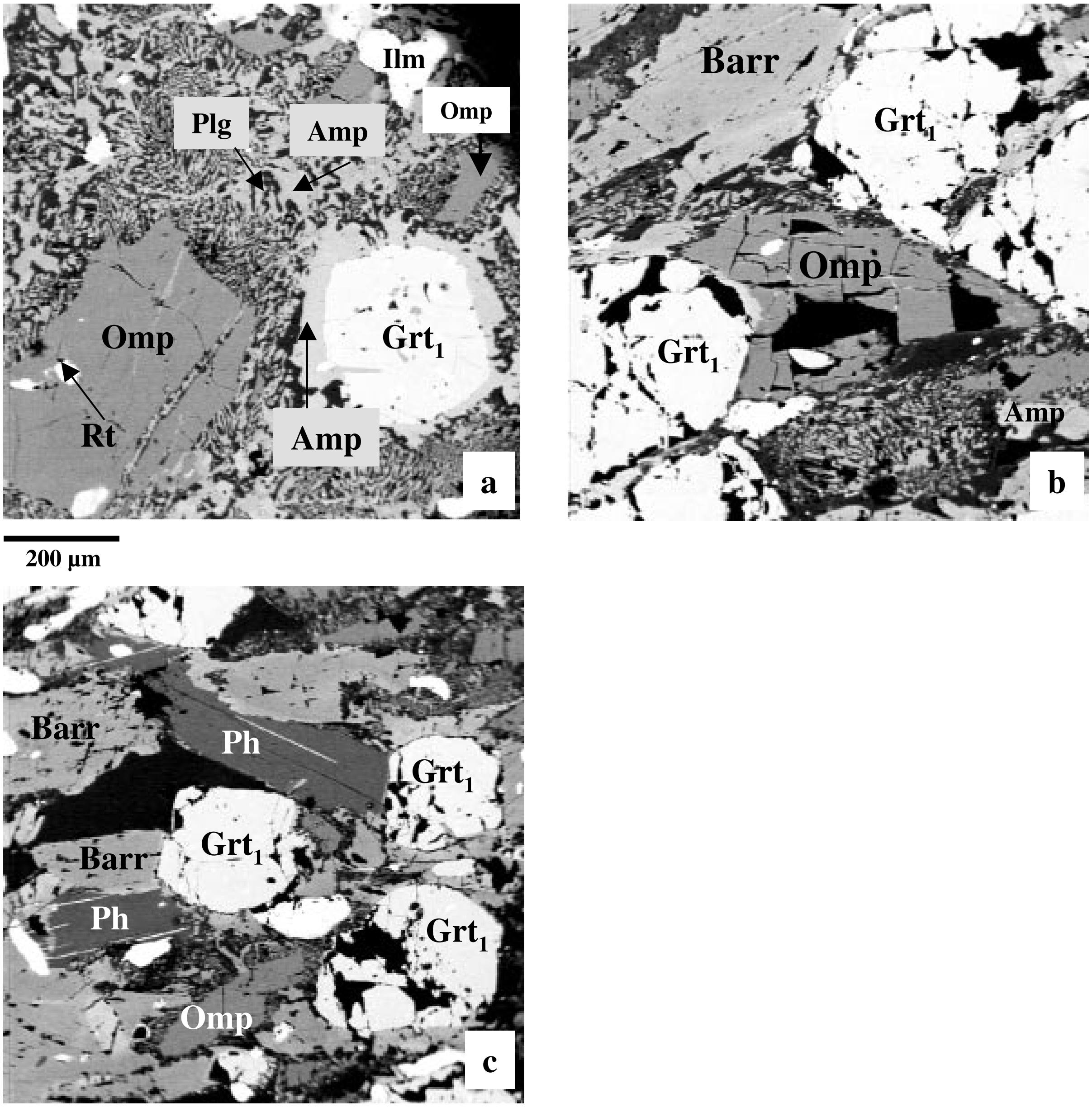


Fig. 4. Photomicrographs showing the relevant mineral assemblages and textural relationships of eclogites from the Acatlan Complex. (*a*) Polycristalline aggregate made of garnet$_1$ (Grt$_1$) porphyroblasts surrounded by an amphibole (Amp) rim of magnesiokatophorite. Symplectites of plagioclase (Plg) – amphibole (magnesiokatophorite) are developed between omphacite (Omp) – garnet$_1$. Rt, rutile; Ilm, ilmenite. (*b*) Polycristalline aggregate made of coarse-grained omphacite, garnet$_1$ and barroisitic amphibole (Barr). (*c*) Polycristalline aggregate of omphacite – phengite (Ph) – garnet$_1$ – barroisite (Barr) with a granuloblastic microstructure.

thermometry were conducted on high-pressure mineral assemblages (garnet$_1$ and omphacite) of eclogites least affected by retrogression. The results for the geothermometers used are summarized in Table 3. The results, particularly the garnet-clinopyroxene temperatures, are strongly dependent on the calculated Fe^{2+} and Fe^{3+} values. For this work, we applied a correction for ferric iron content of natural clinopyroxene based on microprobe analyses as follows: Fe^{3+} = 4–2Si–Al+Na (Ryburn et al. 1976), neglecting insignificant amounts of K, Cr, and Ti. This amount of ferric iron is substracted from the total iron of the microprobe analysis, usually recorded as FeO.

The obtained average temperature using the calibration of Ellis and Green (1979) is 565 ± 37°C. This overlaps with the temperatures estimated from the correction by Ganguly (1979) and falls slightly above the 520 ± 18°C and 580 ± 45°C for Piaxtla and Mimilulco, respectively, obtained from the calibration of Raheim and Green (1974). The garnet-phengite geothermometer based on Krogh and Raheim (1978) yielded an average temperature of 530 ± 20°C for rocks from Mimilulco. A temperature range of 560 ± 60°C is, therefore, considered a reasonable approximation to the actual temperature of eclogite formation.

It is increasingly accepted that partitioning of Al^{IV} in chlorite reflects crystallization temperature in hydrothermal systems (e.g. Cathelineau and Nieva 1985; Cathelineau 1988; Schiffman and Friedleifsson 1991) and low-grade metamorphic terrains (Bevins et al. 1991). Chlorite geothermometry calculated by

Table 2. Representative analyses of chlorite, epidote, and feldspar group from retrogressed eclogites from the Acatlan Complex, southern Mexico.

