TECTONO-MAGMATIC EVOLUTION OF THE PALEOZOIC ACATLAN COMPLEX IN SOUTHERN MEXICO, AND ITS CORRELATION WITH THE APPALACHIAN SYSTEM

by

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A Dissertation submitted to the Faculty of the

DEPARTMENT OF GEOSCIENCES

In Partial Fulfillment of the Requirements

For the Degree of

DOCTOR OF PHILOSOPHY

In the Graduate College

THE UNIVERSITY OF ARIZONA

2001

UMI Number: 3031403

THE UNIVERSITY OF ARIZONA ®
GRADUATE COLLEGE

As members of the Final Examination Committee, we certify that we have read the dissertation prepared by Joel Ramírez Espinosa

entitled Tectono-magmatic evolution of the Paleozoic Acatlán Complex in Southern Mexico, and its correlation with the Appalachian System

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ACKNOWLEDGMENTS

I wish to acknowledge the financial support given by the Universidad Autónoma de Guerrero and Consejo Nacional de Ciencia y Tecnología (CONACyT) during my graduate career. I also wish to thank the community of the Escuela Regional de Ciencias de la Tierra in Taxco, Gro., for the logistic help during the fieldwork, but more strongly for all the moral support. I want to express my gratitude to Miguel Miranda for his friendship and financial support, through his help; BHP minerals financed part of the laboratory expenses.

I would like to thank all my professors at the University of Arizona for their teaching during my graduate studies, but a special debt of gratitude goes to Peter J. Coney for transmitting his motivation and knowledge in regional tectonics. I am really grateful to Joaquín Ruiz for his academic support, advice and friendship throughout my graduate career. I like to specially thank Bo Baylor for her enormous help. I also thank the members of my dissertation committee, Jonathan Patchett, George Gehrels, Mihai Ducea and Elena Centeno; their generous comments were very helpful during the development of this manuscript.

I thank the following people for insightful discussions on my research: Fernando Ortega Gutiérrez, Joaquín Ruiz, Jonathan Patchett, Elena Centeno, Fernanda Campa, Martín Guerrero, Antonio Flores, Miguel Miranda, José Luis Sánchez Zavala, Oscar Talavera and Diana Figueroa. My gratitude goes also to José Luis Arce and José Luis Farfán for their assistance and collaboration during this work.

Finally, I thank my family Lourdes, Olinka and León Felipe for their loving support; this dissertation is dedicated to them.

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ABSTRACT

More than 80 % of the exposed area of the Acatlán Complex was mapped and new outcrops of eclogites (Xayacatlán Formation), granitoids (Esperanza, Hornos, Teticic), and volcanic rocks (Cosoltepec and Tecomate Formations) were identified. 44 selected samples from the new outcrops were geochemically analyzed to establish the tectono-magmatic affinity. Volcanic rocks of the Petlalcingo Group suggest an oceanic setting (MORB and OIB), while bimodal rocks of the Tecomate Formation represent an intraplate event (WPB), in addition to an inherited arc signature. Most granitoids of the Acatlán Complex are peraluminous and classify as S-type intrusions derived from crustal melting. These plutons straddle the VAG+ Syn-ColG and WPG fields in a Nb-Y discriminant diagram and are considered as syn-tectonic, late-tectonic and post-tectonic granites.

Based on the main geologic, stratigraphic, structural and geochemical data it is suggested that the Petlalcingo Group represents a sedimentary sequence deposited in distal passive margin correlated with those siliciclastic deposits of the Early Paleozoic bordering the western margin of Gondwana. The Piaxtla Group (Xayacatlán and Esperanza Granitoids) represents an allochthonous unit deformed, metamorphosed and emplaced over the Petlalcingo Group during the Silurian orogeny (Caledonian) caused by the collision of peri-Gondwanan terranes (Avalon, Oaxaquia) and the subsequent approaching of Gondwana. During the Early Devonian, the Tecomate Formation was deposited because of the relaxation of the Silurian orogen. Finally the subsequent southward translation of Gondwana produced an intense deformation and low grade metamorphism into the Tecomate Formation (Acadian orogeny). Upper Paleozoic sediments deposited over the Acatlán Complex delimit its evolution.

CHAPTER I. INTRODUCTION

1. 1 Statement of Problem

Late Paleozoic-early Mesozoic paleogeographic reconstructions of Pangea show different extents of overlap between the South American continent and the Mexican territory (Bullard et al., 1965; Van der Voo et al., 1976). According to these paleo-reconstructions, the present position of southern Mexico is considered to be allochthonous (Figure 1. 1). To avoid such overlapping, major crustal left lateral faults have been proposed to relocate parts of Mexico during its Mesozoic evolution. Most of these models, which are based on geologic reconstructions, geometric arrangements and paleomagnetic data, suggest the displacement of south-western North America (De Cserna, 1971; Pindell and Dewey, 1982; Anderson and Smith, 1983; Urrutia-Fucugauchi, 1984) (Figure 1. 2).

It is now widely accepted that during the break-up of Pangea, the rifting process displaced 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 subsequently constituted the metamorphic basements in the eastern and southern regions of Mexico and Central America during Mesozoic time.

According to those arguments based on reconstructions of orogenic systems, the Mixteco and Oaxaca terranes of southern Mexico (Campa and Coney, 1983) would be two of the most important Paleozoic and Precambrian outcrops that have been identified as pieces of the Appalachian-Caledonian and Grenville systems respectively (De Cserna, 1971; Ortega-Gutiérrez, 1981; Patchett and Ruiz, 1987; Ruiz et al., 1988; Yáñez et al., 1991). However, their precise locations during the Paleozoic evolution remain uncertain, being a topic of strong debate.

Early interpretations considered the Acatlán and Oaxaca Complexes (metamorphic a)

b)

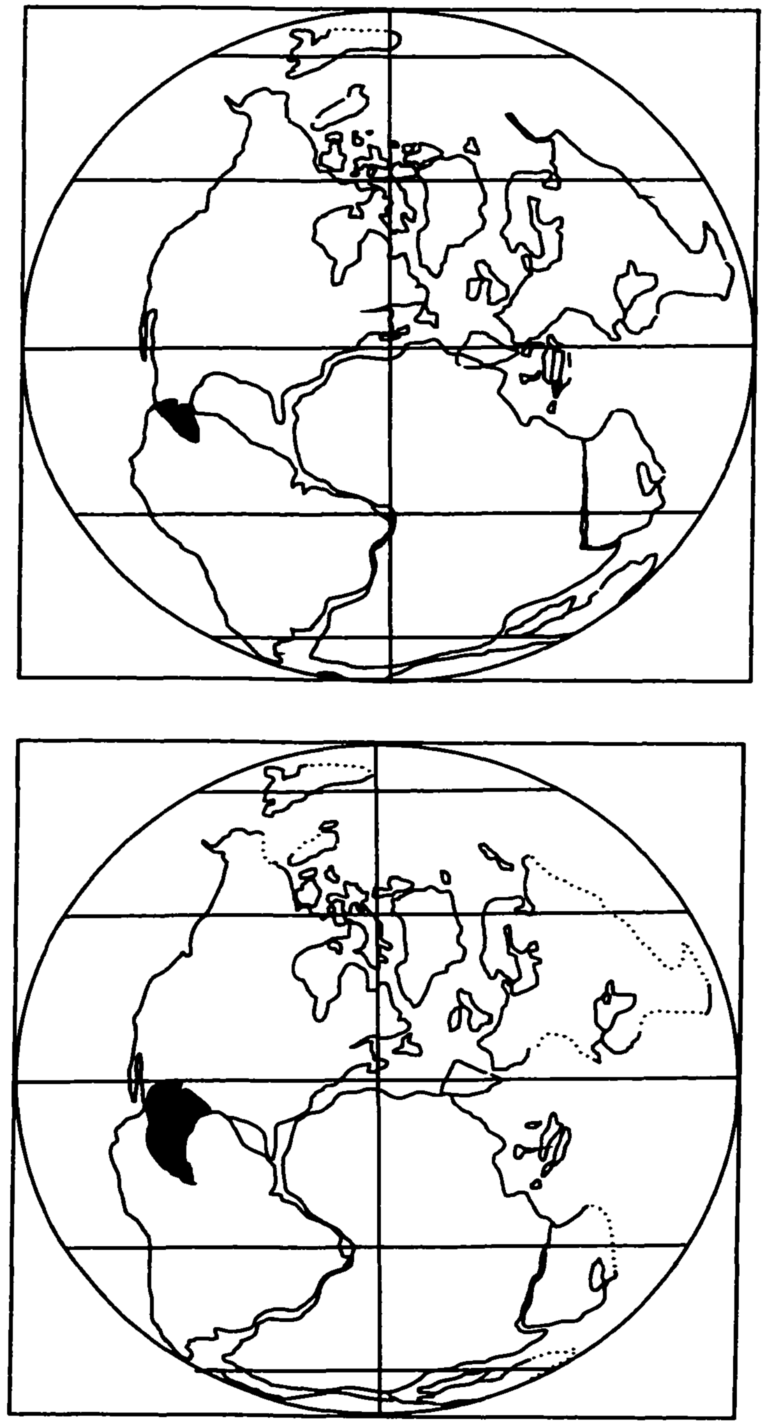


Figure 1.1: Late Paleozoic-Early Mesozoic reconstructions showing different extens of overlapping (shown in black) between the South American continent and México. A) Bullard et al., 1965; b) Van der Voo et al., 1976.

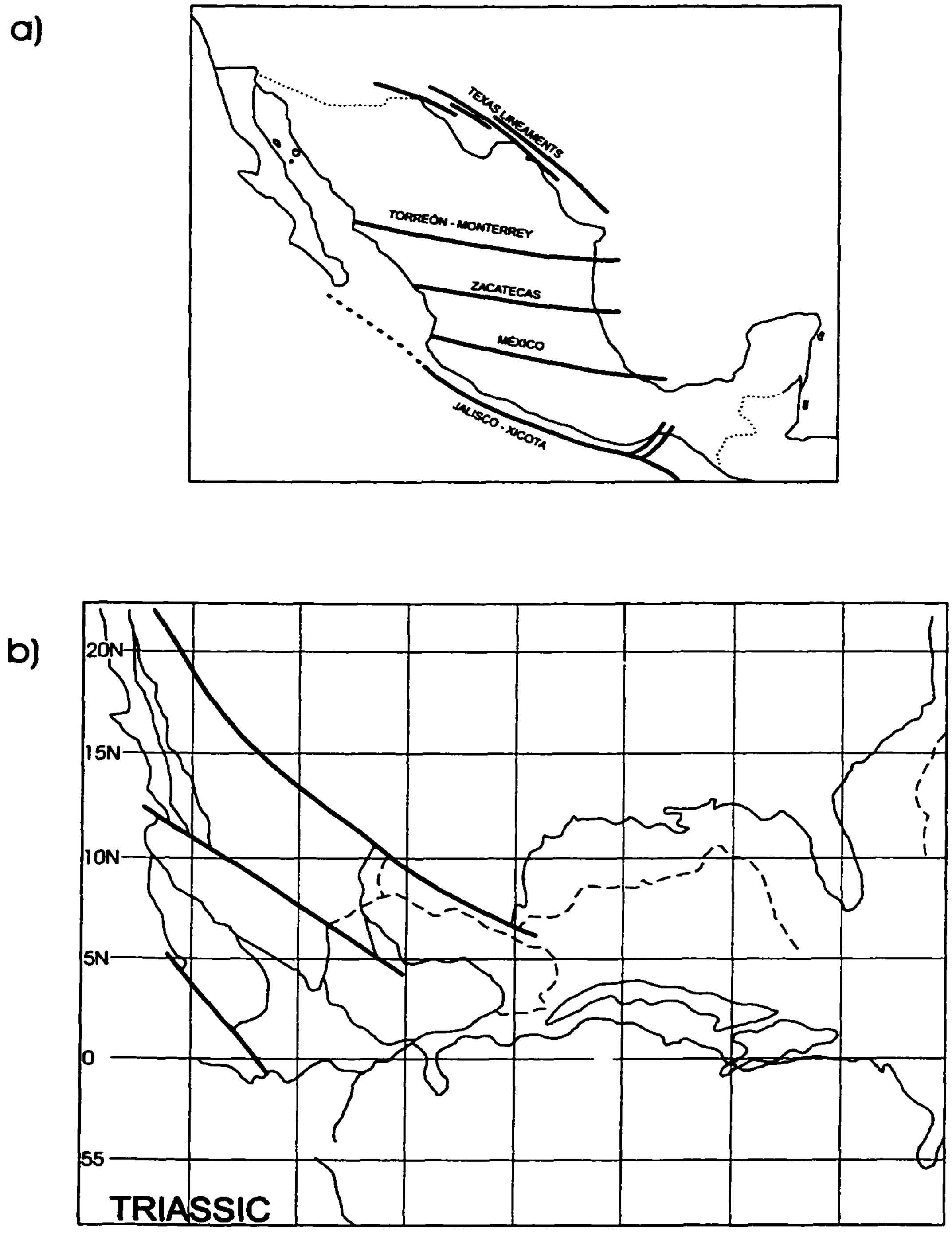


Figure 1.2: Some of the proposed strike-slip faults to relocate Mexico during the break up of Pangea. a) de 'Cerna,1971; b) Anderson and Schmidt, 1983.

basements of the Mixteco and Oaxaca terranes respectively) to be the southward continuation of southeastern North America (Ortega-Gutiérrez, 1981; Ruiz et al., 1988). However, because the configuration of this system in southern Mexico is the opposite to that system in North America (e.g., the Grenvillian basement of Oaxaca to the east, and the Appalachian basement of Acatlán to the west), both terranes were better linked to the evolution of Colombia in northern South America (Van der Voo, 1988; Yáñez et al., 1991). Even more, the presence of sediments with Tremadocian fauna of Gondwana affinity overlying the Grenville basement in Oaxaca strongly supported that correlation (Robinson and Pantoja-Alor, 1968; Pantoja-Alor, 1970; Whittington and Hughes, 1974). This evidence clearly suggests a very complex tectonic-framework where the process of terrane-translation played an important role during the Laurentia-Gondwana interactions in Paleozoic time and the subsequent rifting of Pangea through the Mesozoic (Figure 1. 3).

More recent models suggest earlier collisions between Laurentia and Gondwana, the consequent transference of some terranes during this time, the possible existence of the Oaxaquia microcontinent and the evolution of the Iapetus and Rheic oceans during the Paleozoic. These models make more complex the reconstruction of the Appalachian system and therefore of the Acatlán Complex history (Dalla-Salda et al., 1992; Dalziel, 1995; Ortega-Gutiérrez et al., 1995; Keppie and Ramos, 1999).

During the last two decades, most studies carried out in the Acatlán and Oaxaca Complexes have been focused on establishing the possible connections of these two terranes to the main continental masses. Such studies have included: paleomagnetism (Gose and Sanchez-Barreda, 1981; Mc Cabe et al., 1988; Ballard et al., 1989; Fang et al., 1989), paleontological correlations (Robinson and Pantoja-Alor, 1968, Whittington and Hughes, 1974), Nd and Pb isotopic analyses (Patchett and Ruíz, 1987; Ruiz et al., 1988; Keppie and Ortega-Gutiérrez, 1995; Ruiz et al., 1999), and geochronology in order to date

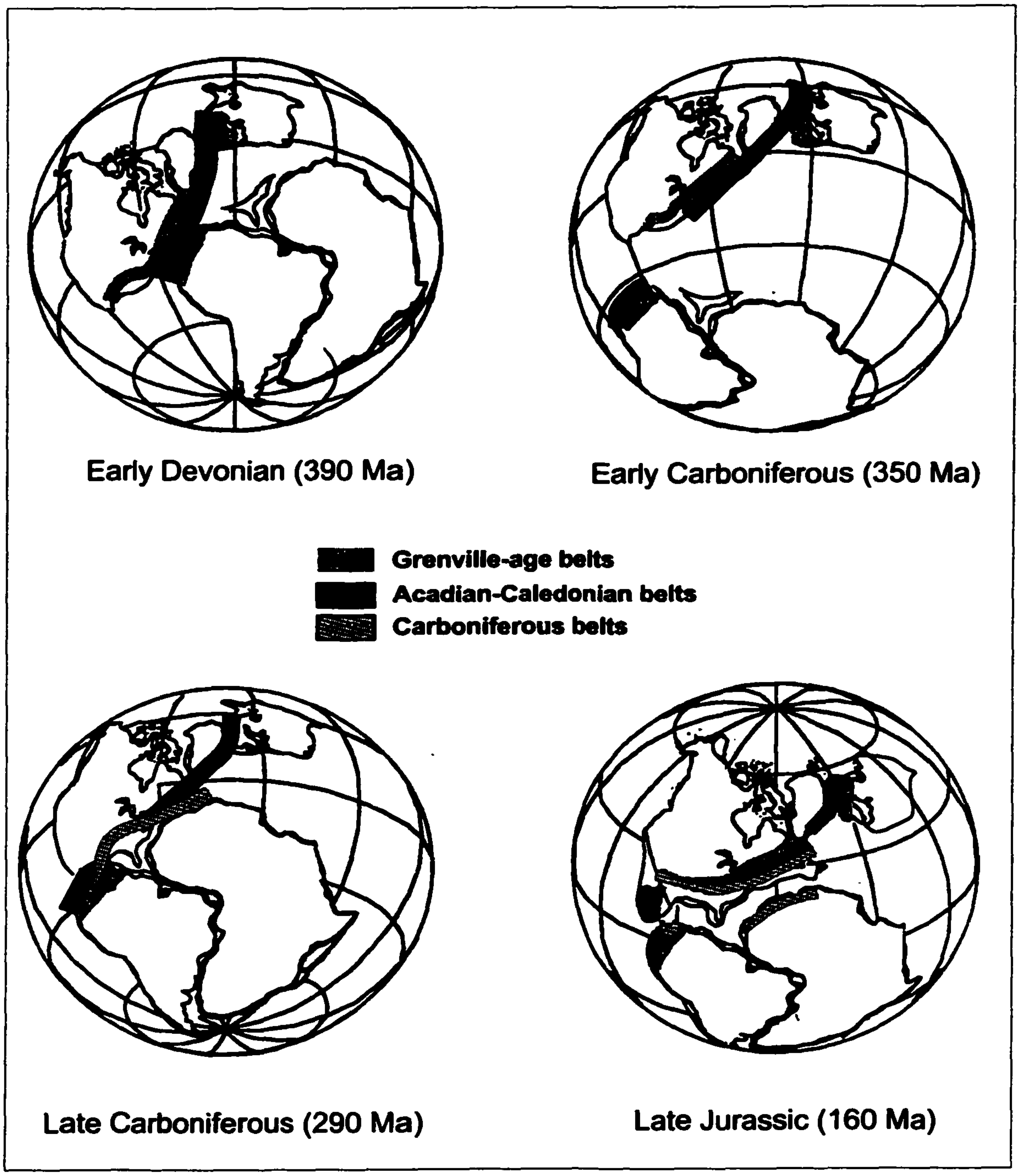


Figure 1. 3: Suggested interactions between Laurentia and Gondwana continents and transference of some terranes through the Paleozoic and Mesozoic times. (From Ruiz et al., 1988; Yañez et al., 1991).

metamorphic and magmatic events (Halpern et al., 1974; Ruíz-Castellanos, 1987; Robinson, 1990; Yáñez et al., 1991; Ortega-Gutiérrez et al., 1999). Most results suggest that the Oaxaca Complex was part of the Grenville system located in between the Laurentian craton and the Amazonian craton before the break up of Rodinia (around the Adirondacks region in northeastern Canada and the Colombian Andes) (Ballard et al., 1989; Keppie and Ortega-Gutiérrez, 1995; Ruíz et al., 1999).