	Chlorite				Plagioclase			Epidote			
Sample	MI-13	MI8a 71	MP1	MP3	MP3	MI8a	M7	MP1	MP1	MI6	MI8a
Analyses	1	2	3	4	5	6	7	8	9	10	11
SiO_2 (wt.%)	26.26	26.67	27.78	28.06	69.13	70.57	70.62	39.59	39.25	38.92	39.24
Al_2O_3	20.40	19.39	22.40	20.99	21.27	20.50	20.45	32.17	29.86	29.12	28.29
TiO_2	0.08	0.01	0.01	—	—	—	—	0.01	0.01	0.02	0.17
FeO	24.11	31.66	20.57	20.57	0.34	0.18	0.30	2.56	5.37	6.18	6.89
MgO	16.98	11.92	19.99	19.29	—	—	—	0.03	0.02	0.01	0.03
CaO	0.04	0.06	0.04	0.12	2.21	0.05	0.37	25.58	26.21	25.57	23.52
MnO	0.31	0.18	0.30	0.31	—	—	—	0.01	0.11	0.19	—
Na_2O	—	0.02	0.02	0.01	10.23	10.66	10.74	—	—	0.02	—
K	—	—	—	—	—	—	0.05	—	—	—	0.01
Total	88.2	88.3	91.1	89.4	103.1	101.9	102.3	100.0	100.8	100.0	98.15
	28 oxygens				8 oxygens			12.5 oxygens			
Si	5.467	5.448	5.457	5.628	2.591	3.004	2.997	2.972	2.963	2.967	3.032
Ti	0.012	0.002	0.001	—	—	—	—	0.001	0.001	0.001	0.010
Al	5.007	4.962	5.186	4.961	1.064	1.028	1.023	2.846	2.657	2.617	2.576
Fe	4.198	5.750	3.378	3.450	0.012	0.007	0.011	0.144	0.304	0.354	0.400
Mn	0.054	0.033	0.050	0.053	—	—	—	0.005	0.006	0.011	—
Mg	5.270	3.858	5.853	5.765	—	—	—	0.003	0.002	0.001	0.004
Ca	0.008	0.013	0.008	0.025	0.101	0.002	0.017	2.057	2.12	2.089	1.947
Na	—	0.006	0.007	0.002	0.842	0.880	0.884	0.001	—	0.003	—
K	—	—	—	—	—	—	0.002	—	—	—	0.001
Total	20.01	20.07	19.94	19.88	4.61	4.92	4.93	8.03	8.05	8.04	7.97
T_1 (°C)	346	349	348	320	—	—	—	—	—	—	—
T_2 (°C)	344	342	354	335	—	—	—	—	—	—	—
Al_{IV}	2.533	2.552	2.543	2.372	—	—	—	0.028	0.037	0.033	—
Al_{VI}	2.475	2.410	2.642	2.588	—	—	—	2.818	2.619	2.584	2.576
Na/(Na+K)	—	—	—	—	—	—	0.997	—	—	—	—
% Cz	—	—	—	—	—	—	—	86.1	95	87.8	86.6
% Ps	—	—	—	—	—	—	—	13.6	4.8	11.9	13.4

Note: MP3 and MP1 are basic eclogites from Piaxtla; MI6, M7, MI8a, and MI-13 are retrogressed eclogite-facies rocks from Mimilulco; Cz, clinozoisite; Ps, pistacite.

this method indicates temperatures in the range of 340 ± 10°C for the overprinting greenschist metamorphism. This overlaps with the temperatures estimated from the plagioclase–amphibole geothermometer of Spear (1980) and obtained for the epidote–amphibolite geothermometer and greenschist-facies metamorphism (Table 3; Figs. 8*a*, 8*b*).

Geobarometry was carried out in the eclogites using the Si content per formula unit of phengite (Si ≈ 3.3–3.5) as proposed by Massonne and Schreyer (1987). Estimates indicate minimum pressure values of 11–13 kbar (1 kbar = 100MPa) for Piaxtla and 10.5–14.5 kbar for Mimilulco eclogites (Table 3). These results are similar to those obtained from jadeite content in pyroxene (Holland 1980), which yielded a pressure range of 12–14 kbar (Table 3; Fig. 9). The minimum pressures of recrystallization of quartz eclogites lacking plagioclase may be calculated by the method of Newton and Perkins (1982), as documented by Newton (1986). The Newton and Perkins'(1982) geobarometer was applied to the eclogites yielding pressures of 14–16 kbar, similar to the range estimated by the other methods (Table 3).

To better constrain a pressure range for the eclogites from the Acatlan Complex, the garnet–clinopyroxene–phengite geobarometer calibration of Holland and Powell (1990) and Waters and Martin (1993) was used. This calibration involved only three phases: $garnet_1$, clinopyroxene and phengite. It is independent of water activity and of silica saturation. No Fe end members are involved, so that the uncertainty related to formula recalculation for Fe^{3+} in clinopyroxene and phengite is minimal. The obtained pressure range was from 15–17.7 kbar for samples from Piaxtla. This barometer calibration overestimates the experimental pressures by a little over 3 kbar according to activity models based on Holland and Powell (1990), considering this, the obtained pressures fit within the range previously defined by the other geobarometers. Application of the geobarometer by Kohn and Spear (1991) to barroisitic amphiboles from Piaxtla eclogites, which are in textural equilibrium with unzoned $garnet_1$, yielded a pressure of 12 kbar. The average pressure range for eclogites from Mimilulco is 11–15 kbar and the average pressure range for those from Piaxtla is 11–13 kbar (Table 3).

Pressures for the epidote–amphibolite and greenschist assemblages have been determined with the empirical

Fig. 5. Ternary diagram showing composition of garnet crystals from the Acatlan Complex. A, B, and C are fields for eclogite groups based on Coleman et al. (1965)

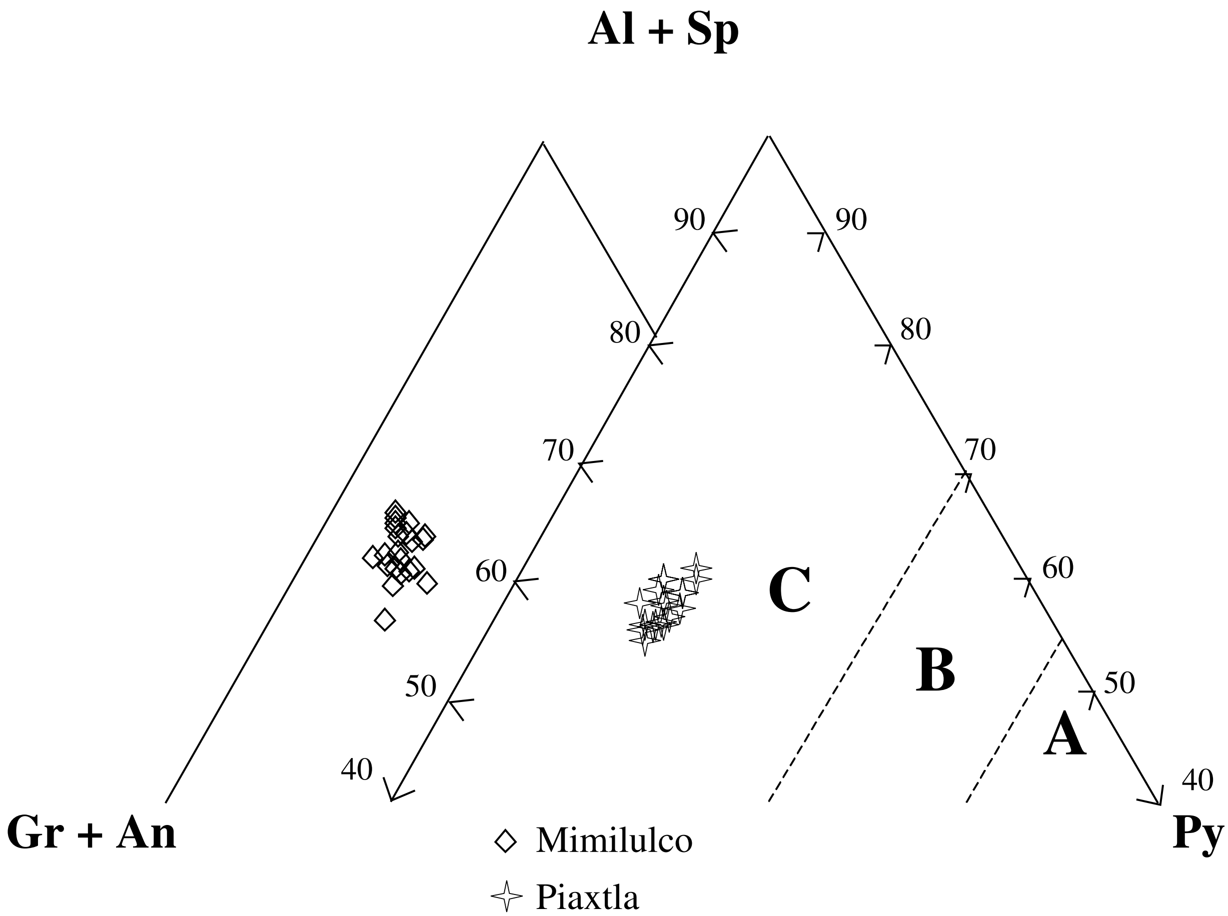


Fig. 6. Composition of metamorphic pyroxenes (formulae calculated on the basis of 6 oxygens).

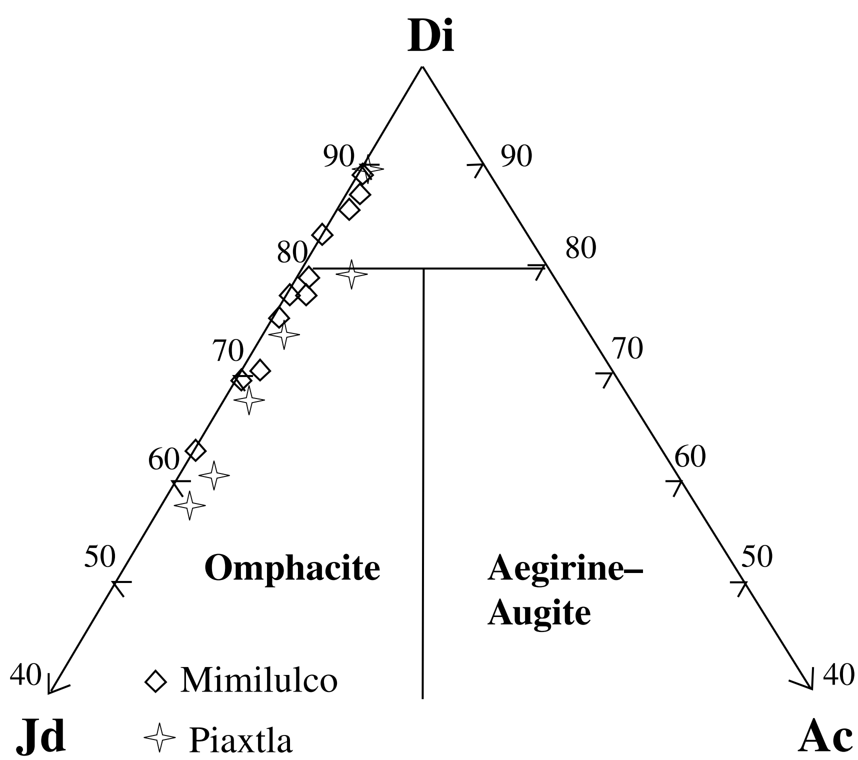


Fig. 7. Leake et al. (1997) classification scheme for calcic–sodic and calcic amphiboles from eclogitic rocks from the Acatlan Complex.