In constrast, analytical data from the Acatlán Complex, which is the objective of the present dissertation, have been inconclusive and some times controversial. Most paleomagnetic studies have not been successful (Fang et al., 1989), and the recorded U/Pb and Sm/Nd isotopic ages defined three main tectono-magmatic and metamorphic events during theSilurian, Devonian and Carboniferous times, such events representing the Caledonian, Acadian and Alleghenian orogenies respectively. On the other hand, the most important Paleozoic events in Colombia were registered in Ordovician and Silurian times during the Taconic and Caledonian orogenies, but no late Paleozoic ages have been recorded yet (Forero-Suárez, 1990; Yáñez et al., 1991; Restrepo-Pace et al., 1997). Therefore, the data obtained preclude direct correlation between the Acatlán Complex and similar metamorphic rocks of northern South America.

The more recently U/Pb analysis conducted on granitic rocks of the Acatlán Complex yielding Late Ordovician-Silurian ages (Robinson, 1990; Ortega-Gutiérrez et al., 1999) are similar to those ages widely reported in the northern Appalachian system as a consequence of the Silurian orogeny. This Silurian orogeny, called Salinian or early Acadian by Dunning et al. (1990); and Hibbard (1993) respectively, was the result of the final collision of the Avalonian microplates against Laurentia and is correlative to the Caledonian orogeny in Europe. According to this data, the possible correlation of the Acatlán Complex to equivalent rocks located in present northeastern North America could be re-established, if it is accepted that much of the eastern North America is made up of peri-Gondwanan terranes transferred during Appalachian orogenesis. However, it is evident that many problems still exist to define the overall evolution of the Acatlán Complex to establish more precise continental correlations.

In addition to the studies indicated above, the Acatlán Complex has been more recently the target of new research attempting to define the tectonic affinity and significance of the different metamorphic lithologies included in this complex. Consequently, whole-rock geochemistry and thermobarometry have been conducted on eclogites and metasediments of the Xayacatlán Formation (Ortega-Gutiérrez and Reyes-Sala, 1997; Meza-Figueroa, 1998). Sedimentologic and provenance studies of the Tecomate Formation are in progress (Sánchez-Zavala and Ortega-Gutiérrez, 1997). New and more reliable U/Pb isotopic ages have been obtained from granitic rocks (Ortega-Gutiérrez et al., 1999).

Finally, as a contribution to understanding and to characterization of the tectonic affinity and evolution of some units of the Acatlán Complex, the present work is focused on conducting whole-rock geochemistry and Nd isotopes in most magmatic units.

1. 2. Purpose of this study

One of the main objectives of the present study was focused on the geologic mapping of the whole outcropping Acatlán Complex at a scale of 1: 250 000. Surprisingly, 25 years after the pioneering study of this complex (Ortega-Gutiérrez, 1975), only 20% of the exposed area was internally mapped. Most studies carried out in the Acatlán Complex during the last 20 years (isotopic analyses, paleomagnetism, geochronology, thermo-barometry, etc) were conducted in the same mapped area, which is the most accessible and nearest to Mexico City (Figure 1. 4).

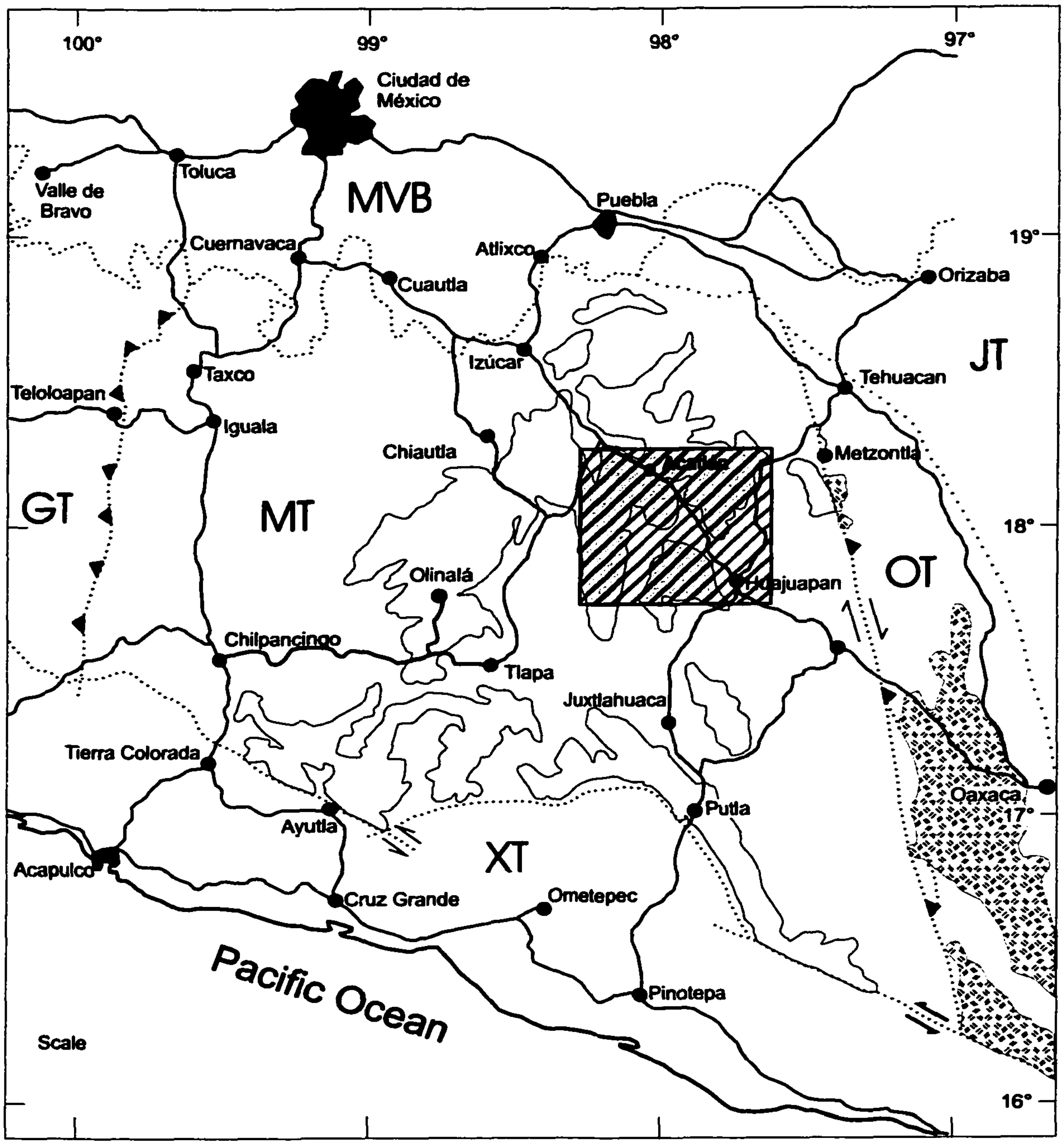


Figure 1. 4: Outcrops distribution of the Acatlán Complex (basement of the Mixteco Terrane). Boxed area studied by Ortega-Gutiérrez, (1975). MT: Mixteco Terrane, GT: Guerrero Terrane, XT: Xolapa Terrane, OT: Oaxaca Terrane, JT: Juárez Terrane, MVB: Mexican Volcanic Belt.

The mapping of the Acatlán Complex was covered up to 85% of the total area; unfortunately the remaining 15% is located in the most risky and uninhabitable region of southern Mexico. Nevertheless, the general distribution of the units included in the Acatlán Complex were satisfactorily outlined (Appendix). In addition to the large-scale mapping, two smaller areas located in the vicinity of the Izucar -Tehuitzingo and Olinalá regions were also mapped in detail (scale 1: 50 000). These two small areas were selected to establish the general geologic framework between formations of the Acatlán Complex, and to control the sampling of the major magmatic units to conduct systematic whole-rock geochemistry. Figure 1. 4 shows the regional distribution where the Acatlán Complex crops out, the area mapped by the author, as well as the area mapped by Ortega-Gutiérrez (1975).

Despite the occurrence of magmatic units (volcanic and plutonic rocks) through the whole stratigraphy of the Acatlán Complex, systematic geochemistry has not been previously done. Only Ortega-Gutierrez (1975) and Fries et al. (1965) had geochemically analyzed some metabasites and some granites of the Xayacatlán Formation and Esperanza granitoids respectivelly, using major elements.

The presence of magmatic rocks in most units of the Acatlán Complex is important to characterize their tectonic setting and therefore a fortunate feature to better define the evolution of this complex. This work presents results of whole-rock geochemistry (major, trace and rare earth elements), and Nd isotopic analyses conducted in meta-volcanic rocks of the Cosoltepec Formation (Petlalcingo group), meta-volcanic rocks of the Tecomate Formation, in addition to the Esperanza granitoids, Los Hornos- Noria granites and the Teticic type stocks.

The new data will be used as a parameter of correlation with the main magmatic events recorded in the evolution of the Appalachian-Caledonian system, and will also provide additional constraints on some of the proposed paleogeographic reconstructions and connections between this terrane and the main continental masses during Paleozoic time.

1. 3. Dissertation Format

The following manuscript, which constitutes the scope of this dissertation, is presented in four chapters. This chapter describes the state of the art about the knowledge of the Acatlán Complex evolution, and the suggested connection to the main continental plates involved in the origin of the Appalachian System. It also defines the objectives of the present work as well as the significance of this study to understand the evolution of the Acatlán Complex.

Chapter two presents the geological framework of the Acatlán Complex, including maps, stratigraphy, geologic sections and structural data gathered during the field work carried out in this study and elsewhere during the last twenty years.

Chapter three presents data of the whole-rock geochemistry (major, trace, rare earth elements, and isotopes) from 44 analyzed samples. This information displayed in tables, rock-classification diagrams, rare-earth and multi-element patterns, as well as diagrams to discriminate between tectonic environments. The geochemical results along with the geological information are together analyzed to define the tectonic setting for each considered unit.

Chapter four presents the possible correlations of the analyzed samples with the main magmatic events registered in the Appalachian System, as well as the possible connection and evolution of this complex during its Paleozoic evolution, from Rodinia to Pangea. Additionally, this chapter also presents the conclusion of this study and suggestions of work to be done in further studies.

CHAPTER II. GEOLOGIC SETTING

2. 1. Previous studies and Definition

Since the first half of the last century, there have been numerous descriptions of the metamorphic rocks outcropping in the Acatlán region, however, the first attempts to establish the geologic framework of these metamorphic rocks were made by Rodríguez-Torres (1969, 1970), Ruiz-Castellanos (1969, 1970) and Ortega-Gutiérrez, (1970). Previously, Fries and Rincón-Orta (1965) and Fries et al. (1970) had conducted isotopic studies trying to date some units of the "Acatlán Formation".

It is indisputable that the basic stratigraphy, petrology and structural geology of the Acatlán Complex were established in a pioneer study by Ortega-Gutiérrez (1975). This important work is still the starting point to analyze the nature and structure of the Acatlán Complex. Since then, the most relevant contributions about this complex have been done by the same author by means of mainly petrologic and isotopic studies, as well as tectonic interpretations (Ortega-Gutiérrez, 1978, 1981a, 1981b, 1991,1993; Ortega-Gutiérrez and Reyes-Salas, 1997; Ortega-Gutiérrez et al., 1997, 1999). Some other important contributions have been made in different times by Ruiz-Castellanos (1979), Robinson (1990), Yáñez et al. (1991), Keppie and Ortega-Gutiérrez (1995), Sánchez-Zavala (1997) and Meza-Figueroa (1998). It is also important to stress the fact that the mapping produced by Ortega-Gutiérrez (1975) was the base to complete any subsequent study (Figure 2. 1).

Initially, the Acatlán Complex was defined by Ortega-Gutiérrez (1975) as a polymetamorphic complex of early Paleozoic age generated in a probable subduction zone comparable with those present in Alpine-type orogens. At the same time, the Acatlán Complex was considered to be the possible continuation of the Paleozoic Caledonian-

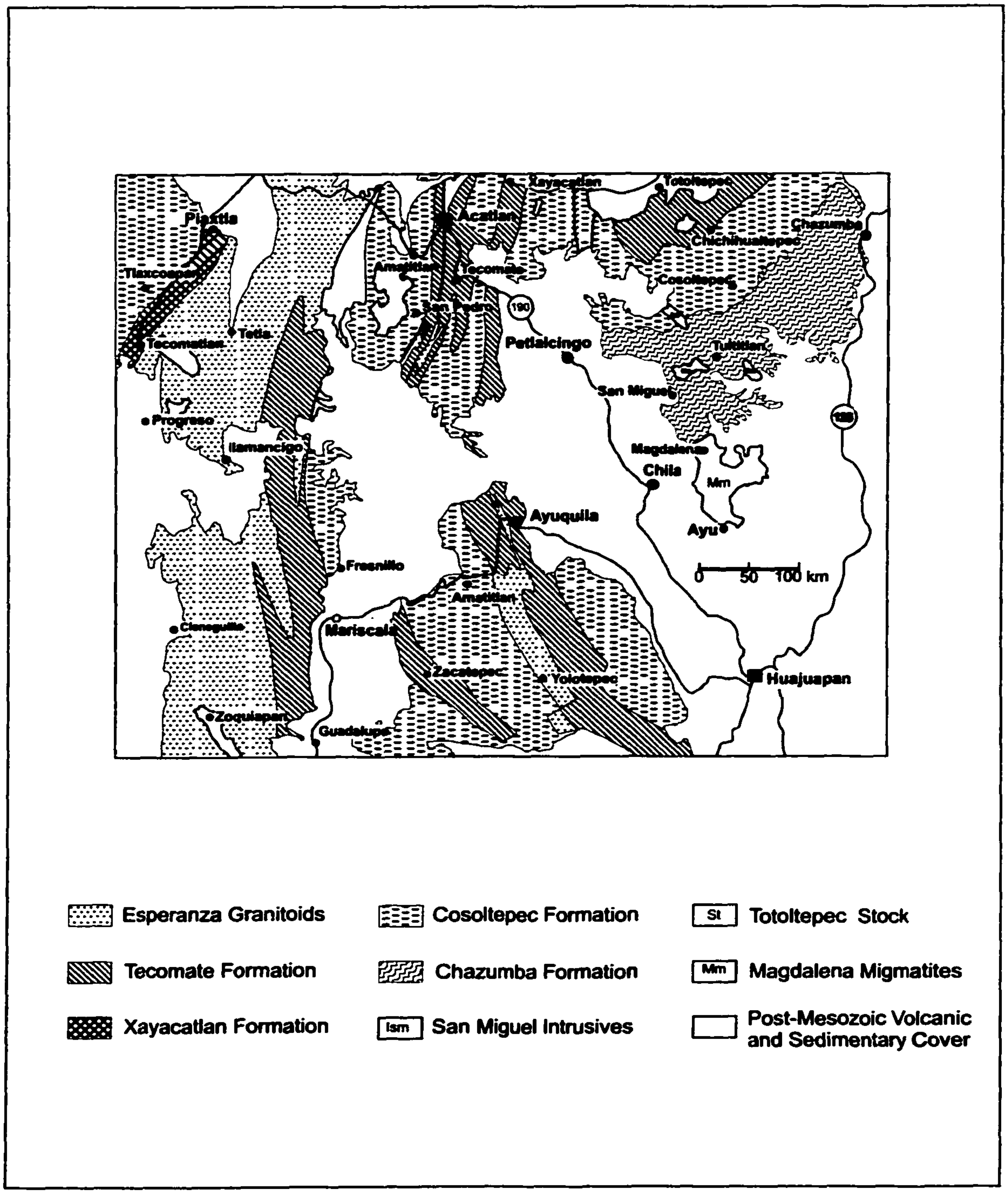


Figure 2. 1: Simplified geologic map of the Acatlán Complex from the area worked by Ortega-Gutiérrez (1975).

Appalachian orogenic system into Mexico.

At present, the Acatlán Complex is defined as a polymetamorphic complex of early to middle Paleozoic age. Lithologically represents the opening and subsequent closure of either the Iapetus or Rheic Oceans carried out by continent-continent collision between Laurentia and Gondwanaduring the middle Paleozoic orogenies (Ortega-Gutiérrez et al., 1993, 1997, 1999; Keppie and Ramos, 1999). Nevertheless, many uncertainties and gaps still exist to reconstruct the whole evolution of this metamorphic complex.

2. 2. Location

The metamorphic Acatlán Complex, located in the Sierra Madre del Sur, 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) (Figure 2. 2a).

To the west, the Guerrero terrane is tectonically juxtaposed with the Mixteco terrane, however, the location and nature of this limit is controversial. According to Campa and Ramírez (1978) and Campa and Coney (1983), the contact is represented by a large thrust fault trending north-south and dipping west located near Teloloapan, Guerrero, and Ixtapan de la Sal, state of Mexico. Throughout this fault the Mesozoic submarine volcanic rocks of the Guerrero terrane tectonically overlie the western border of the Mesozoic sedimentary cover of the Mixteco terrane, locally called Morelos-Guerrero Platform (Figure 2. 2a shows this reverse fault close to the Teloloapan town as reference). On the other hand, Sedlock et al. (1993), and Ortega-Gutiérrez et al. (1999) consider the true boundary between the Guerrero (Nahuatl terrane of Sedlock) and Mixteco terranes along the Papalutla fault. According to them, the Acatlán Complex

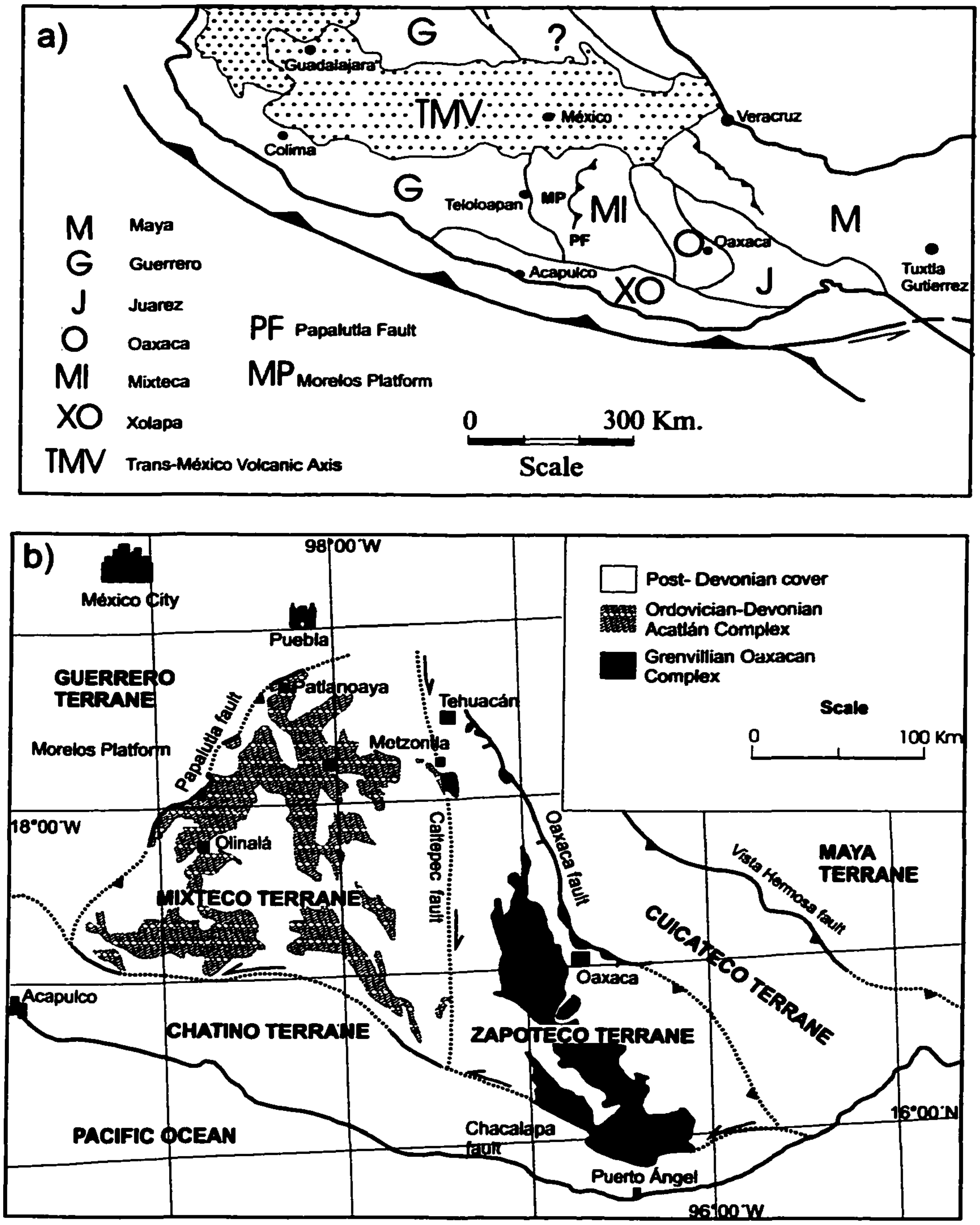


Figure 2.2: Simplified maps of terranes of southern Mexico showing different limits for the Mixteco terrane. a) The western limit includes the Morelos Platform (Campa and Coney, 1983); b) the Western limit corresponds to the Papalutla fault excluding the Morelos platform (Ortega-Gutiérrez et al., 1999)

overrides the Mesozoic sedimentary rocks of the Morelos-Guerrero Platform to the east along this fault (Figure 2. 2b). Basically, the underlying issue of this dispute is about the basement of the Mesozoic platform.

Field work observations and mapping carried out around the Acahuizotla village and along the Chapolapa river (southwestern outcrops of the Acatlán Complex), demonstrate that metamorphic rocks of the Acatlán Complex (Cosoltepec Formation) are overlain by Mesozoic rocks of the Morelos-Guerrero Platform (Appendix 2). This observation strongly supports the boundary suggested by Campa and Ramírez (1978) and Campa and Coney (1983). It is also possible that below the Mesozoic rocks located west of the Chilpancingo-Ocotito highway and around the Ixcuinatoyac region, the Acatlán Complex is present. However, more detailed studies are needed to resolve this question.

To the east, the Mixteco and Oaxaca terranes are juxtaposed by a sub- vertical right-lateral fault trending NNW and dipping ENE. Throughout this fault, the basement of the former terrane (Acatlán Complex) is over thrust by the Grenvillian basement (Oaxaca Complex) of the latter terrane. This relationship is well exposed in Los Reyes Metzontla and Caltepec region, near Tehuacán, Puebla (Ortega-Gutiérrez, 1981; Elías-Herrera and Ortega-Gutiérrez, 1998) (Appendix 2). According to Elías-Herrera and Ortega-Gutiérrez (1998), this fault (Caltepec fault) displays a very complex history since Devonian time, when both terranes were placed in contact by the dextral strike-slip fault in a transpressional regime. Such relationship is attributed to the lateral motion between Laurentia and Gondwana during Devonian time (Van der Voo, 1988; Dalziel, 1995). Following the ideas of Elías-Herrera and Ortega-Gutiérrez, such motion has been recorded by the existence of syntectonic plutons intruding the contact, like the Cozahuico granite present in that region (previously called Caltepec granite by Ruiz Castellanos, 1979), which yielded a U/Pb zircon age of 373 ± 34 Ma. However, a Rb/Sr isochron of 269 ± Ma and a K/Ar age of 266 ± 13 Ma had previously been obtained for the same granite by Ruíz-Castellanos (1979) and Torres et al. (1999) respectively. A more recent and accurate U/Pb zircon dating has yielded a Permian age for this granite (Ortega-Gutiérrez, oral comunication), suggesting post-Permian rather than Devonian tectonic activity.

To the south, the Mixteco terrane is truncated by the west-northwest trending metamorphic structures of the Xolapa terrane (Figure 1. 4). Even though the contact between these terranes is obscured by the presence of Tertiary plutons stitching the contact, some outcrops define a sinistral strike-slip fault with an important vertical component, suggesting a transtensional regime that produced an impressive shear zone. The origin of this contact has been explained by the exhumation of the Xolapa terrane since late Eocene time during the translation of the Chortis block to the east (Robinson, 1990; Riller et al., 1991; Ratschbacher et al., 1991; Tolson, 1998). To the southeast and between the Mixteco, Oaxaca and Xolapa terranes, there exists a small terrane, called Juchatengo (Figure 2. 2b). This terrane is made up mostly of submarine basic igneous rocks interlayered with upper Paleozoic black shales, however its relationships with respect to the surrounded terranes are not sufficiently understood (Grajales-Nishimura, 1988; Grajales-Nishimura et al., 1999).

To the north, the Mixteco terrane is covered by the youngest overlapping sequence of Mexico, the Trans-Mexican Volcanic Belt, which is made up of Pliocene and Quaternary volcanic rocks (Figure 1. 4). The most northern outcrops of the Acatlán Complex are located in the Sierra de Tenzo, in the state of Puebla, where the metamorphic basement thrusts over its Mesozoic cover following the same structural trend with respect to the reverse Papautla fault (Appendix 2). This antitetic Papalutla thrust fault has been correlated to the late Mesozoic-early Tertiary folds and thrusts generated by the Laramide orogeny (De Cerna et al., 1980; Campa, 1985).

The outcrop area of the Acatlán Complex is mainly distributed in the central and eastern regions of the Mixteco terrane, as shown in Figure 1. 4. The studied area is included in five 250 000-scale topographic maps of Orizaba, Cuernavaca, Chilpancingo, Oaxaca and Acapulco. The extension of the mapped area is approximately of 30 000 km^2, however, some outcrops located in the southeastern corner of the regional map (Appendix 2), were excluded because of their distance, inaccessibility and risk. Nevertheless, the extension of the mapped area was strongly improved with respect to the previous work. By means of the accomplished mapping, it is now possible to visualize the general distribution of units included in the whole Acatlán Complex, therefore, this work opens the possibility to carry out more detailed studies in different regions. According to this, two smaller areas were additionally mapped in detail in order to establish the stratigraphic control of the sampled rocks used to conduct systematic geochemestry (Figures 2.3 and 2.4). These two areas are now the target of recent and in progress studies of the Acatlán Complex (Meza, 1999; Campa et al., 2001). These areas are included in two 50 000-scale topographic maps of Olinalá and Tehuitzingo, covering 1600 km^2 (Figures 2.3 and 2.4).

2. 3. Stratigraphy

The main stratigraphy, metamorphic petrology and structural characteristics of the Acatlán Complex were established by Ortega-Gutiérrez (1975). However, some names of the proposed formations were taken from Rodríguez-Torres (1970). Since then, the stratigraphy of this complex has been partly modified and its units have been regrouped according to new concepts and the available information (Ortega-Gutiérrez, 1993, 1999; Ramírez and Talavera, 1997).

Initially, the Acatlan Complex was subdivided into two groups called Petlalcingo and Acateco (Ortega-Gutiérrez, 1975, 1978). The former group includes the Magdalena,

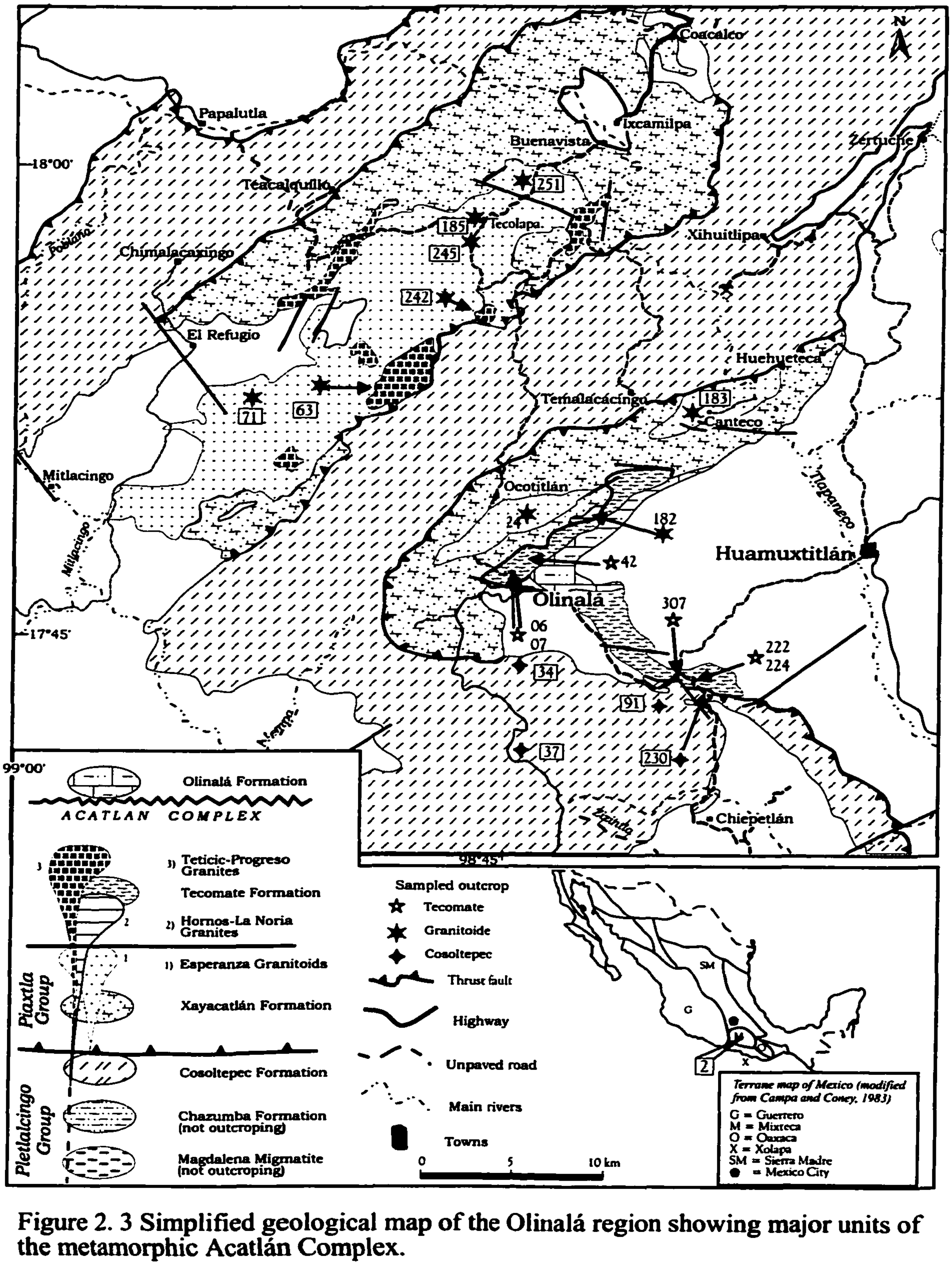


Figure 2. 3 Simplified geological map of the Olinalá region showing major units of the metamorphic Acatlán Complex.

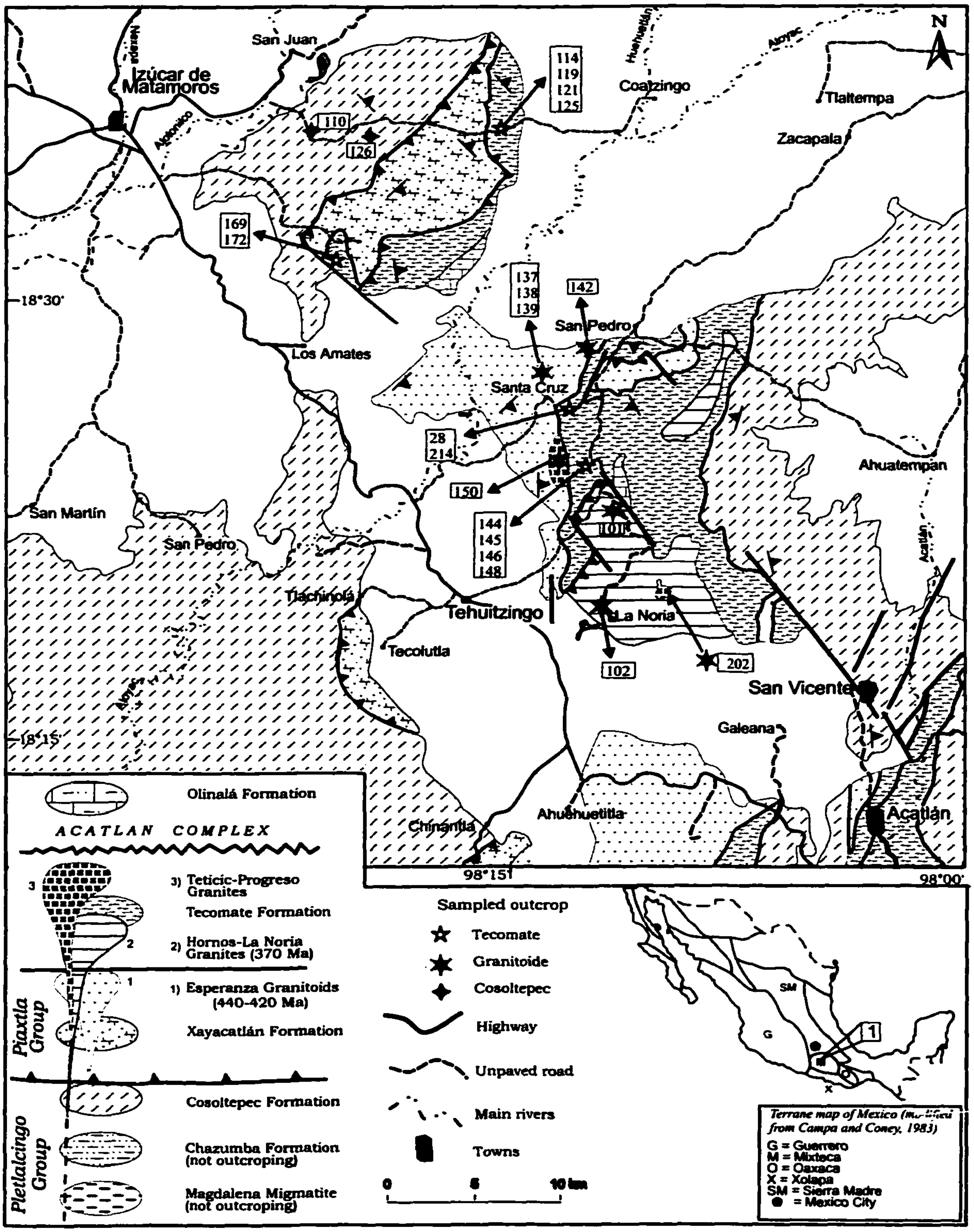


Figure 2. 4 Simplified geological map of the Tehuitzingo region showing majors units of the metamorphic Acatlán Complex.

Chazumba and Cosoltepec formations and it has not been modified through time, whereas the latter group which included the Xayacatlán and Tecomate formations were subsequently regrouped (Ortega-Gutiérrez, 1993). In addition to these groups, intrusive rocks considered to be younger than the other units were differentiated into Esperanza granitoids, Totoltepec pluton and San Miguel dikes (Figure 2. 5a). However, this subdivision, based on structural arguments, presented some confusing statements mainly in the definition and relationships of the Acateco Group and the Esperanza granitoids. At present, granitoids with similar deformation have yielded different ages and relationships have been restudied.

In 1991, Ortega-Gutiérrez suggested that the Xayacatlán Formation and the Esperanza Granitoids form part of an allochthonous assemblage that was affected by an eclogite facies metamorphism during Devonian time. In 1993, based on the new concept of the tectonostratigraphic terranes, Ortega-Gutiérrez presented a new stratigraphic column. Unfortunately, the new stratigraphy introduced confusing units like the Inopilco Formation, as well as the presence of remnants of Oaxacan gneisses (Grenvillian rocks) beneath the Tecomate Formation (Figure 2. 5b). However, the new concept of the litho-tectonic assemblages made more sense to the evolution of the Acatlán Complex.

Based on previous studies, new data, mapping and field observations carried out during the present study, the Acatlán Complex was regrouped into three lithotectonic assemblages (Ramírez and Talavera, 1997). The originally defined Petlalcingo Group, made up of the Magdalena, Chazumba and Cosoltepec formations (1); the newly defined Piaxtla Group, which includes the Xayacatlán Formation and the Esperanza Granitoids (2); and the post-tectonic Tecomate Formation (3). Additionally, two post-tectonic magmatic events recorded by small plutons are also reported (Figure 2. 6a). This column is similar to the last tectonostratigraphic column presented by Ortega-Gutiérrez et al.

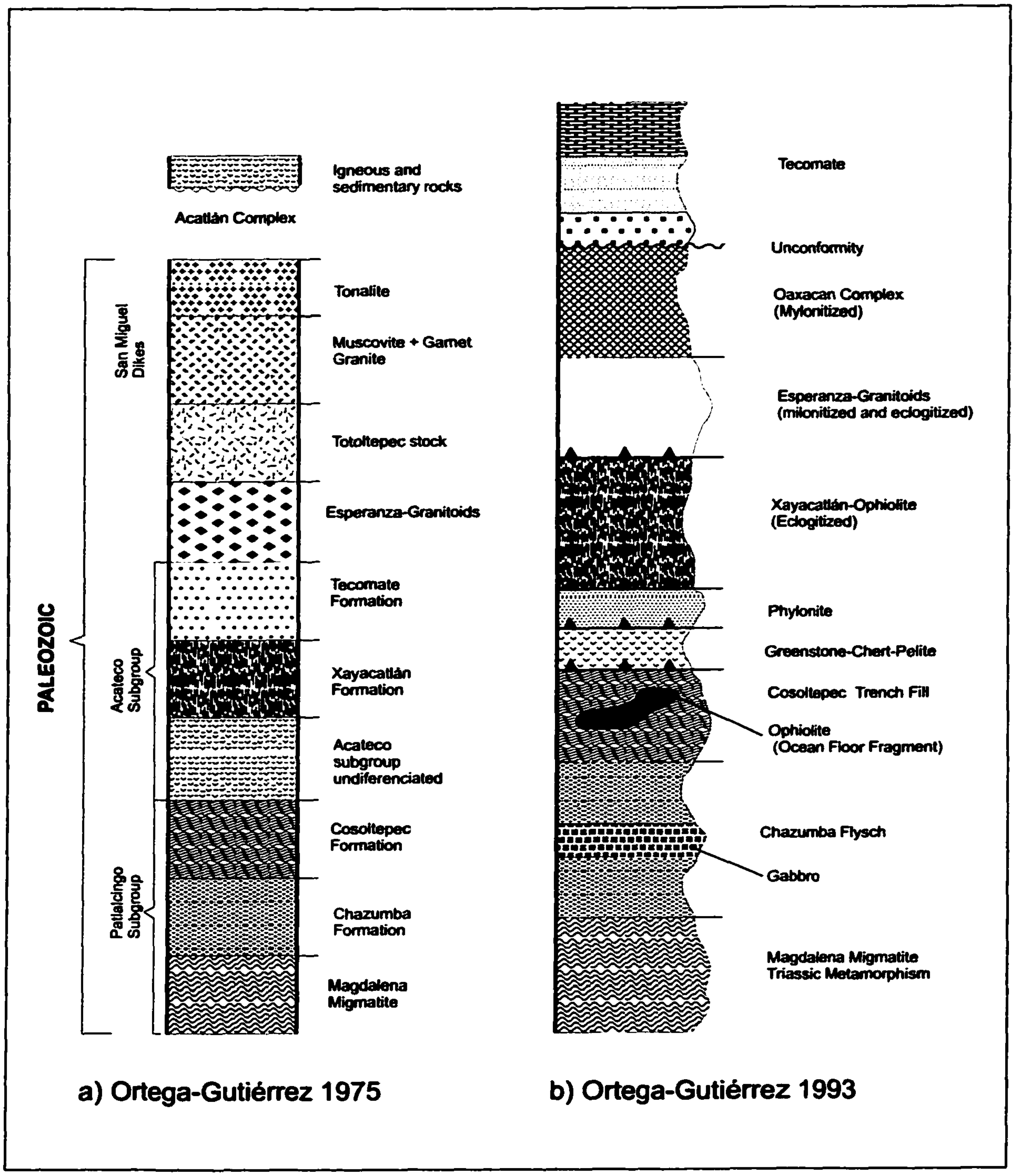


Figure 2. 5 Early stratigraphic columns of the Acatlán Complex suggested by Ortega-Gutiérrez. a) proposed in 1975, the Acateco group included the Xayacatlán and Tecomate formations. b) proposed in 1993, the Xayacatlán Formation and Esperanza granitoids group together.