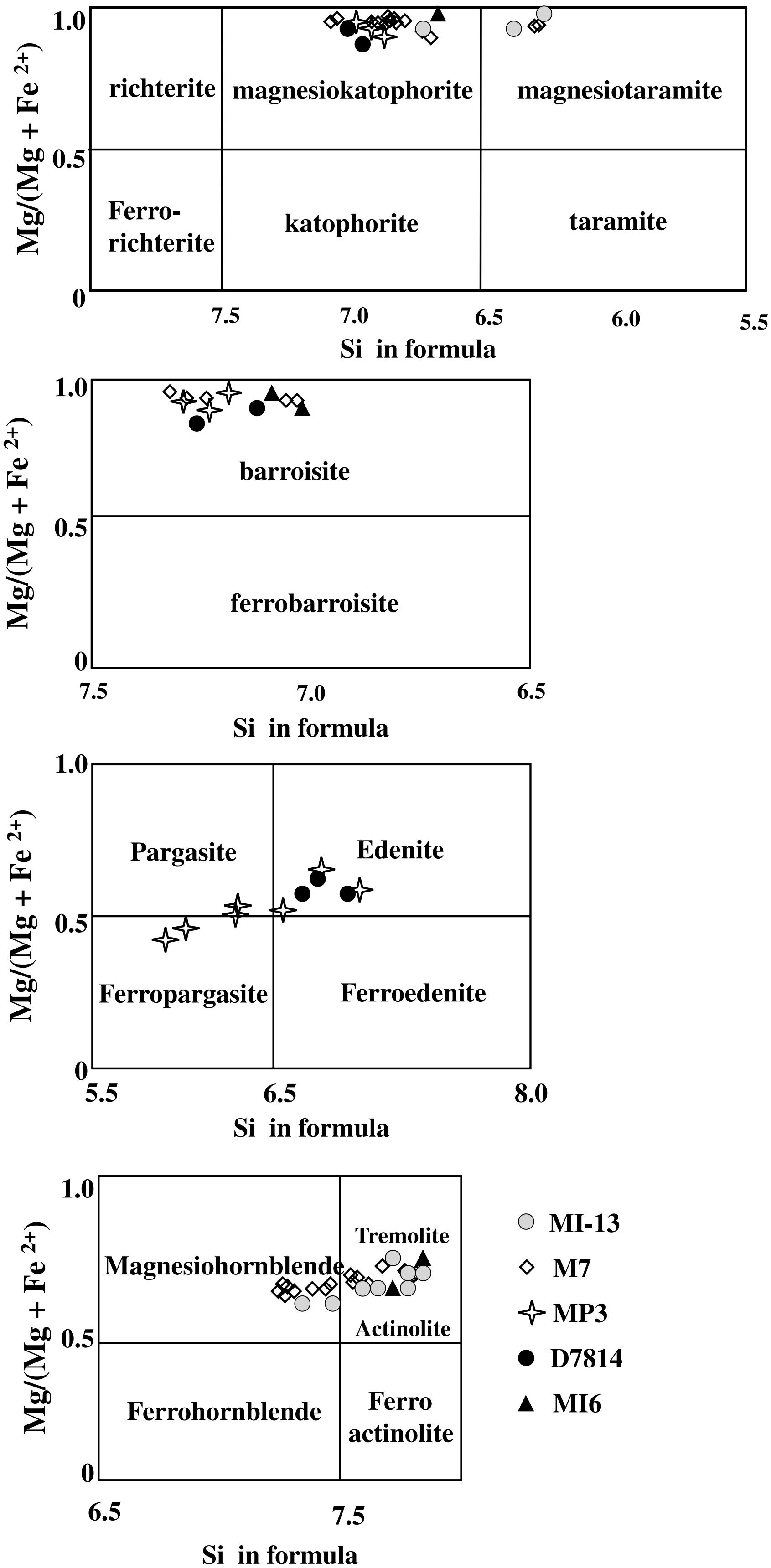


Table 3. Geothermobarometry of eclogites and retrogressed eclogites from the Acatlan Complex, southern Mexico.

	Garnet–Clinopyroxene					Garnet–Phengite	
Sample	X_{Ca}^{Grt}	$\ln K_D$	*T* (°C) (9)	*T* (°C) (10)	*T* (°C) (13)	*T* (°C) (12)	
MP3	0.295	3.05	542 ± 16.0	496 ± 41.7	532 ± 15.2	MI8b	434 ± 41.1
MI8b	0.324	2.69	625 ± 18.0	518 ± 42.8	588 ± 16.3	M3b	571 ± 49.0
MP3	0.307	2.98	561 ± 16.5	485 ± 41.1	541 ± 15.4	M3b	535 ± 47.0
MI6a	0.339	3.03	574 ± 16.5	460 ± 39.7	535 ± 15.2	MP3	488 ± 44.2
MI13	0.301	2.07	577 ± 17.0	559 ± 45.1	542 ± 15.7	MP3	483 ± 44.0
MI13	0.300	2.75	578 ± 16.1	584 ± 46.4	578 ± 16.1	M3b	600 ± 51.0
MI13	0.332	3.00	574 ± 16.5	544 ± 44.3	539 ± 15.4		
MI13	0.379	3.15	578 ± 16.0	520 ± 43.0	518 ± 14.9		
M7	0.300	2.75	600 ± 17.5	—	578 ± 16.1		

Geothermometry

	Eclogite					Epidote–Amphibolite		
	(9)	(10)	(11)	(12)		(5)	(6)	
Mim_{ave}:	565 ± 37	580 ± 45	560 ± 45	535 ± 45	Mim_{ave}:	—	490 ± 20°C	ede
Px_{ave}:	560 ± 20	520 ± 18	544 ± 14	486 ± 44	Px_{ave}:	500°C	510 ± 20°C	ede

Geobarometry

	Eclogite							
	X_{Jd}	Si_{Phen}	*P*(kbar)	Ref	Sample	X_{Jd}	Si_{Phen}	
Mim_{ave}:	20–44%	—	10.5–15	(1)	MI6	19.8	3.50	
Px_{ave}:	36–39%	—	13–15	(1)	MP3	39.3	3.40	
					MP3	36.0	3.48	
Mim_{ave}:	20–44%	—	11–13.5	(2)	MI13	43.0	3.40	
Px_{ave}:	36–40%	—	12–13	(2)	M7	22.3	3.36	
					M7	24.0	—	
Mim_{ave}:	—	3.36–3.45	10.5–14.5	(3)	MI8b	—	3.40	
Px_{ave}:	—	3.40–3.48	11–13	(3)				
					Sample	P_{ave}(Fe)		P_{ave}(Mg)
Mim_{ave}:	—	—	—	(7)	MP3	12.9 ± 0.3		13 ± 0.5
Px_{ave}:	—	—	13	(7)	MP3	13 ± 0.4		12.7 ± 0.5
						lnK		*P* (kbar)
Mim_{ave}:	—	—	11	(8)	MI-13	17.33		10.8
Px_{ave}:	—	—	16 ± 2	(8)	MP3	11.07		17.7
					MP3	15.35		14.1

Note: X, Molar fraction; K_D, equilibrium constant, Px, Piaxtla; Mim, Mimilulco; $_{GS}$, greenschist facies; $_{Amph}$, epidote–amphibolite facies; ede, edenite; $_{Jd}$, jadeite; Piaxtla; MI8b, MI6, MI13, M7, and M3b are retrogressed eclogite-facies from Mimilulco. (1) Holland (1980) and Gasparik (1985); (2) Powell (1985); (3) and Waters and Martin (1993); (9) Ellis and Green (10) Raheim and Green (1974); (11) Krogh (1988); (12) Krogh and Raheim (1987); (13)

method of Brown (1977). Pargasitic and edenitic amphiboles of the epidote–amphibolite stage indicate approximate values of 6–7 kbar, while the greenschist actinolites suggest a pressure of about 3.5–4.0 kbar (Fig. 10, Table 3). Graphic representation of geothermometers and geobarometers used is shown in Fig. 10.