(1999) (Figure 2. 6b). These new columns substitute the Acateco Group and reorder the Xayacatlán and Tecomate Formations along with different plutonic suites.

The following definitions and descriptions of units are based on previous works and observations gathered during the present study, however more detailed descriptions are found elsewhere in the cited references. Nevertheless, major emphasis will be given to those units geochemically analyzed in this dissertation.

Petlalcingo Group

The structural base of the Acatlán Complex is represented by the Petlalcingo Group, which was defined by Ortega-Gutiérrez (1975) as a deep marine sequence of graywacke, quartzite, shale and scarce carbonate, occasionally accompanied by basic volcanic rocks and intrusions. According to Ortega-Gutiérrez (1975), this sequence was affected by a prograde metamorphism developing a chlorite-biotite-garnet-staurolite-sillimanite zoning, therefore, ranging from greenschist to amphibolite facies. The base of this sequence is characterized by intense migmatization.

After mapping the whole Acatlán Complex, it was possible to realize that the whole Petlalcingo Group only crops out east of the Acatlán region, which is the type locality and nowhere else, is completely exposed. On the other hand, the Cosoltepec formation is the most widespread units of the whole Acatlán Complex (Appendix 2).

From the bottom to top, the Petlacingo Group has been subdivided into three units: (1) the Magdalena migmatites; (2) the overlying Chazumba Formation; and (3) the widespread Cosoltepec Formation. Even though the rocks have experienced intense deformation and prograde metamorphism, the relationships between these units seems to be continuous and transitional, however, Ortega-Gutiérrez (1975) suggests a possible

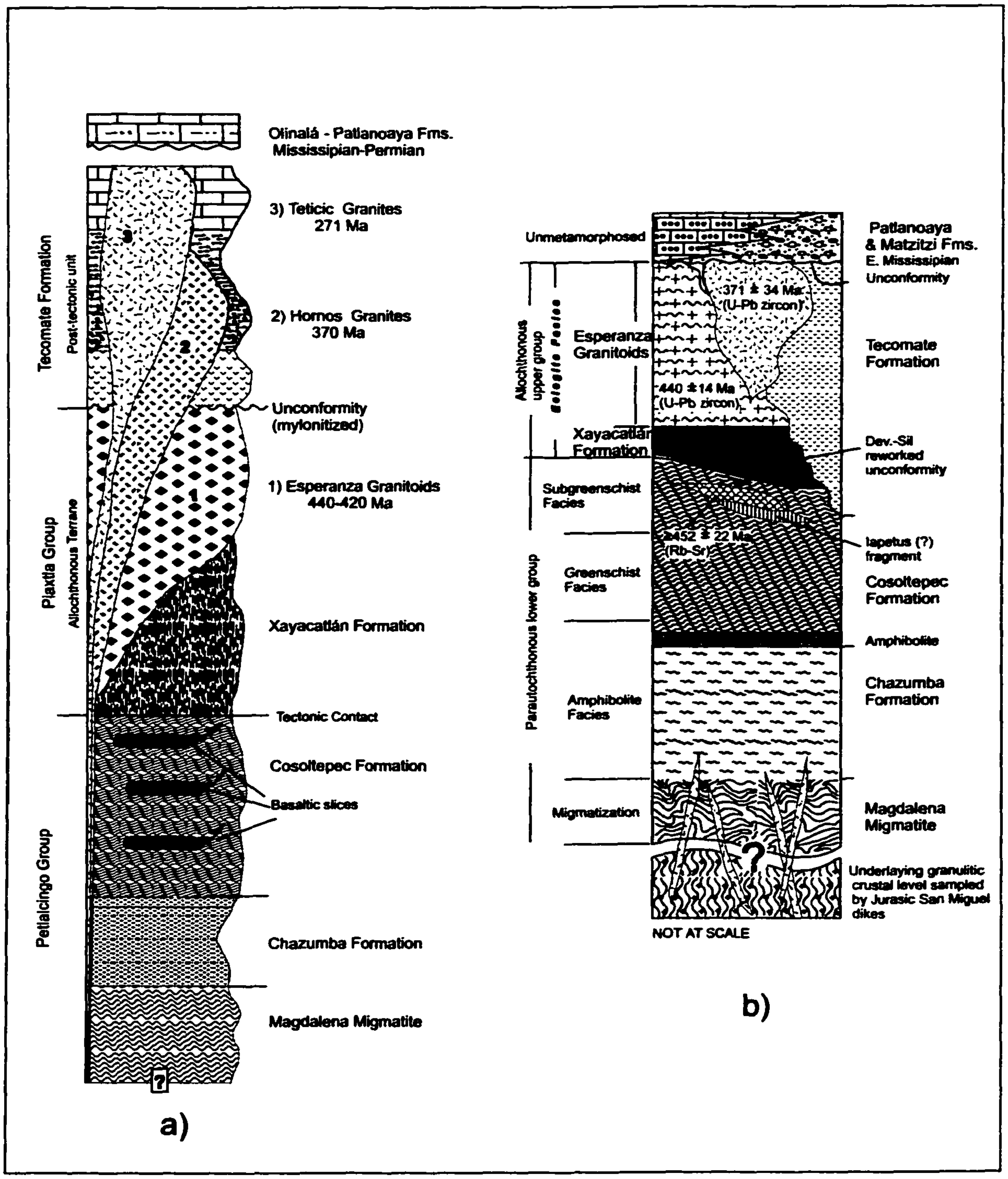


Figure 2. 6 Recent stratigraphic columns of the Acatlán Complex. a) proposed by this author (Ramírez, 1997),and b) proposed by Ortega-Gutiérrez et al.,1999.

unconformity or tectonic boundary between the Chazumba and Cosoltepec Formations. It is clear that the contact between those formations should be analyzed more carefully.

Magdalena Migmatite: According to Ortega-Gutiérrez (1975, 1978), this unit is defined as a classic migmatite which consists of alternating mainly granitic neosomes and biotite-rich paleosomes. This sequence also displays minor hornblende gneiss, amphibolite, rare calc-silicate and marble, gneissic granite, as well as biotite-schist and ultramafic inclusions. Migmatites are characterized by banded, folded and nebulitic structures, and fine-grained granoblastic textures consisting mainly of plagioclase, quartz, biotite and hornblende. This unit is interpreted to consist mostly of pelitic and psammitic sediments as the original protoliths along with scarce carbonates. On the other hand, there is no evidence of a lower unit or basement beneath the Magdalena migmatite and the relationship with the overlaying Chazumba Formation is obliterated by the migmatization event and by the intrusion of post tectonic granites.

Even though pervasively migmatized, the Magdalena migmatite has been structurally analyzed suggesting a complex history of deformations. In some cases up to seven events of deformations have been suggested (Powell et al., 1998), however, more realistic data suggest only four deformational events (Ortega-Gutiérrez, 1975).

On the other hand, the age of the protolith of this unit has not been determined, but the process of migmatization yielded Sm/Nd garnet-whole rock and Rb/Sr muscovite-whole rock ages of 204 ± 6 and 163 ± 2 Ma, respectively (Yáñez et al., 1991). More recent study based on U/Pb analysis of single zircon grains yielded a lower intercept age of 170 Ma, and an upper intercept age of 1053 Ma (Powell et al., 1998). These dates clearly indicate that the migmatization process occurred during the middle Mesozoic (Early to Late Jurassic) possibly related to the break up of Pangea in Mexico or to the genesis of a core complex structure (Powell et al., 1998). In addition, granitoids of the

Magdalena migmatite display crustal residence ages (T_{DM}) ranging from 1.32 to 1.87 Ga, and 0.7 to 0.8 Ga for amphibolites, suggesting recycled crust and addition of juvenile materials respectively (Yáñez et al., 1991). Only one metasedimentary sample from this unit has been geochemicaly analyzed but using major elements (Ortega-Gutiérrez, 1975).

Chazumba Formation: This formation is made up mostly of biotite-rich psammitic and pelitic schist, minor quartzite and scarce mafic and ultramafic rocks. Biotite-rich schists show banded, crenulated and very diverse folded structures along with fine-grained granoblastic to lepidoblastic textures consisting mainly of quartz, plagioclase and biotite, as well as muscovite and garnet as main accessories.

Mafic rocks consist of gabbroic bodies of either massive or apparently stratified structures included in the schists. Rocks of the Chazumba Formation clearly represent a marine sedimentary sequence highly deformed and affected by an amphibolite facies metamorphism (Ortega-Gutiérrez, 1975). The upper contact with the overlaying Cosoltepec Formation seems to be concordant with similar northward dipping foliation. However, the observation made by Ortega-Gutiérrez (1975) about a possible unconformity or tectonic contact should be taken into account.

The only available isotopic ages of the Chazumba Formation were obtained by a Sm/Nd single garnet-whole-rock pair, and Rb/Sr muscovite-whole rock pair isochrons yielding ages of 429 Ma and 349 Ma respectively (Yáñez et al., 1991). These dates with a wide variation, therefore, seem to represent cooling ages. A more consistent 1.4 Ga crustal residence age was also obtained from this unit (Yáñez et al., 1991). Additionally, geochemical analysis using major elements were conducted on ultramafic rock from this unit (Ortega-Gutiérrez, 1975).

Cosoltepec Formation: This formation is the most widespread and thick unit of the Acatlán Complex. It represents almost 70% of the outcrops of this complex and it is distributed all over the central and eastern Mixteco terrane. Those terranes adjacent to the Acatlán Complex, basement of the Mixteco terrane are in tectonic contact with the Cosoltepec Formation. To the east, the Oaxacan Complex overlies this formation, whereas to the southwest it is tectonically juxtaposed with rocks of the Xolapa terrane. To the west, the Cosoltepec Formation is exposed along the Papalutla fault and Tenzo region where it is thrust over the Mesozoic cover of the Morelos-Guerrero Platform (Appendix 2). However, the most extensive outcrops of this formation are located south and northwest of the exposed Acatlán Complex. It is also observed in the cores of the tight anticlines along the Petlalcingo-Acatlán-Totoltepec region underlying the Piaxtla Group and Tecomate Formation.

The structural base of the Cosoltepec Formation is only exposed near its type locality where it is in contact with the underlying Chazumba Formation. A similar relationship could be found along the southeastern outcrops of the Acatlán Complex, between the towns of Juxtlahuaca and Putla, where the Cosoltepec Formation is in contact with biotite-rich schists which can be correlated with the Chazumba Formation. However, because of the Xolapa terrane is also represented by biotite-schists, this correlation is ambiguous. The Cosoltepec Formation is either tectonically overthrust by the allochthonous Piaxtla Group or unconformably overlain by the post tectonic Tecomate Formation. On the other hand, south of Huajuapan de León town, the Cosoltepec Formation is thrust over the Tecomate unit following a NW -trending and E-dipping reverse fault (Appendix 2). This formation can be geomorphologically distinguished by its higher elevation and lighter color, possibly as a consequence of the abundant presence of quartzite.

The Cosoltepec Formation is made up mostly of black slate, phyllite and fine-grained quartzite (Figure 2. 7). These lithologies are preferentially associated and geographically distributed. Black slates and phyllite are mainly associated and distributed toward the southern and southwestern outcrops (Tlacoapa-Malinaltepec and Ayutla-Chapolapa-Colotlipa regions), as well as in the Huajuapan region. Quartzite and phyllite are mainly distributed toward the central and northern outcrops (Olinalá-Papalutla and Tecomatlan-Acatlán-Izucar de Matamoros regions), however, such lithologies can alternate locally.

Most sedimentary structures of the Cosoltepec Formation have been obliterated by the strong deformation and greenschist facies metamorphism. Nevertheless the present deformation, lamination, gradation and slump structures can be recognized within the pelitic and psammitic sediments. Graded bedding is common but the rippled part of the Bouma sequence is not seen. Nevertheles, this sequence is best interpreted as a turbiditic deposit. Quartzite beds vary in thickness from thin (1-2 cm) to massive beds (3-4 m). Noteworthy is the absence of limestone, only nearby Chila village along the road to Chiautla town, there is a small outcrop made up of lenses and thin beds of strongly recrystalized limestone. Ortega-Gutiérrez (1975) also reports an outcrop of limestone 10-km NE of Petlalcingo. According to Ortega-Gutierrez (1975), near its type area, the Cosoltepec Formation displays a prograde metamorphism developing the chlorite-biotite-garnet metamorphic zones, whereas in any other place, it is only characterized by the development of chlorite. Therefore, the metamorphic facies varies from greenschist to sub-greenschist.

Massive and pillowed basaltic rocks are frequently included as tectonic slivers within the Cosoltepec Formation and share the same deformation and metamorphism (Figure 2. 8). These volcanic rocks are frequently observed throughout the Cosoltepec Formation. Ortega-Gutiérrez (1975), and Ortega-Gutiérrez et al. (1999) also point out the presence of amphibolite at the base of this unit. Most outcrops of these volcanic rocks are

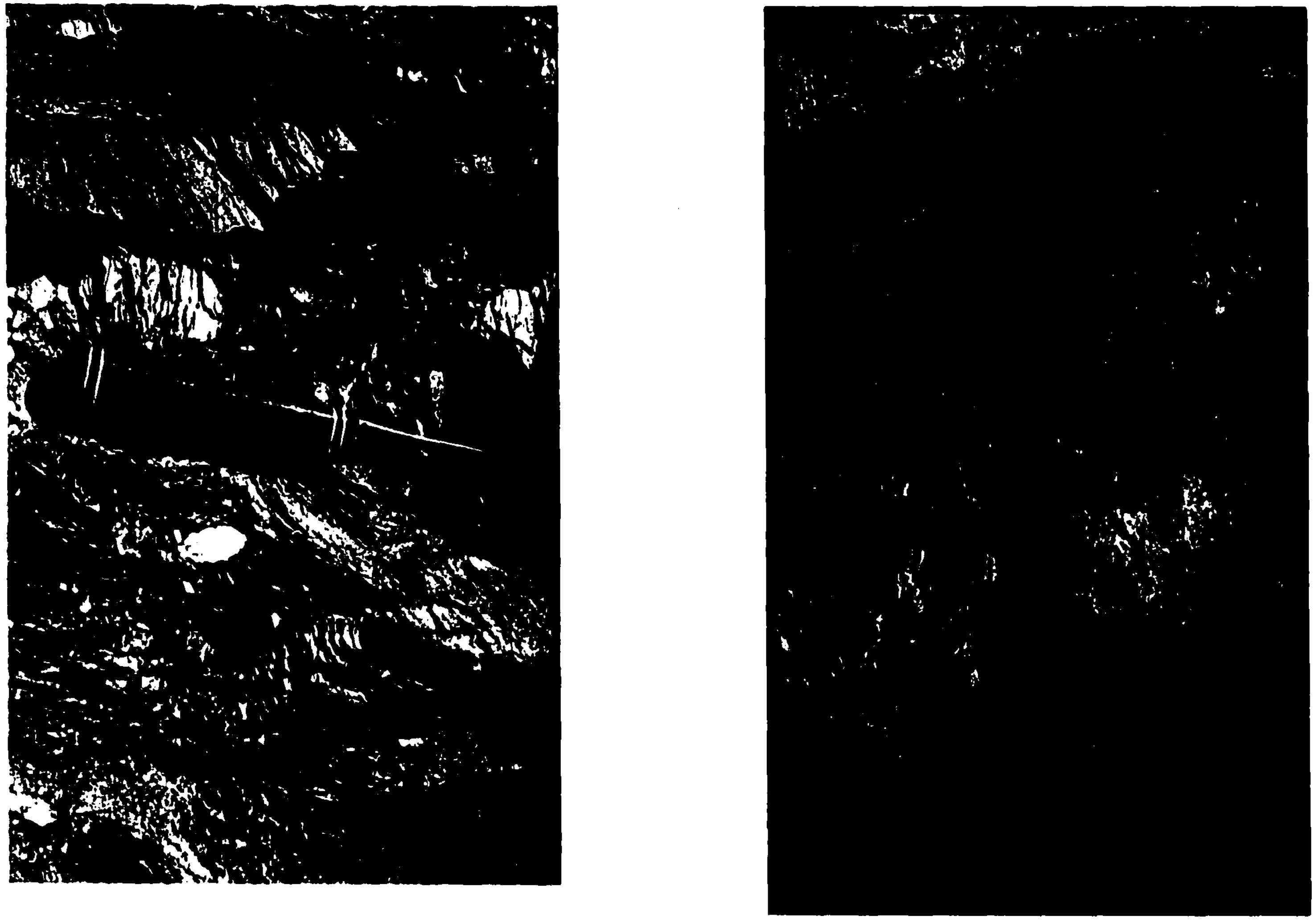


Figure 2.7. Main lithologies of the Cosoltepec Formation. To the left, phyllites showing folded foliation, to the right, quartzites displyng thin to middle bedding and planar foliation.

Figure 2.8a. Example of a volcanic sliver within the Cosoltepec Formation. Outcrop located along the road to Olinala town. Sample 39

Figure 2.8b. Pillowed basalt from the volcanic sliver located along the road to Olinala town. Cosltepec Formation (sample 37).

several tens of meters thick and because their dark color contrasts with the phyllite and quartzite of the surrounding sediments, they are easily identified. Approximately fourteen outcrops of these rocks have been recorded, however, only six samples were geochemically analyzed and results concerning their petrography and geochemistry are described in Chapter III.

Unfortunately, the age of the Cosoltepec Formation has not been determined; no fossils have been found and isotopic data are limited. A preliminary U/Pb study of detrital zircons from quartzites yielded an average age of 1800 Ma (Robinson et al., 1989; Robinson, 1990), and crustal residence ages (T_{DM}) ranging from 1.43 to 1.65 Ga strongly suggesting a Proterozoic continental source for the sediments (Yáñez et al., 1991). Considering that this unit is tectonically overlain by the Piaxtla Group of Late Ordovician-Early Silurian, as well as by the unconformable Tecomate Formation of probable Devonian age, the best estimated age for the deposition of the Cosoltepec Formation is considered as Cambrian (?)- Ordovician. Recently, one basaltic sample from this unit was analyzed by Ar^{39}/Ar^{40} yielding an integrated plateau age of 288 ± 13 Ma (Campa and López-Martínez, 2000), however, this age seems to be reset by the late Paleozoic magmatism reported by Torres et al. (1999). A Rb/Sr whole-rock age of 452 ± 22 Ma was reported by Armstrong (1979) *in* Ortega-Gutiérrez (1999). Therefore, the crystalization age of these rocks is not well defined.

Piaxtla Group

I introduce the new name of Piaxtla Group because it is the locality where the Xayacatlán Formation and the Esperanza Granitoids are closely related (Appendix 2). It is also the place where Ortega-Gutiérrez (1975, 1991, 1993, 1997) has studied and described most of the petrologic and structural characteristics of these units (Figure 2. 6a).

This group is made up of eclogite facies metabasites and metasediments of the Xayacatlán Formation, along with strongly deformed intrusives called Esperanza Granitoids. It is the most studied group within the Acatlán Complex (Ortega-Gutiérrez, 1974, 1975, 1991, 2000; Ortega-Gutiérrez et al., 1999; Ortega-Gutiérrez and Reyes-Salas, 1997; Meza-Figueroa et al., 1996; Meza-Figueroa, 1998). Originally, the Xayacatlán Formation and the Esperanza granitoids were included in different lithotectonic groups (Ortega-Gutiérrez, 1975) (Figure 2. 5a). Later on, these units were regrouped on the basis of the similar metmorphic grade, deformation style and tectonic relationship with respect to the other formations (Ortega-Gutiérrez, 1991) (Figure 2. 5b). Even more confusing is the fact that around the type locality of the Xayacatlán Formation crops out the Tecomate Formation, both of which were considered to be part of the Acateco Group (Ortega-Gutiérrez, 1975) (Figure 2. 5a). For these reasons, the introduction of the new name of Piaxtla Group is necessary.

The Piaxtla Group is mainly exposed in the northwestern and central-western regions of the outcropping Acatlán Complex (Appendix 2). It is represented by narrow belts of 5 to 10 km wide and 15 to 35 km long, trending NE-SW. In all the mapped outcrops, both the Xayacatlán Formation and the Esperanza Granitoids are closely associated sharing the same mylonitic deformation and eclogite facies metamorphism. This group represents an allochthonous high-pressure assemblage that tectonically overrides the low- pressure and moderate temperature assemblage of the Petlalcingo Group for more than 100 km, following a westward emplacement (Ortega-Gutiérrez, 1991, 1993). The Piaxtla Group is in turn overlain by the Tecomate Formation in a highly tectonized but certainly unconformable relationship (Figure 2. 6a-b) (Ramírez and Talavera, 1997; Ortega-Gutiérrez et al., 1999).

Xayacatlán Formation: The presence of eclogites around the Acatlán region was first reported and later termed Xayacatlán Formation by Ortega-Gutiérrez (1974, 1975). He defined this formation as a suite of igneous basic and ultrabasic rocks made up mostly of banded metagabbro, serpentinite, eclogitic rocks and greenschist. He also pointed out the difficulty to separate this unit from the Tecomate Formation and considered the Xayacatlán town as the type locality. Fieldwork carried out during the present study concluded that this locality is underlain by the Tecomate Formation, not by the Xayacatlán unit. However, this name was conserved because of its frequent use in most of the publications.

According to Ortega-Gutiérrez (1975, 1991) the high-pressure mineral associations developed in these eclogites, at least in the Piaxtla region, consists of omphacite-quartz-ganet-barroisite-phengite-rutile. He also calculated the peak metamorphic conditions of temperature and pressure in 550^{o} C and 12 kb respectively for the eclogite facies. Later, Meza-Figueroa (1998) also calculated similar values of 550^{o} C of temperature, but higher pressure conditions up to 15 kb for the eclogites of Piaxtla and Izucar de Matamoros regions. Commonly, eclogites are affected by intense retrogression in such a way that this lithology is widely recognized in the field as garnet amphibolites displaying amphibole, garnet, chlorite, biotite and epidote as the final product of metamorphic retrogression (Figure 2. 9a-b).

On the other hand, metasediments associated with the eclogites mostly include quartzite and pelitic micaschist rich in phengite, quartz, chlorite, garnet and rutile, which also represent an assemblage of high-pressure metamorphism. Muscovite, the most conspicuous mineral of these metasediments represents the final product of phengite retrogression (Ortega-Gutiérrez, 1991; Ortega-Gutiérrez and Reyes-Salas, 1997) (Figure 2. 10a-b)

The Xayacatlán Formation is one of the units that have been geochemically

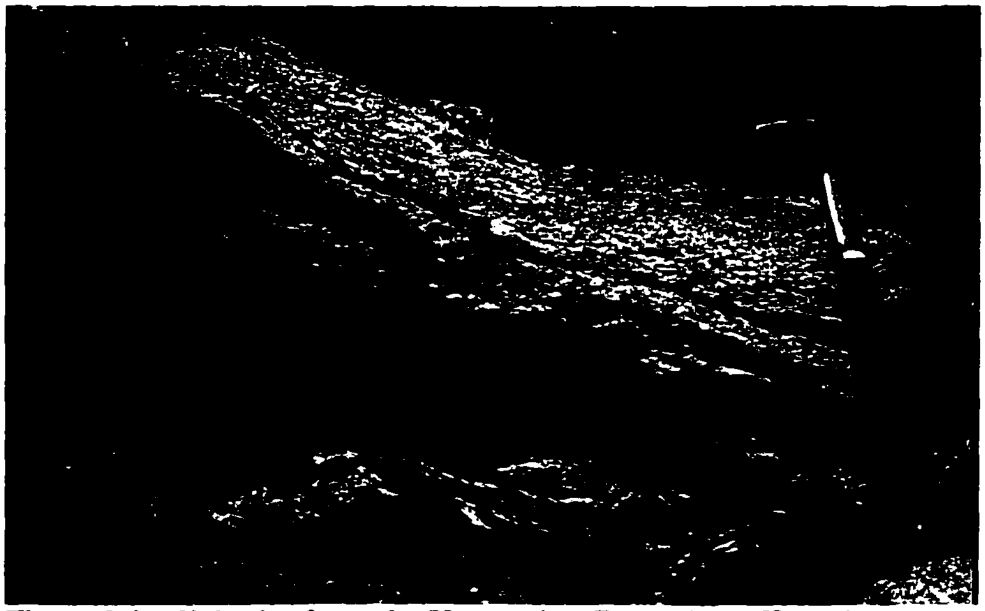


Figure 2.9a. Eclogite from the Xayacatlan Formation affected by migmatization. Outcrop from Izucar de Matamoros region (see figure 2.4)

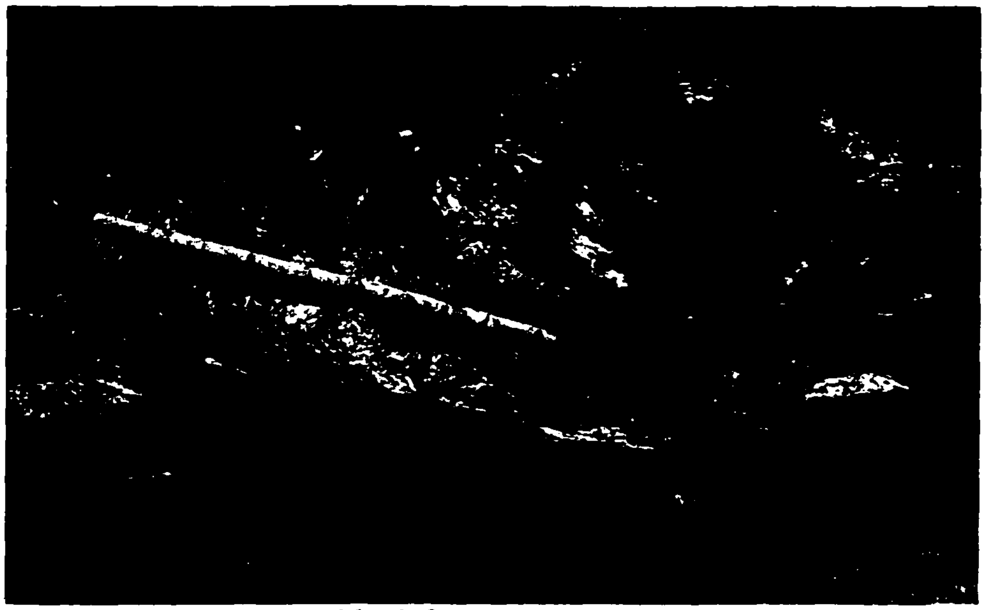


Figure 2.9b. Serpentine block from the Xayacatlan Formation. Outcrop from the Tlachinola region (see figure 2.4)

Figure 2.10a. Pelitic mica-schist rich in muscovite, quartz and garnet (metasediments of the Xayacatlan Formation. Outcrop from Izucar de Matamoros region.

Figure 2.10b. Detail of the pelitic mica-schist of the Xayacatlan Formation showing abundant garnet and muscovite. Outcrop from Izucar de Matamoros region.

analyzed by Ortega-Gutiérrez (1975) and Meza-Figueroa (1998). The former author presents results from twenty samples analyzed for major elements, three of which really belong to the Xayacatlán Formation (Piaxtla region), while the others seem to be part of the Tecomate Formation (lava flows and basic intrusions) that were originally mapped as the undifferentiated Acteco Group. Based on such data, in addition to petrographic information, he concluded that this unit represents part of an ophiolithic complex.

On the other hand, Meza-Figueroa (1998) presents major, REE and trace elements concentrations from eclogites of the Piaxtla and Izucar de Matamoros regions, which suggest magmas generated preferentially into two different oceanic settings: OIB (Ocenic Island Basalt)-MORB (Mid Ocean Ridge Basalt) and IAT (Island Arc Tholeiite) respectively. Meza-Figueroa (1998) points out that the high-pressure metamorphism that generated the eclogites, produced insignificant mobility in HFSE (High Field Strength Elements) and REE (Rare Earth Elements) concentrations such that the determined tectonic setting is tightly constrained.

Eclogites of the Xayacatlán Formation yielded a Devonian metamorphic age of 388 ± 44 Ma by means of Sm/Nd garnet-whole rock isochron (Yáñez et al., 1991). However, the crystallization age of this unit is still not defined, but necessarily it would be older than the 425 to 440 Ma calculated for the Esperanza Granitoids which intrude the Xayacatlán Formation. Therefore the protolith age could be Cambrian to Early Ordovician.

Esperanza Granitoids: One of the most outstanding features of the Acatlán Complex is the presence of highly deformed and metamorphosed plutonic bodies, which were collectively defined as Esperanza Granitoids (Ortega-Gutiérrez, 1975; 1978; 1981). Early description of these intrusive bodies included all granitoids, aplites and pegmatites dynamically deformed (mylonitized) and poly-metamorphosed. Original relationships were not clearly established but considered to be mainly intrusive into the Acateco Group and apparently intrusive into the Petlalcingo Group and even into the Grenvillian Oaxaca Complex further east (Ortega-Gutiérrez, 1975). Therefore, this name was used to differentiate the most deformed granitoids from the least deformed Totoltepec stock and the clearly post-tectonic and undeformed San Miguel dykes (Figure 2. 5a). Later on, the Esperanza Granitoids were definitively associated with the Xayacatlán Formation due to the similar metamorphic grade but erroneously considered to be of Devonian age (Ortega-Gutiérrez, 1991; Yáñez et al., 1991). More accurate, Late Ordovician-Silurian U/Pb zircon ages ranging from 440 to 425 Ma has been obtained from two different localities of the Esperanza granitoids (Robinson, 1990; Ortega-Gutiérrez et al., 1999). This data leads to the conclusion that there were different granitoids within the Acatlán Complex sharing similar mylonitic deformation but different metamorphism and ages.

At present, the best definition of the Esperanza Granitoids includes all those intrusive bodies highly deformed and metamorphosed to the eclogite facies closely associated and intruding the Xayacatlán Formation which together constitute an allochthonous assemblage that overthrusts the Petlalcingo Group. In turn, the Esperanza granitoids are unconformably overlain by the Tecomate Formation (Figure 2. 6a-b). In addition, the Esperanza Granitoids are clearly intruded by a suite of leucocratic granites, which also display variable degree of mylonitic deformation. Those granites are of undetermined age but probably could be correlated with the Totoltepec pluton of 287 Ma.

The Esperanza Granitoids crop out as elongated to tabular-shaped bodies of very different dimension, composition and fabric. The longest outcrop is a north-south trending plutonic complex (ca. 100 km long; 15 km wide) running from the northern to the central region of the Acatlán Complex (south of Coayuca to west of Mariscala de Juárez towns) (Appendix). Northwest of the Olinalá region another important plutonic complex of the

Esperanza Granitoids (30 km long; 10 km wide) crops out. It is trending in a NE-SW direction from Linderos del Sur to Zompazolco towns. In addition, two smaller outcrops (8-10 km long; 4-2 km wide) around Olinalá village are running in the same NE-SW direction. Finally, northwest of Huajuapan de León, granitoids of this unit (20 km long; 5 km wide) trend in a NW-SE direction (Appendix 2). The are also some outcrops in the vecinity of the Rodeo and Las Minas towns (east of Izucar de Matamoros region), but because of the small size they were not mapped. In all cases the Esperanza Granitoids are always associated with the Xayacatlán Formation.

The largest outcrops of the Esperanza Granitoids constitute plutonic complexes displaying a wide variation in composition and texture. Some outcrops range from granites to gabbro (Olinalá region) and some others from granite to diorite (Tehuitzingo region). Because of the observed variations, these two regions were selected to conduct geochemical analyses on selected samples (Figures. 2. 3 and 2. 4).

The most representative lithology consists of megacrystic K-feldspar augen gneiss, although displaying important variations from fine grained augen gneiss to augen schist (Figure 2. 11a-b). The most abundant minerals depend on the composition of the sampled body, most consist of quartz, K-feldspar, plagioclase, biotite, amphibole and garnet. However, a high-pressure assemblage has been reported by Ortega-Gutiérrez (1975, 1991, 1997) including phengite-garnet-rutile-zoisite and garnet-biotite-phengite-rutile-tourmaline, which define the eclogitic facies. Therefore, the most abundant and obvious mineralogy constitutes the final product of retrogression. Some outcrops also consist of garnet micaschist and migmatites (Figure 2. 12a).

All textures are cataclastic and vary from protomylonites to mylonites and even ultramylonites. This type of deformation is shared by the Xayacatlán Formation, in such a way that the relationship between both units is expressed by a parallel foliation. However,

Figure 2.11a. Megacrystic k-feldspar augen gneiss from the type locallity of the Esperanza Granitoids. Outcrop along the highway to Acatlan town (see figure 2.4)

Figure 2.11b. Esperanza granitoid showing mylonitic texture and rotated porphyroblasts.

Figure 2.12a. Mylonitic Esperanza granitoid showing evidence of migmatization. Outcrop from Santa Cruz region (see figure 2.4).

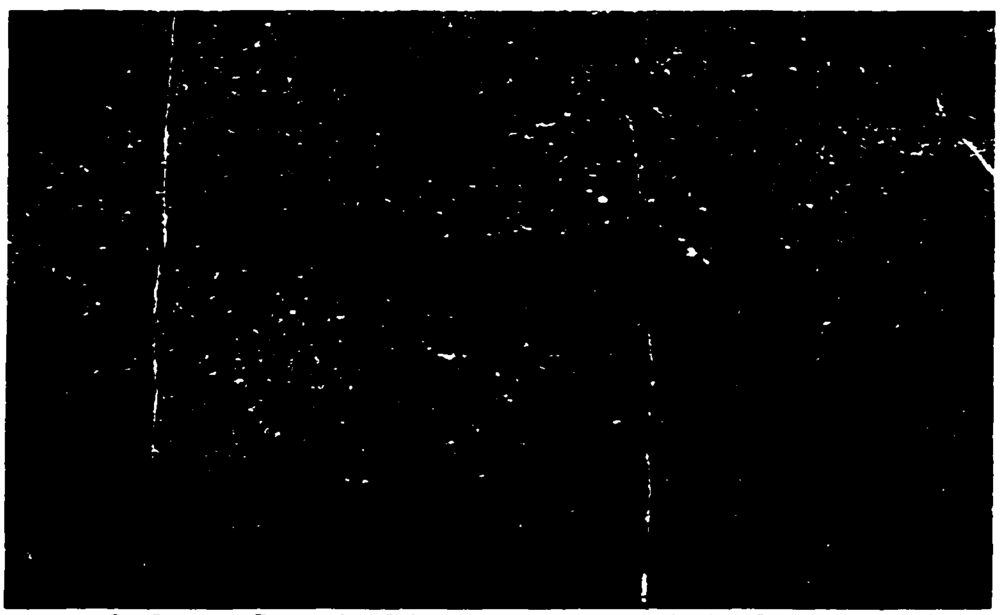


Figure 2.12b. Mafic rock of the Esperanza granitoid. Sample 245 from the region of Tecolapa (see figure 2.3).

the inclusion of large xenolithic blocks of the Xayacatlán Formation within the Esperanza Granitoids, as well as the presence of foliated sills within the Xayacatlán Formation close and parallel to the contact zone with the Esperanza Granitoids suggest that the former unit is intruded by the latter. In addition, the Esperanza Granitoids are clearly intruded by a suite of mylonitic and leucocratic granites, as seen by the presence of numerous stocks in the Olinalá region (Figure 2. 4).

Early attempts to date the Acatlán Complex were done by Rb/Sr isotopic analyses (mineral and isochron ages) conducted on the most representative augen-schist lithology of the Esperanza Granitoids (Fries and Rincón-Orta, 1965; Halpern et al., 1974; Ruiz-Castellanos, 1979). Such isotopic analyses systematically yielded Ordovician ages ranging from 443 to 480 Ma, but all of them with large error values. More precise U/Pb isotopic analyses carried out in the same previously dated lithology have yielded younger ages ranging from 425 to 440 Ma. (Robinson, 1990; Ortega-Gutiérrez et al., 1999). These Late Ordovician-Early Silurian ages are, at this time, the most accurate for the Esperanza Granitoids. However, because of the large extent and compositional variation of these plutonic complexes, further isotopic analyses are necessary to accurately determine the whole magmatic event.

Tecomate Formation

Initially, the Tecomate Formation was defined by Rodríguez-Torres (1970) as a late Paleozoic marine sequence consisting mainly of siltstone, graywacke, conglomerate, quartzose sandstone and laminated limestone. Subsequently, the Tecomate and Xayacatlan Formations were regrouped by Ortega-Gutiérrez (1975, 1978) to define the Acateco Group of pre-Mississipian age. In addition, the concept of undifferentiated Acateco Group was also used for those outcrops where the main characteristics of either the Tecomate or

Xayacatlán Formations were not clearly determined (Ortega-Gutierrez, 1975, 1978). Field obsrvations suggest that both the Tecomate and Xayacatlán Formations show very distinctive metamorphic grade and lithology and can be easily distinguished. Thus, the use of the Acateco and undifferentiated Acateco Group names should be abandoned (Figures 2. 6a-b).