Discussion

Some eclogites from the Acatlan Complex show apparent textural equilibrium among garnet$_1$, omphacite and barrositic amphibole (peak metamorphic assemblages, Fig. 4*b*). It has been documented that amphiboles that are apparently compatible with garnet and clinopyroxene appear in most group B and C eclogites (Laird and Albee 1981; Newton 1986). Such rocks are referred as amphibole eclogites (Newton 1986). This implies the availability of fluids during metamorphism and "wet" protoliths.

A greenschist assemblage of amphibole, epidote, and other minerals replacing garnet and omphacite is common (Maresch and Abraham 1980; Newton 1986). This may result from continuous reaction during uplift from deep burial or from one or more discrete later episodes of lower pressure metamorphism; these contrasting processes are often hard to distinguish in their effects, and it may, in addition, be hard to tell by textures whether some of the minor minerals in an eclogite were compatible with garnet and omphacite during eclogite-facies conditions or were formed in the uplift process.

Textural and mineralogical evidence indicate three metamorphic episodes in the Acatlan mafic eclogites: (1) an early episode defined by the assemblage garnet$_1$ + omphacite + phengite$_1$ + rutile ± zoisite–clinozoisite ± quartz ± Ca–Na amphibole (M_1) (Fig. 4*c*); (2) a retrograde formation of the

Table 3 (*concluded*).

Plagioclase			Amphibole				
X_{Ca}	X_{Na}	ln(An/Ab)	XNa_B	XCa_B	XAl_{IV}	ln(Ca/Na)	
MP3	0.101	0.842	–2.121	0.272	1.55	1.682	1.74
MP3	0.033	0.838	–3.23	0.023	1.868	1.159	4.40
MP3	0.033	0.838	–3.23	0.479	1.376	1.103	1.06
MP3	0.101	0.842	–2.121	0.143	1.622	2.082	2.41
M7	0.015	0.859	–4.07	0.312	1.519	0.309	1.58
MI13	0.010	0.797	–5.07	0.023	1.945	0.135	4.43
MI8a	0.002	0.880	–6.08	0.157	1.838	0.355	2.46

					Greenschist			
					(13)			
					330 ± 25	act–plg		
					300 ± 25	act–plg		
					Geothermometer			
	Amphibolite				Chlorite (14)			
	Na(M4)	Al_{IV}	Sample		Sample	Al_{IV}	T_1 (°C)	T_2 (°C)
	0.272	1.682	MP3		MI13	2.533	346	344
	0.023	1.159				2.508	342	338
	0.479	1.103				2.491	339	336
	0.435	0.959	M3b		MI8b	2.496	340	340
	0.509	0.761				2.436	330	333
	0.727	0.939				2.518	343	345
	0.023	0.135	MI13			2.466	335	338
	0.157	0.355			MP1	2.543	348	354
					2.372	320	335	
P_{amph} (kbar)	4–6	(4)	M3b	2.475	336	341		
P_{amph} (kbar)	6	(5)		2.466	335	337		
				M7	2.495	340	353	
				$Mim_{ave\ (GS)}$:	339°C	341°C		
				$Px_{ave\ (GS)}$:	334°C	345°C		
				P_{GSave}:	2–3 kbar	(4)		

$_{Phen}$, phengite; act–plg, actinolite–plagioclase; $_{ave}$, average; ^{Grt}, garnet; *T*, temperature; *P*, pressure; Ref, reference; MP3 and MP1, basic eclogites from Massonne and Schreyer (1987); (4) Brown (1977); (5) Plyusnina (1982); (6) Spear (1980); (7) Kohn and Spear (1991); (8) Holland and Powell (1990) Ganguly (1979); (14) Cathelineau (1988).

assemblage garnet$_2$ + Ca-pyroxene + Ca-amphibole + plagioclase ± phengite$_2$ ± epidote ± quartz (M_2); and then (3) a low-grade metamorphism characterized by the assemblage Ca-amphibole + plagioclase + chlorite + epidote (M_3).

Internally, eclogites from the Xayacatlan Formation, show the effects of at least three phases of penetrative deformation (Fig. 3). The oldest recognized phase of deformation (D_1) is visible as a foliation (S_1) defined by quartz, rutile, titanite, ilmenite, and, sometimes phengite and zircon, and has been preserved in garnet and rutile porphyroblasts. Its corresponding folding phase (F_1) may be represented by isoclinal folds, sometimes present in the limbs of later enclosed folds assigned to the second phase of deformation. Commonly, the inclusions defining S_1 curve in spiral or spherical patterns implying that the eclogite-facies metamorphism (M_1) partly evolved syntectonically with D_1. This oldest deformation presumably happened in the Silurian to Early Devonian according to geochronological and structural data (Yañez et al. (1991; Weber et al. 1997).

The second phase of deformation (D_2) was accompanied by mylonitization and produced a penetrative, ductile foliation (S_2) and associated subisoclinal to tight folds with sharp hinges. The mylonitic foliation is commonly axial to these folds and altogether define the dominant trend of the regional foliation and tectonic trend of the eclogitic belts. Because garnet grains and, in general, all the high-pressure phases are wrapped by the S_2 foliation, it is inferred that this phase of deformation postdated the high-pressure metamorphism. The main retrogressive metamorphic event (M_2) probably accompanied the second phase of deformation, which may be linked to the emplacement of the eclogites above the lower grade quartzites and phyllites of the Cosoltepec Formation. The third penetrative deformation (D_3) developed a spaced subvertical crenulation cleavage axial to north-trending regional

Fig. 8. (*a*) Plot of Al_{IV} vs. temperature for chlorites from eclogites affected by greenschist metamorphism from the two studied areas, based on the geothermometer of Cathelineau (1988); (*b*) plagioclase–amphibole geothermometer after Spear (1980) showing temperatures for edenite–magnesiohornblende from eclogites affected by amphibolite facies.

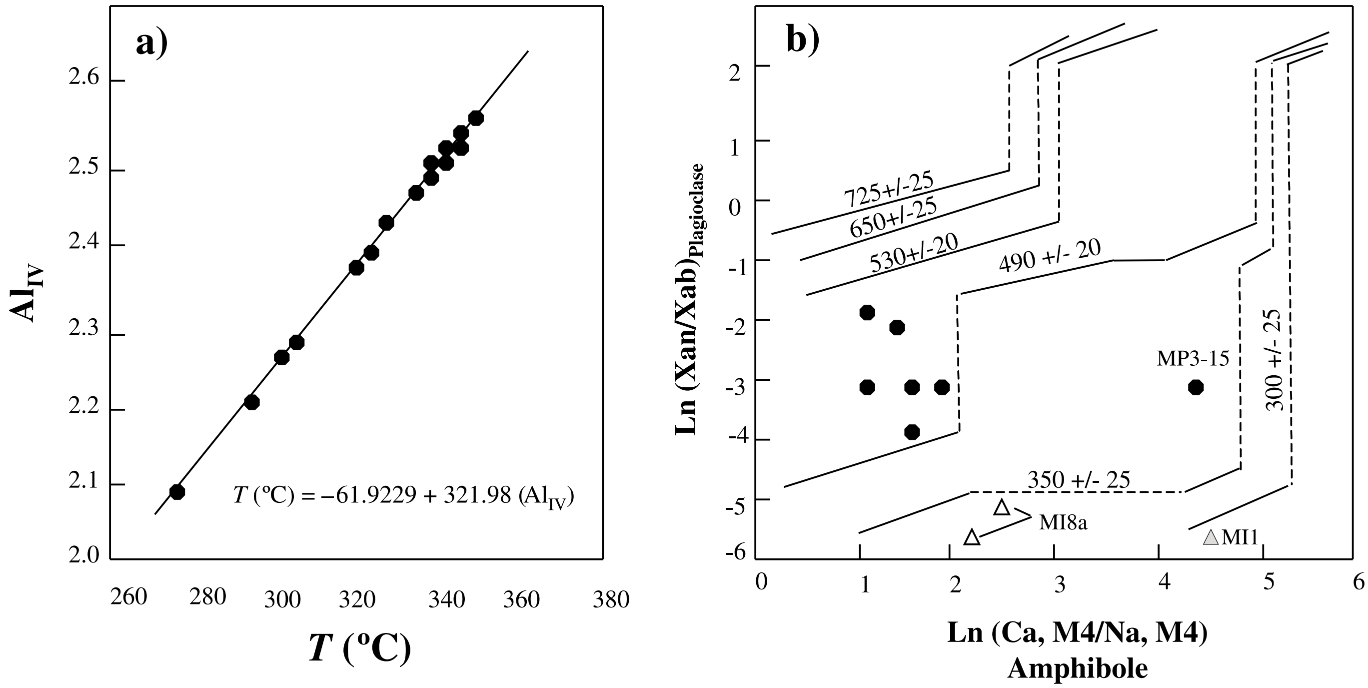
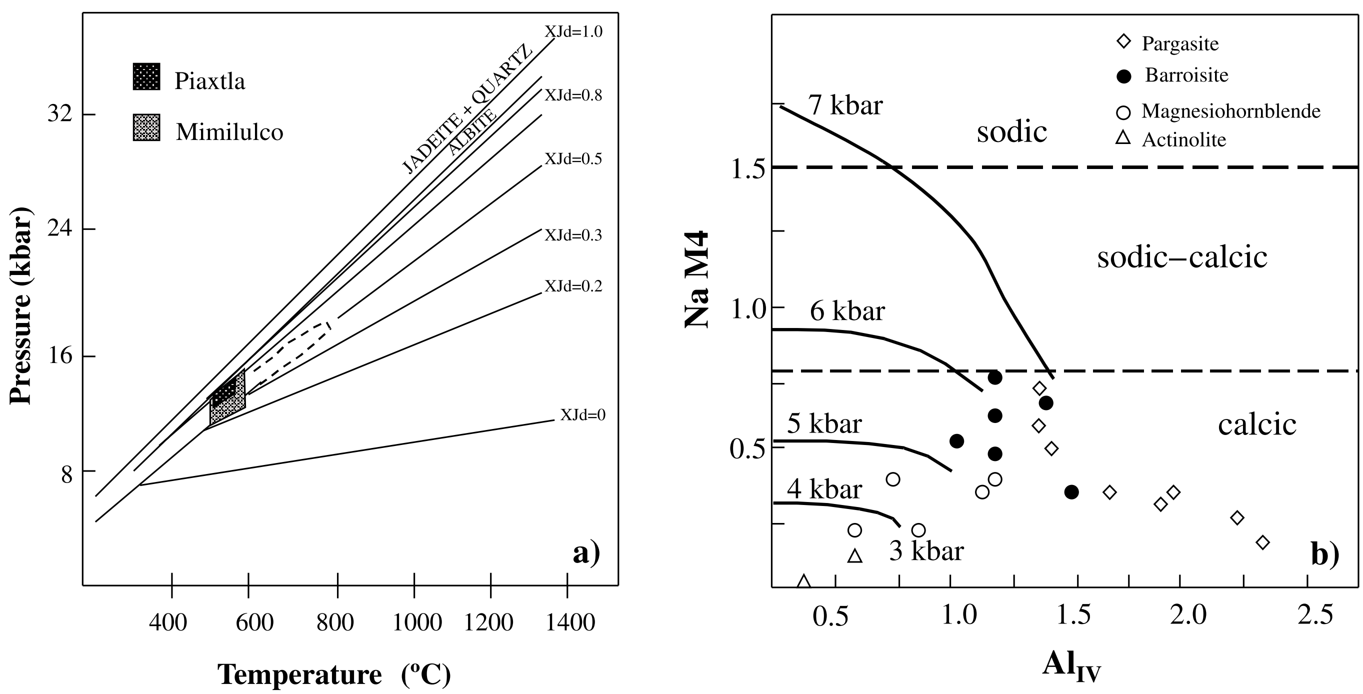