The Tecomate Formation is widely distributed in the north-central region of the Acatlán Complex (Appendix). Most of these outcrops trend in NNE-SSW to NW-SE directions following large folded west-verging-structures for more than 25 km. Smaller exposures of this formation are also found around Olinalá town forming part of a NE-SW trending syncline. At this time, the Tecomate Formation is considered to be the uppermost metamorphic unit of the Acatlán Complex, therefore representing, the earliest post-tectonic assemblage (Figure 2. 6a-b).

Even though the lower contact is highly tectonized (mylonitized), studies based on structural data and sedimentary provenance suggest that this formation is unconformably overlying either the Petlalcingo or the Piaxtla Groups (Ortega-Gutiérrez, 1993; Sánchez-Zavala and Ortega-Gutiérrez, 1997, 1998). In turn, the Tecomate Formation is unconformably overlain by unmetamorphosed shallow marine and continental sediments of Carboniferous and Permian ages (Patlanoaya, Olinalá and Matzitzi Formations). Besides, the Tecomate Formation is also affected by multiple granitic intrusions like the Hornos granite of Late Devonian, the Totoltepec stock of Late Pennsylvanian and even some granites of probable Permian age (Figures 2. 3 and 2. 4).

Earliest descriptions of the Tecomate Formation only considered the presence of shallow marine sediments made up mostly of banded semipelites, coarse grained psammites, as well as minor conglomerates and very distinctive horizons of marble. More recently, it was redefined as a volcano-sedimentary sequence including an important package of volcanic and associated intrusive rocks (Ramírez and Talavera, 1997). The lower part of this formation consists of coarse to fine grained sandstones interlayered with mainly massive lava flows. Sandstone varies from mostly arkose to graywacke and quartzite, and finer sandstone show tuffaceous influence (Figure 2. 13a-b). Main minerals in the arkosic sandstones are K-feldspar, plagioclase and quartz, while the matrix contains abundant white mica. Graywackes consist mostly of plagioclase and quartz within a matrix enriched in white mica. Stratification varies from thin to massive beds.

On the other hand, volcanic rocks are mainly basaltic in composition with scarce rhyolites; this association suggests a bimodal magmatic suite where andesites are absent. Even though the basaltic lavas were ejected in a marine environment, the absence of pillow lavas is noteworthy. Most basalts are massive flows and some present small nodules of recrystallized limestone. Rhyolites present porphyritic and fluidal textures displaying small phenocrysts of quartz and K-feldspar. Gabbros and sub-volcanic plutons of basic composition also intrude both sediments and lavas from the lower part of this formation. Gabbros consist mainly of plagioclase and pyroxene displaying textural variations, ranging from phaneritic in gabbros to porphyritic in sub-volcanics rocks. Some basic inrusives are completely serpentinized (i.e., around Olinalá town).

The whole sequence is metamorphosed to greenschist facies, in addition to the strong deformation which develops isoclinal folds, crenulation and very intense cataclasis toward the base (Figure 2. 13a, and 2. 14a-b). Magmatic rocks of this formation were geochemically analyzed and results are presented in the following chapter.

The upper part of the Tecomate Formation consists of a thick sequence of graywacke, arkose, quartzite, conglomerate and shallow marine limestone. Despite the strong deformation and greenschist metamorphism, primary structures in this part of the formation can be identified (Figure 2. 13b). Laminations, slumps and graded bedding are

Figure 2.13a. Metasediments of the Tecomate Formation showing isoclinal folds

Figure 2.13b. Metasediments of the Tecomate Formation displying primary structures.

Figure 2.14a. Strongly deformed sediments of the Tecomate Formation.

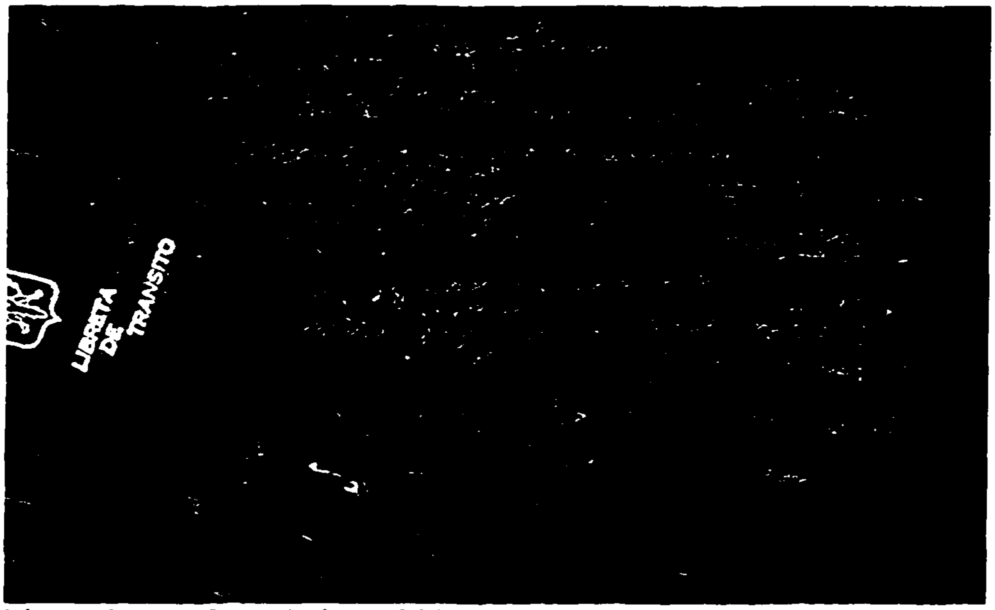


Figure 2.14b. Crenulation within the metasediments of the Tecomate Formation.

some of the identified structures. Along the road between Totoltepec and Xayacatlán towns, sandstones display detrital hornblende and pebbles of gabbros and basic gneisses (5 to 10 cm). However, some fragments are larger reaching up to 20 cm. Most characteristic features of the distinctive limestones and conglomerates of this formation have been described by Ortega-Gutiérrez (1975, 1978) and Sánchez-Zavala and Ortega-Gutiérrez (1997). However, in conglomerates, the most important feature to be taken into account is the presence of granitic fragments derived from either the Esperanza granitoids or the Oaxaca Complex. In limestones fossiliferous horizons include crinoids and mollusks, unfortunately most organisms are not well preserved and they are not age-diagnostic (Ortega-Gutiérrez, 1978).

The age of the Tecomate Formation is constrained by its stratigraphic position and plutonic cross cutting relationships. The unit is unconformably overlain by the Patlanoaya, Olinala and Matzitzi Formations considered to range from Early Missisippian to Permian age by means of paleontological data (Flores and Buitrón, 1981; Vázquez-Echavarria, 1986; Villaseñor-Martínez et al., 1987; Silva-Pineda, 1970; González-Arreola et al., 1994; Brunner, 1994). On the other hand, the underlying Piaxtla Group, including the Esperanza granitoids of Late Ordovician-Silurian age, constrains the lower limit of the Tecomate Formation. Therefore, the best estimated age for this assemblage is Early-Middle Devonian. This age is supported by the Late Devonian granite that intrudes the Tecomate Formation around Los Hornos and La Noria towns in the Tehuitzingo region (Yáñez et al., 1991) (Figure 2. 4).

Middle to Upper Paleozoic Granites

In addition to the described units of the Acatlán Complex, there are strongly to slightly deformed granites that clearly intrude the Petlalcingo and Piaxtla groups as well as the Tecomate Formation and the unmetamorphosed upper Paleozoic sediments. Most of these granites show mineralogic and textural variation with respect to the Esperanza granitoids, however, some times they could be mistaken due to the similar mylonitic textures that some of them display. Based on field relationships as well as chronological, textural, mineralogical, and chemical parameters, the analyzed granitoids have been subdivided into three different groups defined as: 1) Hornos-La Noria type granite; 2) Teticic type granites and 3) Totoltepec granite. However, further work certainly will require additional subdivisions or new regrouping if chronological data are available. Representative samples of these granites were geochemically analyzed and sampled outcrops are shown in figures 2. 3, and 2. 4. The following description is based on chronological order from older to younger granites.

Hornos-La Noria granites. This type of granite has only been identified in the northern Tehuitzingo region as two isolated but possibly continuous bodies (Figure 2. 4 and Appendix 2). In the Tehuitzingo area, these granites intrude rocks of the Tecomate Formation and are in turn unconformably overlain by upper Paleozoic sediments of the Olinalá-Patlanoaya Formations. The Hornos-La Noria granites are characterized by a very distinctive megacrystic texture where pink K-feldspar grains are enclosed in a coarse granular matrix made up of quartz, plagioclase, biotite and scarce hornblende (Figure 2. 15a-b).

In a cross section from La Noria to La Colonia villages (Figure 2. 4), this pluton exhibits different degrees of cataclasis and mylonitic deformation, sometimes producing an augen-schistose fabric very similar to that present in the Esperanza granitoids (Figure 2. 16a-b). Further detailed mapping and petrographic analyses in large deformed augen-megacrystic plutons all over the Acatlán Complex will certainly separate the Hornos granite from the augen-schistose lithology of the Esperanza-type granitoids. Despite the

Figure 2.15a. Typical Hornos granite showing megacrystic texture. K-feldspar grains are enclosed in a coarse granular matrix of quartz, plagiclase and biotite.

Figure 2.15b. The Hornos granite displying scarce crystals of hornblende.

Figure 2.16a. The Hornos granite displying an incipient mylonitic deformation.

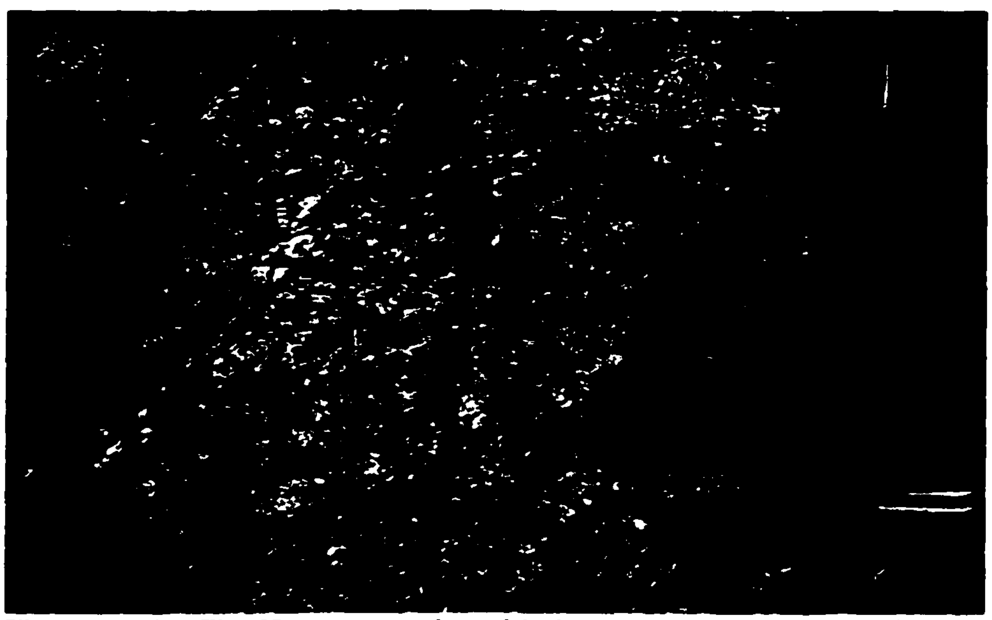


Figure 2.16b. The Hornos granite with the presence of innumerables enclaves. Sample 101 (see figure 2.4).

close proximity of these granites to the Esperanza granitoid the relationship between the plutonic bodies was not observed, however, it is assumed to be intrusive (Figure 2. 4). Detailed petrographic description from the Hornos granite will be provided in the following chapter in addition to geochemical analyses.

This type of granite was dated in La Noria locality by zircon U/Pb geochronology and yielded a Middle Devonian age of 371 ± 34 Ma. (Yáñez et al., 1991). For years, this age was erroneously considered to be the representative age of the highly deformed Esperanza granitoids mainly because of the similar mylonitic deformation which is shared by different granites and the lack of more chronologic data (Ortega-Gutiérrez, 1991, 1993).

Totoltepec granite. This granite crops out northeast of Acatlán town and displays the typical shape of a stock (Appendix 2). To the south, it intrudes the Tecomate Formation which adopts a circular distribution similar to the pluton, whereas to the north, it is affected by a normal fault which juxtaposes Mesozoic rocks at the same level of the granite. Structurally, the Totoltepec granite shows a weak foliation and more locally shear planes. On the other hand, the essential minerals are plagioclase, K-feldspar, quartz and biotite, where biotite is mostly altered to chlorite. This granite was geochemically analyzed in the present work and results are given in the following chapter.

The slightly metamorphosed Totoltepec granite was firstly studied by Fries et al. (1963) who obtained an Ordovician age using the Pb/alpha method. The granite was geochemically analyzed and classified as a leuco-granodiorite. Subsequently, Ortega-Gutiérrez, (1975) classified this pluton as a trondhjemite but correlated it to the Esperanza granitoids. Finally, Yáñez et al. (1991) conducted a new geochronologic study obtaining a concordant U/Pb zircon age of 287 ± 2 Ma (Late Pennsylvanian) for this granite.

Teticic granites. These granites are mostly represented by a very white high-silica phase (leucocratic) which includes aplitic to fine-grained dikes, as well as medium to fine- grained stocks. These high-level granites constitute the only group that presents clear intrusive relationships to all the metamorphic groups included in the Acatlán Complex (Petlalcingo, Piaxtla and Acateco) (Figures 2. 3 and 2. 4). This type of granite also exhibits textural variations ranging from mylonites to slightly deformed fine-grained fabrics (Figure 2. 17a-b). Locally, some dikes display a well-developed micro-augen schistose phase with small crystals of quartz (0.5 to 1 cm) as the main porphyroblasts. Most mylonitized stocks with banded foliation are closely associated with the Esperanza granitoids, whereas those intrusives displaying medium size grains and weak foliation seem to be more associated with the Tecomate Formation and therefore seem to be a younger event (Figure 2. 18 a-b).

Unfortunatelly, none of these intrusive bodies have been dated, but a correlation to the slightly deformed and well-dated Totoltepec granite could be possible suggesting a Late Carboniferous age. Some of these leucocratic granites were geochemically analyzed and results are shown in the following chapter.

Pemian Intrusives. There is evidence of an even younger magmatic event represented by coarse-grained granites, but some of them also display mylonitic fabrics. The classic example is the Cozahuico granite, which intrudes the contact between the Acatlán and the Grenvillian Oaxaca Complex nearby the Caltepec region (Appendix 2). This granite is also affected by variable extent of cataclasis ranging from relatively undeformed to augen schistose and mylonitic textures. This granite was originally considered as Devonian (U/Pb zircon age of 373 Ma) by Elías-Herrera and Ortega-Gutiérrez (1998), and correlated with the Hornos-La Noria granite. However, this granite, previously named Caltepec by Ruíz-Castellanos (1979), yielded a Rb/Sr isochron age of 269 ± 21 Ma (Early Permian), in addition to another K/Ar age of 266 ± 13 Ma determined

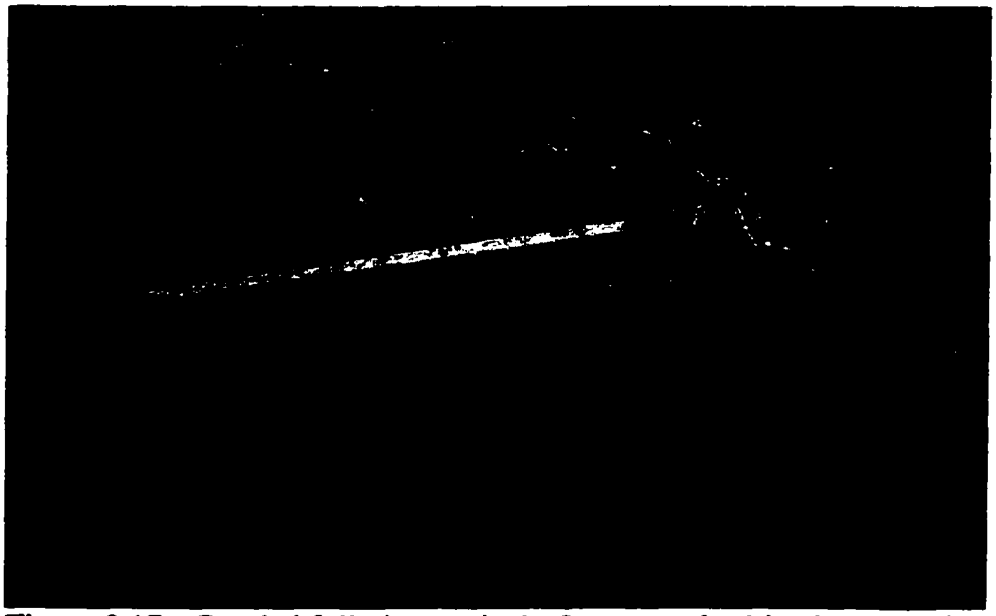


Figure 2.17a. Banded foliation typical of most mylonitized stocks of the Teticic granite. This type of granite is more associted to the Esperanza granitoids.

Figure 2.17b. Metric dikes included within the Teticic granites clearly intrude the Tecomate Formation.

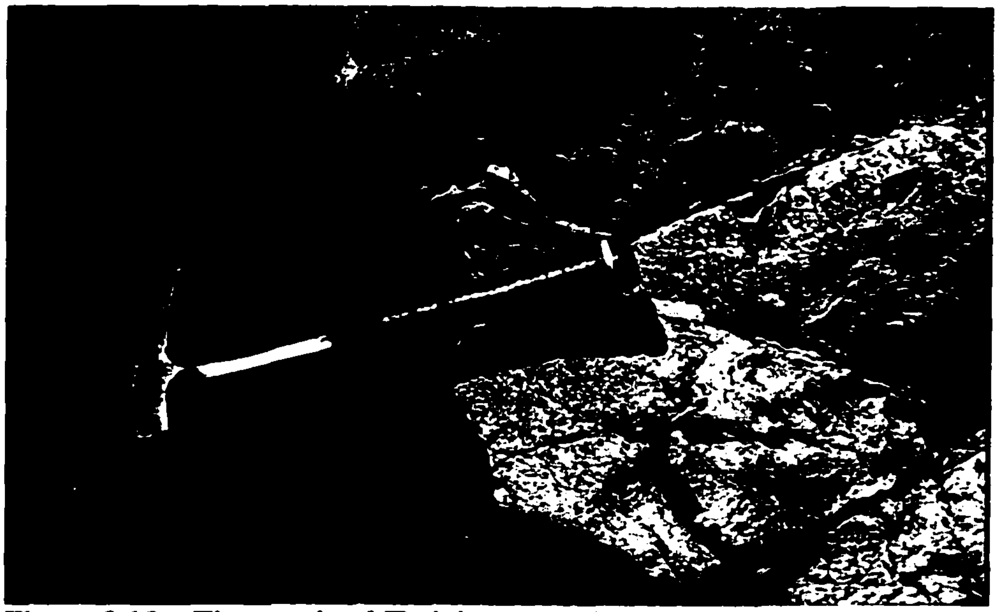


Figure 2.18a. Fine-grained Teticic granite intruding metasediments of the Tecomate Formation.

Figure 2.18b. Detaile of a fine-grained Teticic granite showing inclusions from the Tecomate Formation.

by Torres et al. (1986, 1999). More recently, the Permian age of this granite has been confirmed by a concordant U/Pb zircon age (Ortega-Gutiérrez, oral comunication). It is important to point out that due to the lack of more isotopic ages and similar deformation present in different granites, some tectonic interpretations have been mistaken.