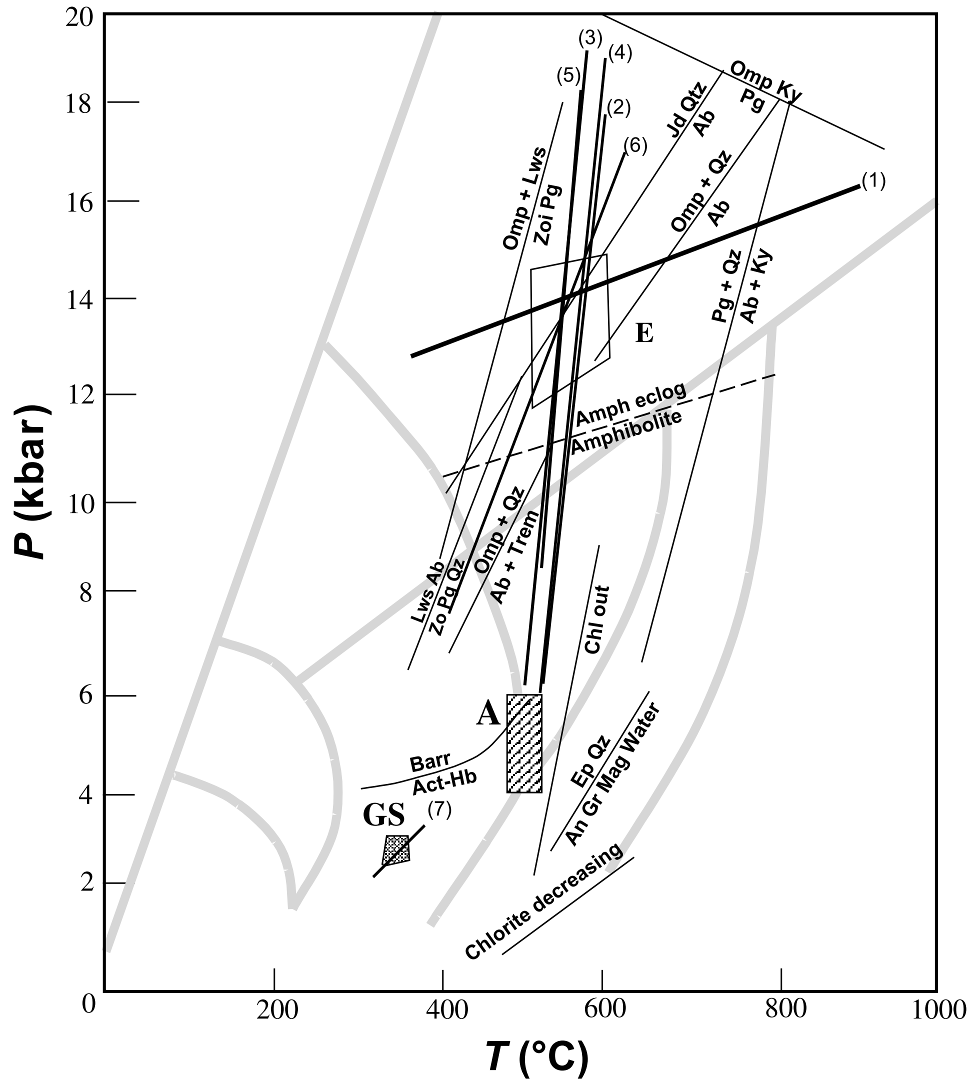
Fig. 9. (*a*) Univariant reaction curve for albite = jadeite + quartz (after Holland 1980) and isopleths of jadeite content in disordered (C2/c) and ordered (P2/n) clinopyroxenes coexisting with albite and quartz. (*b*) Geobarometry of amphiboles after Brown (1977).



folds of open to tight profile geometry. Interference patterns are commonly associated with this last intense deformation that affected the Acatlan Complex and its eclogitic rocks during Late Carboniferous to Permian.

Mineral assemblages record only the peak metamorphic conditions, and the retrograde *P–T* path of the multistage history of the Acatlan Complex eclogites. Prograde path is inferred based on the presence of blue-amphibole (glaucophane) found in blueschists preserved as lenses within greenschist associated to garnet metabasites from the Xayacatlan Formation

Fig. 10. Petrogenetic grid for the eclogitic, amphibolite, and greenschist metamorphism in the Acatlan Complex metabasites. Variation of equilibration temperature with temperature and with pressure, calculated using the relations proposed by (1) Kohn and Spear (1991), (2) Ellis and Green (1979): Mimilulco average, (3) Ellis and Green (1979): Piaxtla average, (3) Raheim and Green (1974), (5) Ganguly (1979) average, (6) Krogh and Raheim (1987) average, (7) Cathelineau (1988), Spear (1980) and Brown (1977), (8) Plyusnina (1982), (9) Brown (1977) and Spear (1980). GS, greenschist; E, eclogite; A, amphibolite; Amph eclog, Amphibole eclogite (garnet + clinopyroxene + amphibole + water); Amphibolite, amphibole + plagioclase ± garnet ± clinopyroxene, from Newton (1986).



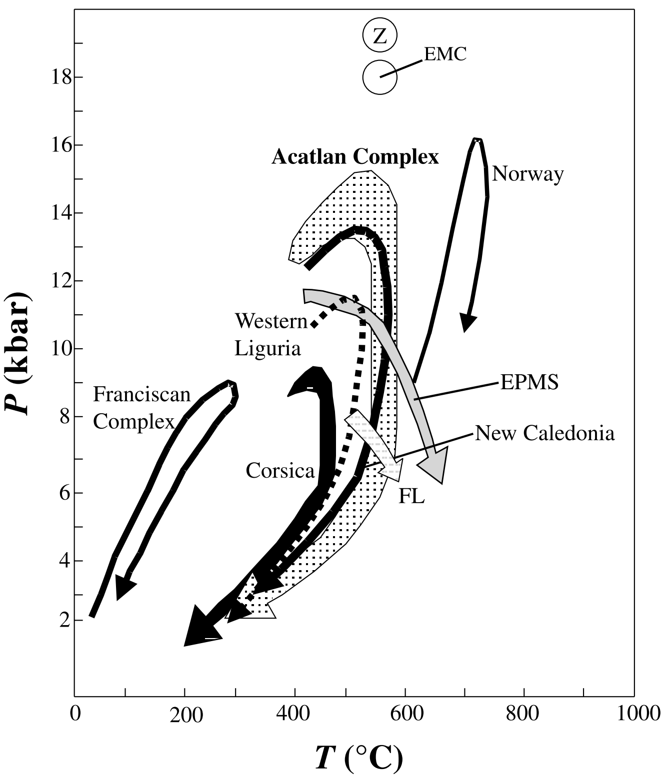
in the western part of the Acatlan Complex (Meza-Figueroa 1998).

The decompressional evolution postdating the eclogitic climax is evidenced by different amphibole generations, since they recrystallized under varying pressure–thermal conditions (M_2 and M_3). This indicates reequilibration of high-pressure relics within an intermediate *P–T* regime, followed by an overall retrogression by cooling during uplift. The metamorphic evolution is summarized by the *P–T* path on Fig. 11.

P–T paths give important information regarding convergent tectonics (Ernst 1988). Figure 11 also shows *P–T* paths for different high-pressure metamorphic terranes with a marked greenschist–amphibolite-facies overprint. The Alpine-type path (e.g., Western Liguria and Corsica) is characterized by nearly isothermal decompression, which can only be caused by very rapid exhumation, this mechanism places hot rocks in shallow crustal levels and metamorphic overprinting occurs at low pressure. The rapid exhumation is explained by the shift from the subduction of oceanic crust to the collision and attempted subduction of buoyant sialic crust. Subduction will continue only until a significant crustal mass, such as when an island arc, continental fragment or a continent collides and impinges on the subduction zone. On the other hand, Franciscan-type paths require the rocks to be cooled as they are exhumed. This requires a very slow exhumation rate so that the rocks mantain thermal equilibrium with their surroundings (Ernst 1988).

The determined *P–T* trajectory for the Acatlan Complex metamorphic rocks is somewhat intermediate between the discussed Franciscan and Alpine paths (Fig. 11). The obtained data indicates that the Acatlan Complex *P–T* path was not produced by simple subduction; it neither presents typical very high-pressure mineralogical assemblages found in continental collision involving large masses (Fig. 11, *P–T*

Fig. 11. *P–T* diagram based on present study. *P–T* paths from subduction and continental collision complexes are shown for comparison. Data from Ghent et al. (1987), Jamieson (1990), and Carswell (1990). Z, Zermatt-Saas zone (metaophiolites); EMC, eclogitic micaschist complex of the Sesia zone; EPMS, East Pond Metamorphic Suite; FL, Fleur de Lys Supergroup.



estimates for Z: Zermatt-Saas zone and EMC: eclogitic micaschist complex of the Sesia zone). The eclogites from the Acatlan Complex plot well along a trend defined by complexes, such as Fleur de Lys Supergroup, East Pond Metamorphic suite, Sifnos, New Caledonia, and Cabo Ortegal, which are characterized by peak metamorphic pressures below 16 kbar. Most alpine eclogites plot above this trend, and they present a development or coexistence of paragonite and kyanite with the rest of the high-pressure mineral assemblages. This feature has not been found in eclogites from the Acatlan Complex.

High-pressure rocks associated with the peri-Gondwanan terrane in the northern Appalachians have been reported in New Brunswick (Brunswick subduction complex), where the Gander and Dunnage terranes continuously interact to develop a complex evolution (van Staal 1987; van Staal and de Roo 1995). Even though the high-pressure rocks of this region (blueschist) can not be directly compared to those of the Acatlan Complex, the emplacement of the Dunnage terrane matches the relationship between the Piaxtla Group (Dunnage-type terrane) over the Petlalcingo Group (Gander-type terrane) of similar age (Ramírez-Espinosa 2001).

High pressure rocks of similar age are present in the Caledonides in Greenland and Scandinavia (collisional process between Laurentia and Baltica). There are also high-pressure rock in Iberia and France, but they are related to the collision of Armorica against Baltica during the late Paleozoic (Strachan et al. 1995; Robinson et al. 1988).

Intraoceanic arcs during the Ordovician evolution of the Iapetus Ocean are clearly represented by the Dunnage and Piedmont terranes along the length of the Appalachians. In the Acatlan Complex, rocks older than Silurian are the Xayacatlan Formation (which is intruded by the 440–425 Ma Esperanza Granitoids) and the Petlalcingo Group, which is correlated with the Cambrian – Middle Ordovician siliciclastic miogeocline of western Gondwana (Ramírez-Espinosa 2001). These sequences contain metavolcanic rocks with MORB and OIB geochemical signatures, closely associated to quartzite and schist (Meza-Figueroa 1998). According to Ramírez-Espinosa (2001), those groups could represent different regions of the same oceanic plate within a distal passive margin or represent different parts of a more complex setting, including an intraoceanic arc (Meza-Figueroa 1998).

Based on the geological similarities between the peri-Gondwanan region of the northern Appalachians and the Piaxtla and Petlalcingo Groups, Ramírez-Espinosa (2001) suggested that a possible location of the Acatlan Complex could be southward of the Brunswick Complex following the Caledonian high-pressure trend. The main difference with this region is the vergence of folding and thrusting: southeastward in the Dunnage–Gander relationship and northwestward in the Acatlan Complex units.

In the Acatlan Complex, the main phase of metamorphism is apparently Taconian (Ortega-Gutierrez et al. 1999). However, the precise timing of the various stages of metamorphism remain uncertain. Yañez et al. (1991) reported peak metamorphism to be Acadian. Whether these Silurian and Devonian ages represent minimum age of metamorphism or reflect subsequent overprinting by an Acadian thermal event is subject of a current investigation.

Conclusions

Pressure and temperature peak metamorphic conditions are 11–15 kbar and 565 ± 37°C for metaeclogites from Mimilulco similar to those of Piaxtla. Pressure and temperature estimates obtained for the epidote–amphibolite and the greenschist facies allowed for the construction of a *P–T* path for the complex. On the basis of textural and mineral associations, at least three simplified metamorphic stages can be defined: (1) an eclogitic stage defined by high-pressure mineral associations; (2) an epidote–amphibolite stage characterized by zoisite–clinozoisite – pargasite and edenite, which represents the earliest decompressional assemblage; and (3) actinolite–albite–epidote–chlorite latest stage of reequilibration, which corresponds to the greenschist facies.

The Acatlan Complex metaeclogites are similar to low–intermediate-temperature complexes which are not associated to classical continental-collisional scenarios. This does not discard the possibility of a continental-collisional scenario for the Acatlan Complex eclogites, but the data suggest that if it occurred it would represent a more complex continental-collisional setting, including intraoceanic arcs, than previously proposed models. Further studies should be conducted in associated rocks to the eclogites at a regional scale to clarify this.

Based on the *P–T* path, as well as on the nature of the protolith, the petrogenetic history of the Acatlan eclogites can be summarized as follows:

(1) Extrusion of OIB and MORB.
(2) Collision of the plates (?) and formation of the stage E eclogite (Acatecan collisional orogeny) assemblage at minimum conditions P = 11 kbar and T = 560 ± 60°C and maximum pressure of almost 15 kbar.
(3) Slow rise and destabilization of omphacites, because of the release of pressure, to produce the early symplectite of plagioclase–amphibole.
(4) Sudden ascent toward the crust and disruption and tectonic emplacement within the disrupted crustal sequence.
(5) Healing of the sequence and partial reequilibration within upper-crustal conditions (6 kbar) marked by complete amphibolitization of the small relics and formation of symplectites in the largest ones.
(6) Reequilibration to greenschist facies (343°C, 3.5 kbar) probably during Mississippian times.

The island-arc basalt, MORB, and OIB eclogite transitions result from a long and complex history. The evolution scheme seems now to be partially clarified, but many important issues are still unclear and must await further investigations.

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