Upper Paleozoic Sediments

The Acatlán Complex is unconformably overlain by an undeformed shallow marine to continental sequence, similar to that covering the Oaxaca Complex. The well-defined unconformity of this sequence places the upper limit on the metamorphic evolution of the Acatlán Complex and clearly determines the minimum age for the amalgamation of the Mixteco and Oaxaca terranes. The sequence overlapping the Acatlán Complex consists of three formations: Olinalá, Patlanoaya and Matzitzi.

The Matzitzi Formation crops out to the notheastern region covering the Mixteco-Oaxaca boundary (Appendix 2). This unit consists mostly of clastic sediments including sandstone and shale interbedded with thick layers of conglomerates. The finest sediments present abundant fossil flora which has been dated as Pennsylvanian to Permian (Silva-Pineda, 1970; Weber and Ceballos, 1994). The Olinalá Formation crops out near to Olinalá town (Figure 2.3). It consists of clastic sediments including sandstone, shale and sandy conglomerate, whereas carbonates include sandy limestone. This formation represents a deltaic to shallow marine setting and fossils suggest an Early Permian age (Flores de Dios et al., 2000; González-Arreola et al., 1994). The Patlanoaya Formation crops out to the northwest (Appendix 2), being the most complete sequence of the Late Paleozoic. It is similar to the Olinalá Formation including sandstone, shale, sandy conglomerate and limestone; however, the abundant fossil fauna indicate a Mississippian to Early Permian age (Villaseñor-Martínez et al., 1987; Brunner, 1987).

CHAPTER III. PETROGRAPHY AND GEOCHEMISTRY

3. 1. Objectives, sampling and analytical methods

As previously described the Acatlán Complex displays several magmatic units through the whole stratigraphy, but none of the previous studies had conducted systematic geochemical analyses. The most important works carried out in this complex emphasized petrologic and isotopic analyses (Ortega-Gutiérrez, 1974, 1993; Robinson, 1990; Yáñez et al, 1991), but all of them lacked systematic geochemistry. Accordingly, this chapter presents the results of whole-rock geochemistry and Sm/Nd isotopic analyses conducted in metavolcanic rocks of the Cosoltepec Formation (Petlalcingo Subgroup), metavolcanic rocks of the Tecomate Formation and three different metaplutonic suites herein called Esperanza, Hornos and Teticic granites respectively. These data, will contribute to better definition of the tectonic setting and evolution of the main magmatic units included in the Acatlán Complex.

Tables 3. 1, 3. 2, and 3. 3 (Appendix 1), present whole-rock geochemical data of 44 selected samples after 135 rocks were petrographically analyzed and screened to minimize the effects of alteration and weathering. Six samples belong to the Cosoltepec Formation, twenty to the Tecomate Formation and the remaining 18 to plutonic rocks. In addition, 21 samples were selected to conduct Nd isotopic analyses (Table 3. 4). Eleven samples are from the Tecomate Formation, and the other ten samples belong to different granitoids (6 from the Esperanza granitoids, 1 from the Hornos and 3 from the Teticic granites).

Whole-rock analyses (major, rare-earth and selected trace elements) were conducted at SGS-XRAL Laboratories, a Division of SGS Canada Incorporation in Ontario, Canada. Major-element oxides and the trace elements Rb, Ba, and in some samples Zr, were analyzed by X-ray fluorescence. Precision in those cases is better than 0.01% in oxides and 2 ppm for trace elements. The trace elements Hf, Nb, Sr, REE and most Zr were determined by ICP-MS, whereas Th, and Ta were analyzed by INAA. Precision varies from 0.2 to 0.05 ppm in ICP-MS and 0.5 to 1 ppm in INAA.

Nd-isotopic analyses were conducted on a VG-354 mass spectrometer at the Geosciences Department of the University of Arizona in Tucson, using techniques described by White and Patchett (1984) and Patchett and Ruiz (1987). Epsilon Nd (ε_{Nd}) for volcanic rocks of the Tecomate Formation were calculated assuming an absolute age of 400 Ma, and present-day bulk Earth ratios of $^{147}Sm/^{144}Nd = 0.1966$, and $^{143}Nd/^{144}Nd = 0.512638$ (Jacobsen and Wasserburg, 1981). All $^{143}Nd/^{144}Nd$ ratios were also corrected for isotope fractionation to $^{146}Nd/^{144}Nd = 0.7219$. Epsilon Nd ($\varepsilon_{Nd}$) for granitoids were calculated assuming the absolute ages of 440, 370 and 287 Ma corresponding to the Esperanza, Hornos and Teticic granites respectively. Crustal residence ages (T_{DM}) were determined using De Paolo's depleted-mantle evolution curve (1981).

Basic volcanic rocks are classified using major elements according to the TAS diagram (Total Alkalai-Silica) of Cox et al. (1979). However, because of the metamorphism of these rocks, the classification suggested by Winchester & Floyd (1977) based on HFS elements was also used, as these elements are immobile during metamorphism and alteration. On the other hand, rhyolites and granitic rocks are classified using the TAS diagram of Wilson (1989), in addition to the normative classification (An-Ab-Or) proposed by Barker (1979).

Rare-earth elements in volcanic and plutonic rocks were normalized to C1 chondrite average according to values determined by Evensen et al. (1978). In basic volcanic rocks, trace elements are analyzed and normalized using the N-MORB (Mid

Ocean Ridge Basalt) concentrations suggested by Pearce (1983), while rhyolites and granites are normalized using the ORG (Oceanic Ridge Granite) pattern proposed by Pearce (1984).

In order to distinguish between basalts produced in different tectonic evironments, discriminant diagrams proposed by Pearce and Cann (1973), Shervais (1982) and Wood (1980), were preferentially used. Discrimination diagrams employed for rocks of granitic composition were those recommended by Pearce et al. (1984), and Maniar and Piccoli (1989). Additional diagrams were used to characterize the aluminium index introduced by Shand (1927) and alkali-lime index of Peacock (1931), as well as the diagrams of Debon and Lefort (1983) and Chappell and White (1992) to correlate chemical and mineralogic characteristics.

Petrography and geochemical analyses are first presented for the volcanic units: Cosoltepec Formation (Petlalcingo Subgroup) and Tecomate Formation and then for the granitoids: Esperanza, Hornos and Teticic granites. Location of the analyzed samples is shown in figures 2. 3, and 2. 4.

3. 2. Cosoltepec Formation (Petlalcingo Group)

Petrography of igneous rocks

Metavolcanic rocks in the Cosoltepec Formation occur as massive and pillowed flows forming tectonic slices within the sequence of quartzites and phylites. Most samples display strong alteration and penetrative deformation, however, igneous textures are easily reconized. All samples are basaltic in composition with silica content (SiO_2) lower than 49.8 wt% (Table 3.1). Under the microscope, samples display quenched textures (skeletal, variolitic, and sub-variolitic), but in general textures vary from aphyric to doleritic and highly porphyritic. Quenched textures are present in the outer part of most pillows and on top of massive flows, whereas doleritic textures are restricted to the inner zone of the pillows. Porphyritic textures (30% vol. of phenocrysts) are well-developed in massive pillows. These textures are characterized by the presence of microphenocrysts of fresh clinopyroxene and partly or completely altered plagioclase both included in a highly recrystallized groundmass, which contains abundant chlorite-smectite, epidote and oxides.

Aphyric basalts are composed of epidote pseudomorphs after plagioclase or microphenocrysts of pyroxene enclosed in a groundmass of plagioclase varioles and microlites. Doleritic basalts include ophitic intergrowths of plagioclase, clinopyroxene, oxides, and interstitial recrystallized glass. Most rocks contain amygdules filled of variable proportions of calcite, quartz, chlorite-smectite, epidote and pumpellyite.

Whole- rock Geochemistry of igneous rocks

Six samples from the Cosoltepec Formation were analyzed for major, rare earth and trace elements (concentration are shown in Table 3. 1). Despite the narrow range in silica concetration, volcanic rocks exhibit high variation in K_2O (0.04 - 0.55 wt%), Na_2O (1.46 - 6.22 wt%), and CaO (4.05 - 13.2 wt%). Same variation is observed in the most incompatible and mobile trace elements Rb (<2 - 25.5 ppm), Ba (68.5 - 1760 ppm), and Sr (44 - 608 ppm). Such variations could be produced by the low-grade metamorphism that affected the Cosoltepec Formation. Most rocks are characterized by low to moderate contents in MgO (2.4 - 8.58 wt%), Cr (38 - 390 ppm), and Ni (42 - 132 ppm). These values are coincident with the low to moderate Mg # number [100 (MgO/(MgO + FeO + Fe_2O_3)] present in these samples, which range from 15.4 to 50.2, suggesting that the magmas underwent an early differentiation event.

On the other hand, the wide variation in concentrations of the less mobile HFS elements: TiO_2 (0.94 - 3.8 wt%), Nb (3 - 37 ppm), Zr (41 - 299), Y (38 - 16), as well as the strong variation in $(La/Yb)_N$ = 1.0 - 12.9 ratios, suggest that volcanic rocks of the Cosoltepec Formation are not related one to another. These contrasting values but could reflect magma genesis in more than one tectonic setting (Pearce and Cann, 1973).

According to the TAS classification (Cox et al., 1979), but mainly based on the classification of Winchester and Floyd (1977), samples group in alkaline and subalkaline basaltic rocks (Figure 3. 1). This result is reinforced by TiO_2 concentrations, which define two very different volcanic suites: (i) Ti-rich-alkaline basalts with concentrations higher than 2.4 wt% (samples 37 and 110), and (ii) Ti-poor tholeiitic basalts with TiO_2 concentrations lower than 1.5 wt% (samples 39, 91, 126 and 230).

Ti-rich alkaline basalts contain high concentrations of TiO_2, Zr, Y and Nb, which is very typical of oceanic island basalts (OIB) settings (Pearce and Cann, 1973). MORB-normalized multi-element patterns also display a general enrichment in both low field strength elements (LFSE) and high field strength elements (HFSE), with the exception of Y and Yb wich do not exhibit such enrichment (Figure 3. 2a). These patterns are similar to those present in within plate settings where magmas are derived from an enriched mantle source (Pearce, 1982). Chondrite-normalized REE patterns (Evensen et al., 1978) show strong enrichment in light REE relative to heavy REE. $(La/Yb)_N$ ratio = 4.0 in sample 37 is typical of transitional OIB magmas, whereas sample 110 shows an even greater $(La/Yb)_N$ ratio = 12.9, which is very distinctive of alkaline OIB (Figure 3. 2b).

On the other hand, tholeiitic basalts with low TiO_2 content also display lower concentrations in Nb, Zr, and Hf, as well as lower Zr/Y ratios. Multi-element patterns of this group of samples are characterized by rather flat spectra depicting variation in the most mobile elements (Figure 3. 2a). These patterns are quite similar to those reported in typical MORB settings (Pearce, 1983). However, most samples exhibit very small but distinctive negative anomalies in Ti and Zr, a characteristic commonly attributed to the

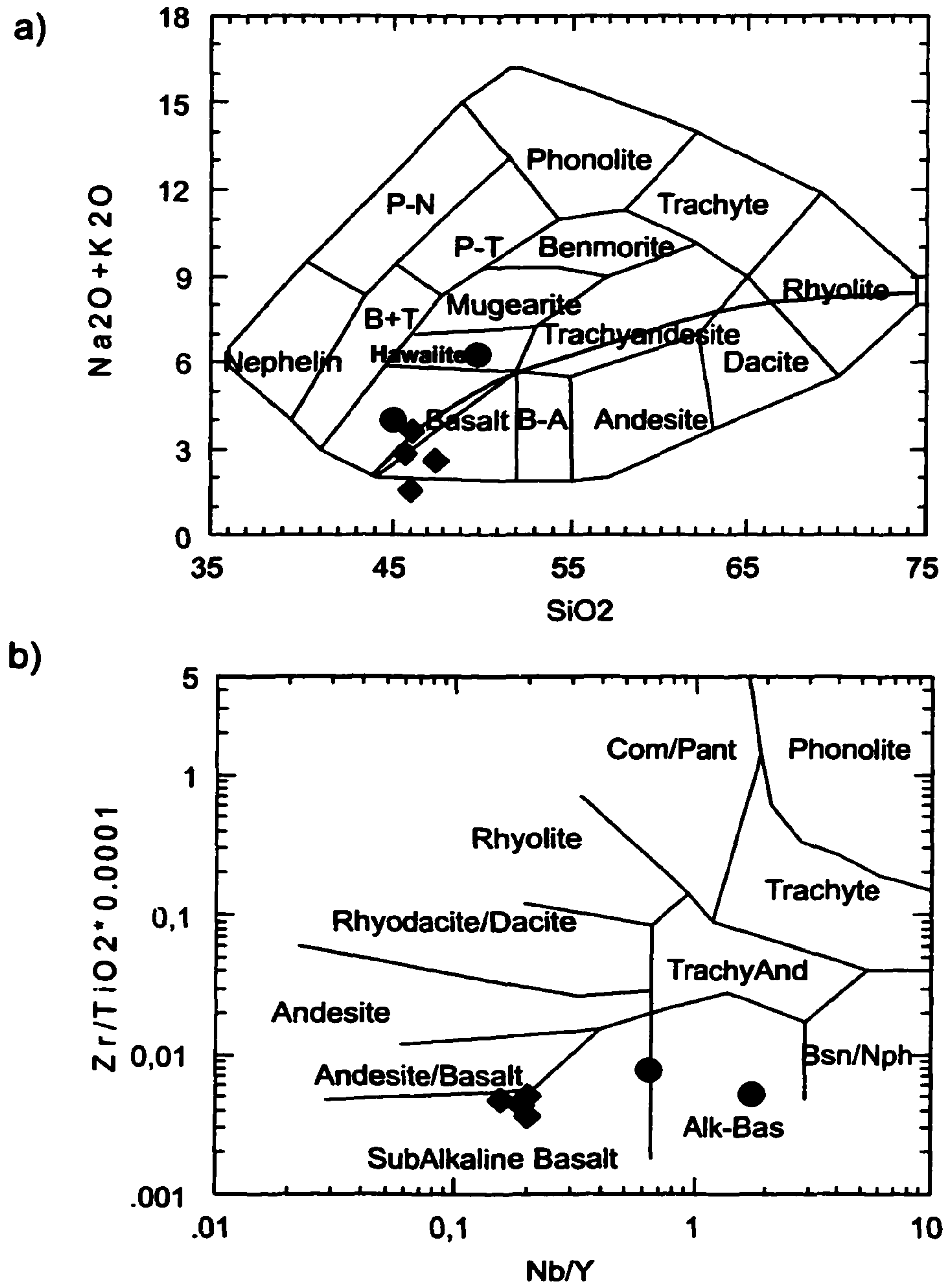


Figure 3.1 Metavolcanic rocks of the Cosoltepec Formation classify as alkaline basalts (black circles) and subalkaline basalts (black diamonds). (a) Na2O + K2O Vs SiO2, after Cox et al. (1979). (b) Zr/TiO2 - Nb/Y, after Winchester and Floyd (1977).

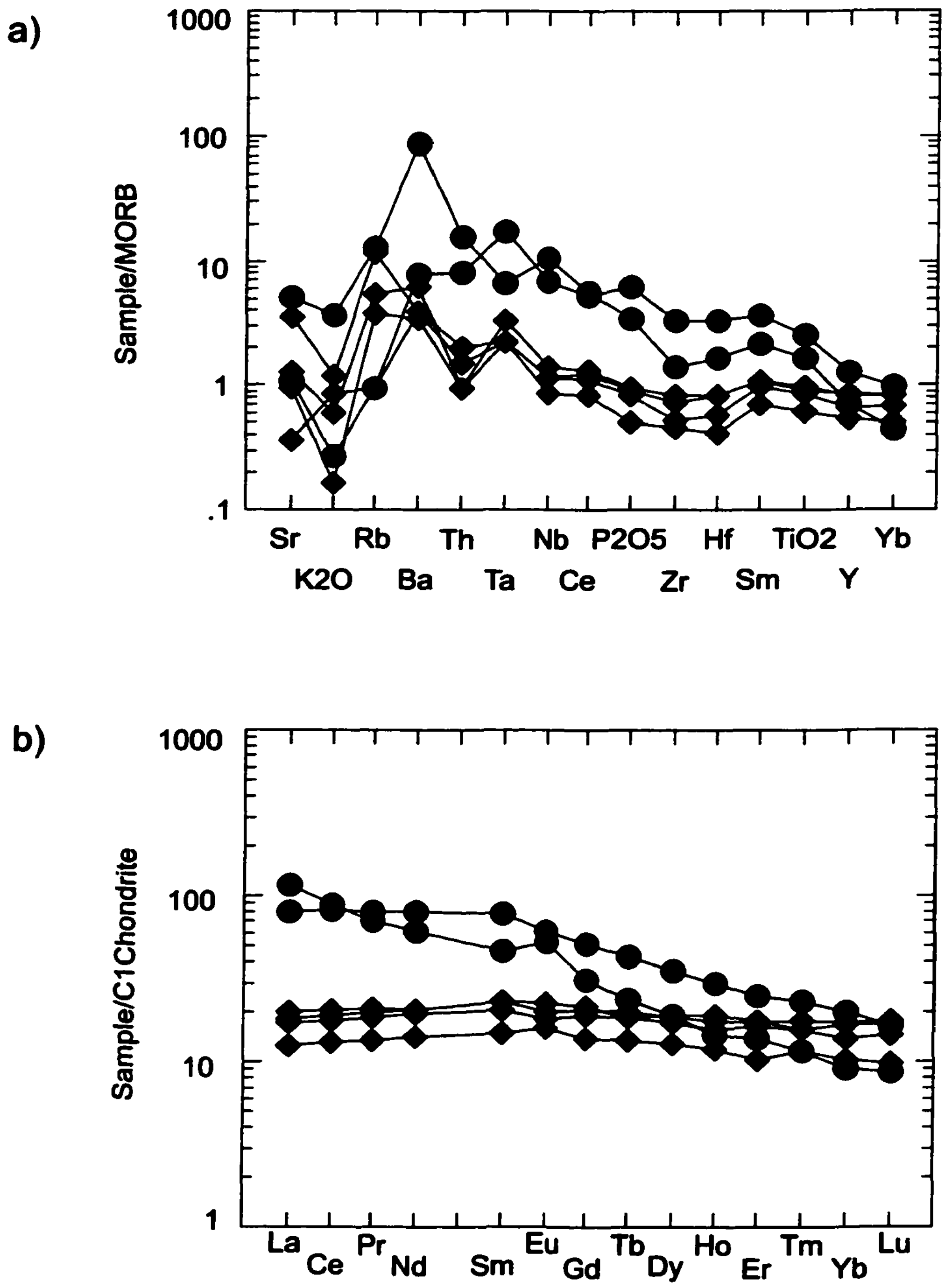


Figure 3. 2. (a) Multi-element diagram, and (b) REE patterns of metavolcanic rocks of the Cosoltepec Formation normalized to MORB and Chondrite values suggested by Pearce (1983) and Evensen et al., (1978) respectively. Symbols as in figure 3. 1

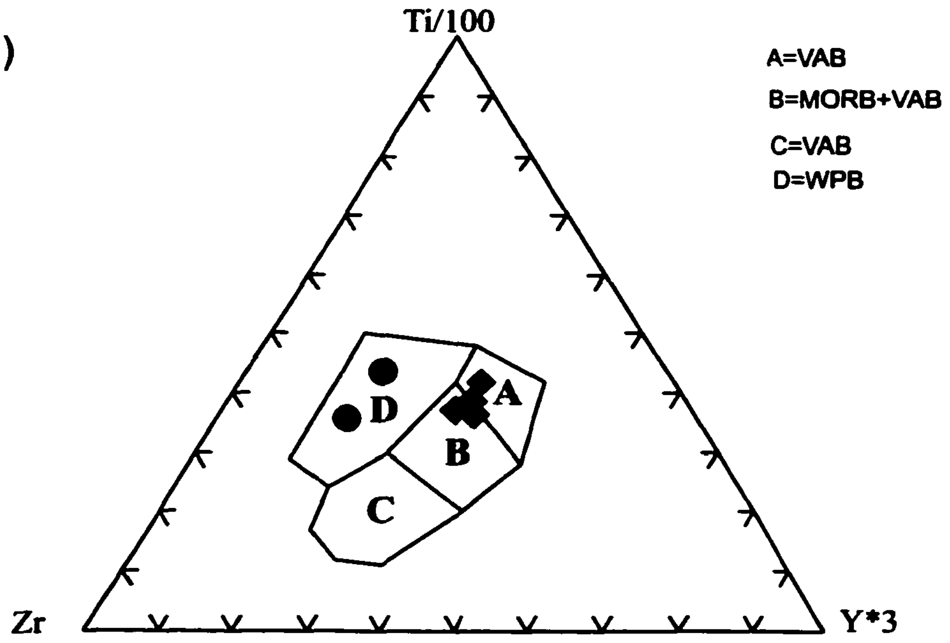
effect of an orogenic component in the magma genesis (Pearce and Norry, 1979). Such features have been described in many present-day back-arc basin settings where a mantle source has been modified by fluids coming from a neighboring arc (Tarney et al., 1977). However, because of the low number of analyzed samples, a back-arc basin setting is not a firm conclusion. Chondrite-normalized REE patterns are almost flat, showing slight depletion or slight enrichment in LREE relative to HREE (Figure 3. 2b). $(La/Yb)_N$ and Y/Nb ratios ranging from 0.99 to 1.23 and 2.63 to 2.88 respectively are indicative of typical tholeiitic series (Pearce and Cann, 1973).

According to the Ti-Zr-Y discriminant diagram proposed by Pearce and Cann (1973), all Ti-rich samples clearly fall within the OIB field, whereas tholeiites with low TiO_2 concentrations straddle the MORB and IAT (Island Arc Tholeiite) fields (Figure 3. 3a). However, using the Hf-Th-Nb discriminant diagram of Wood (1980) to better differentiate volcanic arc from ocean ridge settings, samples exclusively plot in the MORB and OIB fields, arguing against an IAT source (Figure 3. 3b).

The indisputable presence of MORB rocks plus the existence of OIB-type samples, strongly suggest an oceanic basin as the most appropriate tectonic setting for the metavolcanic rocks of the Cosoltepec Formation.

In conclusion, the lithological and geochemical characteristics of the Cosoltepec Formation, along with those geological features reported for the whole Petlalcingo Group (Magdalena and Chazumba Formations), suggest that sediments from this group were derived from a Proterozoic continental margin and deposited over an ancient oceanic floor in a very distal passive tectonic setting. The ancient oceanic crust could be part of either the Iapetus or Rheic oceans.

a)



b)

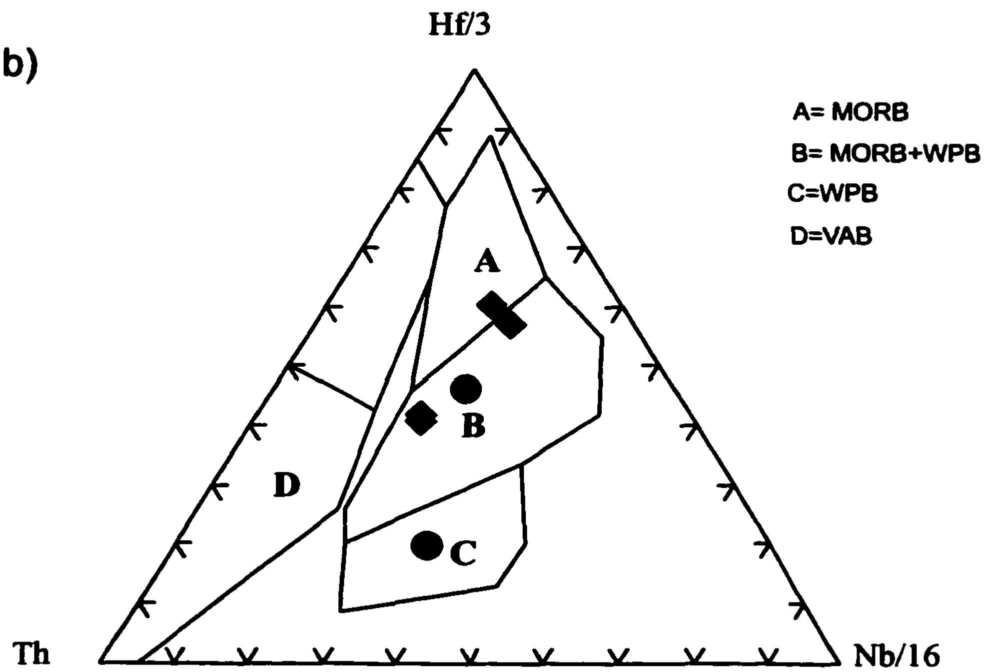


Figure 3. 3 (a) Ti-Zr-Y discriminant diagram after Pearce and Cann, (1973), alkaline basalts plot in the OIB field. (b) Hf-Th-Nb discriminant diagram after Wood (1980), subalkaline basalt plot in the MORB field. Symbols as in Figure 3.1

3. 3. Tecomate Formation

Petrography of igneous rocks

Igneous rocks of the Tecomate Formation comprise a bimodal assemblage with dominant mafic lavas (basalts and gabbros) and scarce felsic lavas (rhyolites). These rocks have been severely deformed and metamorphosed under greenschist facies conditions and primary textures and mineralogy have been completely replaced by metamorphic fabrics and phases. Exceptions are some microgabbroic dykes showing relict porphyritic to doleritic textures displaying clinopyroxene and amphibole phenocrysts.

Mafic rocks show a well-developed foliation composed of alternating actinolitic amphibole-rich and plagioclase-rich bands. Chlorite, sericite, microgranular epidote and oxides are the typical accessory minerals. Quartz is only present in a few samples and normally appears associated with plagioclase, sericite and chlorite.

Microgabbros exhibit completely altered plagioclase and brown amphibole pseudomorphs after clinopyroxene. Actinolitic amphibole develops as discontinuous coronas around clinopyroxene and brown amphiboles. Locally, greenish amphibole defines the foliation surface. Chlorite is present in all samples as an interstitial phase, whereas serpentine partly or completely replaces clinopyroxene and amphibole crystals in some samples.

Felsic rocks show relict microporphyritic and microgranular textures, however, in most samples, deformation is evident by the development of shear zones, incipient schistosity, crenulation and undulating surfaces defined by sericite and quartz ribbons. Felsic rocks consist of quartz, rare plagioclase and oxide microphenocrysts mostly enclosed in a microgranular groundmass of recrystallized quartz, plagioclase, k-feldspar and rare oxides. Metamorphism can be recognized by the development of sutured to lobated surfaces in quartz aggregates.

Whole-rock Geochemistry

Whole-rock geochemical data of igneous rocks of the Tecomate Formation (major, REE, trace element concentrations) are reported in Table 3. 2, whereas Nd data are shown in Table 3.4. On the other hand, locations of samples are shown in Figures 2. 2, and 2. 3.

Most samples display concentrations lower than 51.5 wt % indicative of basaltic rocks, although some samples show values higher than 68 wt% suggesting a bimodal suite, with a significant gap between 52 to 68 wt% of SiO_2. According to the TAS diagram of Cox et al. (1979), mafic rocks are classified as basalts overlapping the alkaline and subalkaline fields (Figure 3. 4a). However, sub-alkaline rocks defined by the Zr/Ti vs Nb/Y diagram of Winchester and Floyd (1977) plot in the basalt-andesite field, displaying an apparent transition between these and the alkaline basalts. On the other hand, felsic rocks are classified as rhyodacites to dacites using the same diagram (Figure 3. 4b).

Whole-rock geochemistry indicates that mafic rocks of the Tecomate Formation define a transition between two end-members, hereafter called groups A and B. Rocks of group A are found at the bottom of the volcanic sequence, and are characterized by TiO_2 concentrations lower than 1.49 wt%, whereas rocks of the upper group B display TiO_2 concentrations higher than 2.14 wt%. Some samples with concentration at about1.8 wt % could represent the transition or evolution between both end-members.

At present, the lower group A is represented by 9 samples including 5 basalts, 1 gabbro and 3 felsic rocks which emphasizes its bimodal character. Mafic rocks display strong variations in TiO_2 content ranging from very low (0.26 wt%) to low values (1.49 %). This group also displays low concentrations in Nb (2 to 7ppm), Zr (30 to 125 ppm) and Y (11 to 37 ppm). The low Nb/Y ratios ranging from 0.06 to 0.2 indicate the tholeiitic nature of these rocks. The low concentrations in the HFS elements Ti, Nb, Zr, and Y have

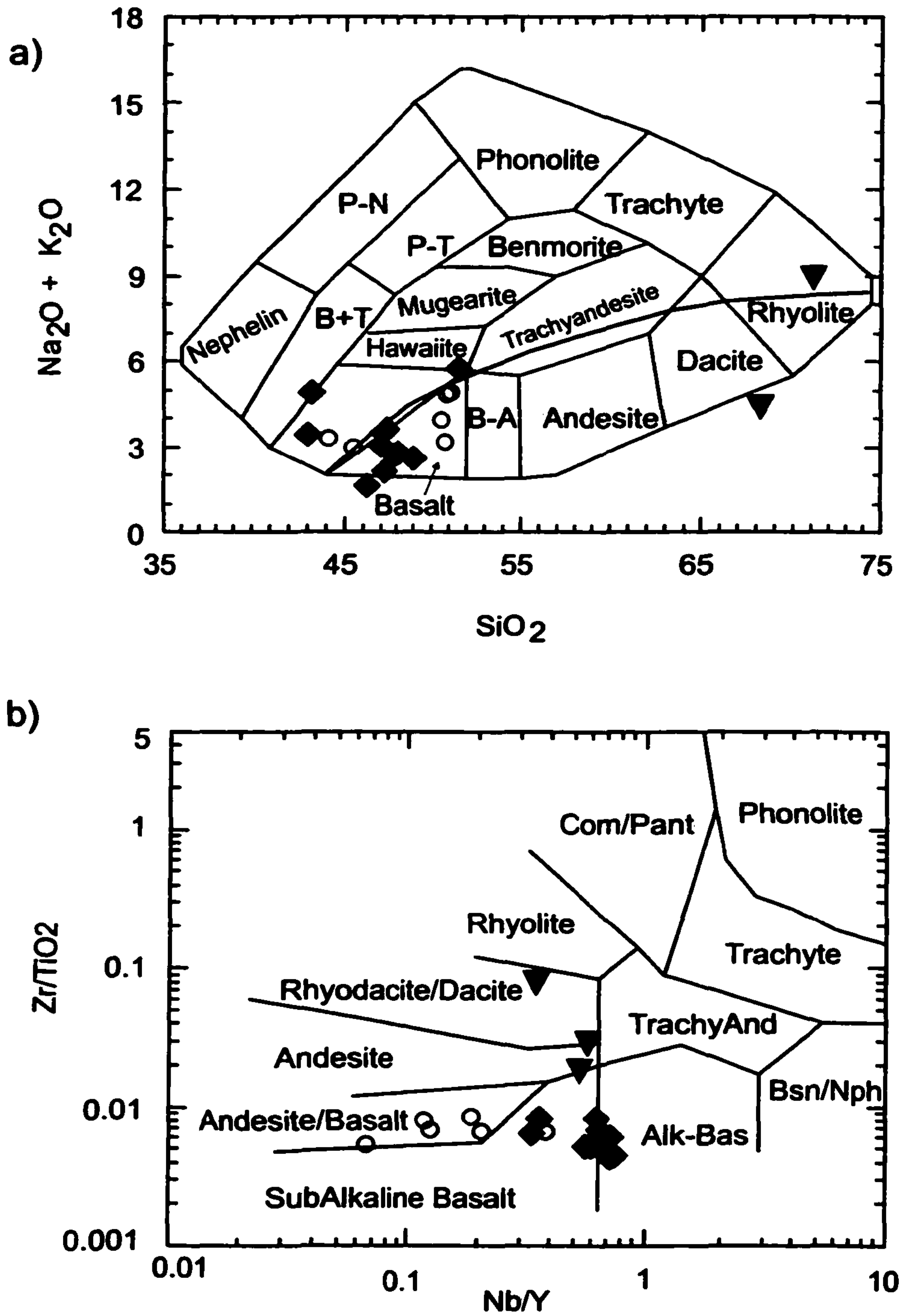


Figure 3. 4. Igneous samples of the Tecomate Formation classify as subalkaline to transitional and alkaline rocks. (a) Na2O+K2O, after Cox et al., (1979). (b) Zr/TiO_2 - Nb/Y, after Winchester and Floyd, (1977). Mafic rocks of group A (open circles), mafic rocks of group B (Black diamonds). Felsic rocks (black triangles).

been reported in most island arc tholeiites (IAT) (Pearce and Cann, 1973; Pearce and Norry, 1979). According to the Mg # these tholeiitic rocks derive from low (Mg # from 51 to 61) to mildly fractionated magmas (from 43 to 37). Ni and Cr exhibit moderate to high concentrations according to the TiO_2 and Mg # values. On the other hand, felsic rocks also display low concentrations in TiO_2 (from 0.26 to 0.85 wt%), and Mg # varies from 18.5 to 22.1 suggesting strong fractionation.

Multi-element patterns for basalts and rhyolites of suite A were normalized to MORB and ORG using values of Pearce (1983), and Pearce et al. (1984) respectively. In general, mafic rocks display high enrichment in LFSE with constant peak values in Rb and Ba, whereas HFSE show almost flat patterns with significant negative anomalies in the "orogenic trace elements" Nb, Zr and Ti, similar to those patterns reported in IAT settings (Figure 3. 5a). The same depletion in Nb and Zr characterizes the multi-element patterns of rhyolites (Figure 3. 5b).

In contrast, REE patterns in mafic rocks vary from almost flat to slightly enriched in LREE relative to HREE (Figure 3. 5c). La_N and Yb_N values range from 11 to 53 and 6 to 25 times-chondrite respectively. $(La/Yb)_N$ ratios bracketing in between 1.5 to 2.1 along with the very low Nb/Y ratios emphasize the tholeiitic nature of these rocks. REE patterns in felsic rocks are more homogeneous, with LREE highly enriched relative to HREE and La_N and Yb_N values ranging from 90 to 300 and 15 to 40 times-chondrite respectively (Figure 3. 5d). In addition, $(La/Yb)_N$ values range from 4.6 to 6.9. Small negative Eu anomalies are present only in the most fractionated samples.

According to the Ti-Zr -Y discriminant diagram of Pearce and Cann (1973), mafic rocks of group A consistently plot in the MORB + VAB field (Figure 3. 7a). However, such a diagram is the most appropriate to identify within plate basalts (WPB), but not to discriminate between MORB and VAB settings (Pearce, 1996). Therefore, using the Hf-

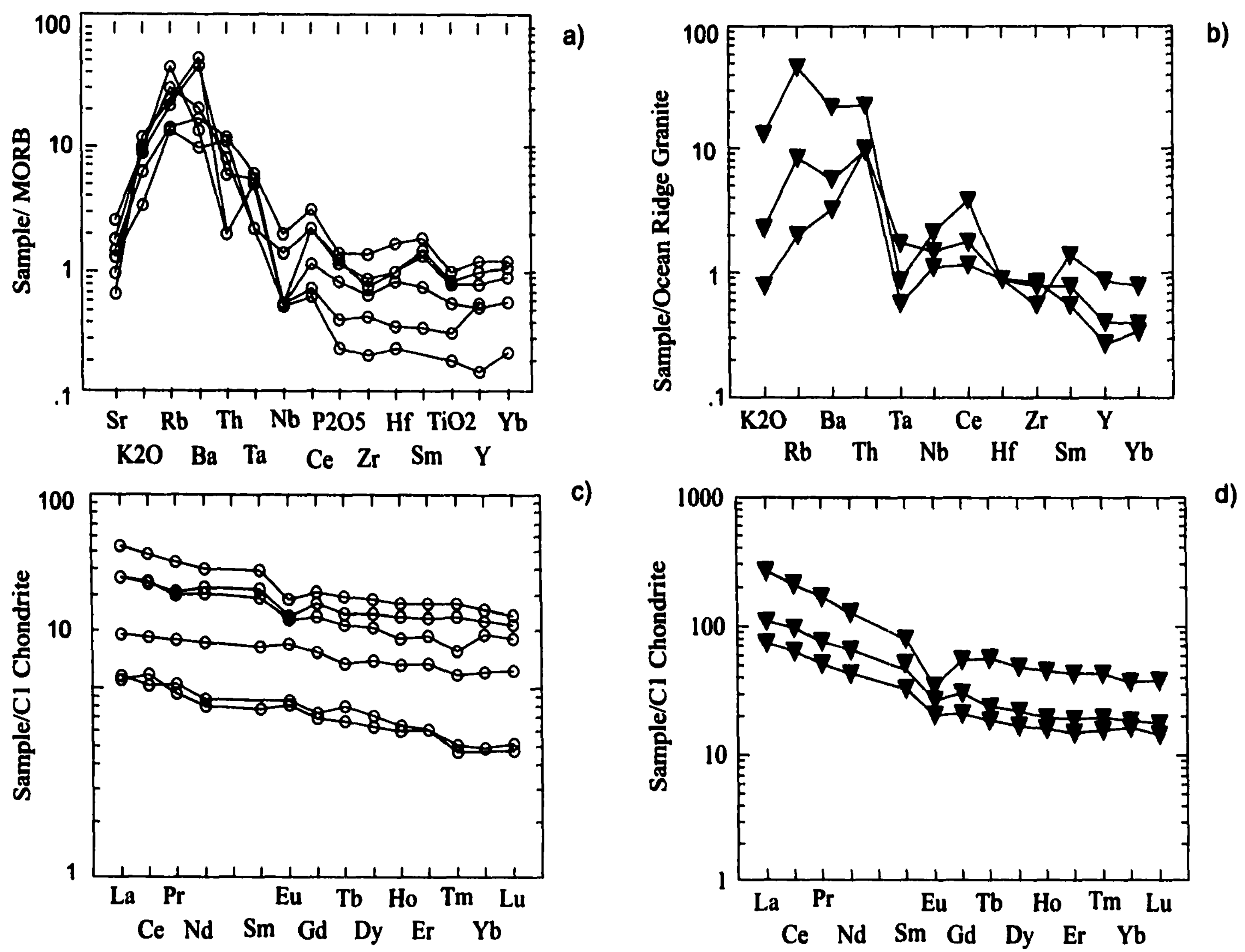


Figure 3.5 (a-b) Multi-element diagrams and (c-d) REE patterns of mafic and felsic rocks from Group A of the Tecomate Formation normalized to MORB, ORG and Chondrite values of Pearce, (1983), Pearce et al., (1984) and Evensen et al., (1978) respectively. Symbols as in Figure 3.4

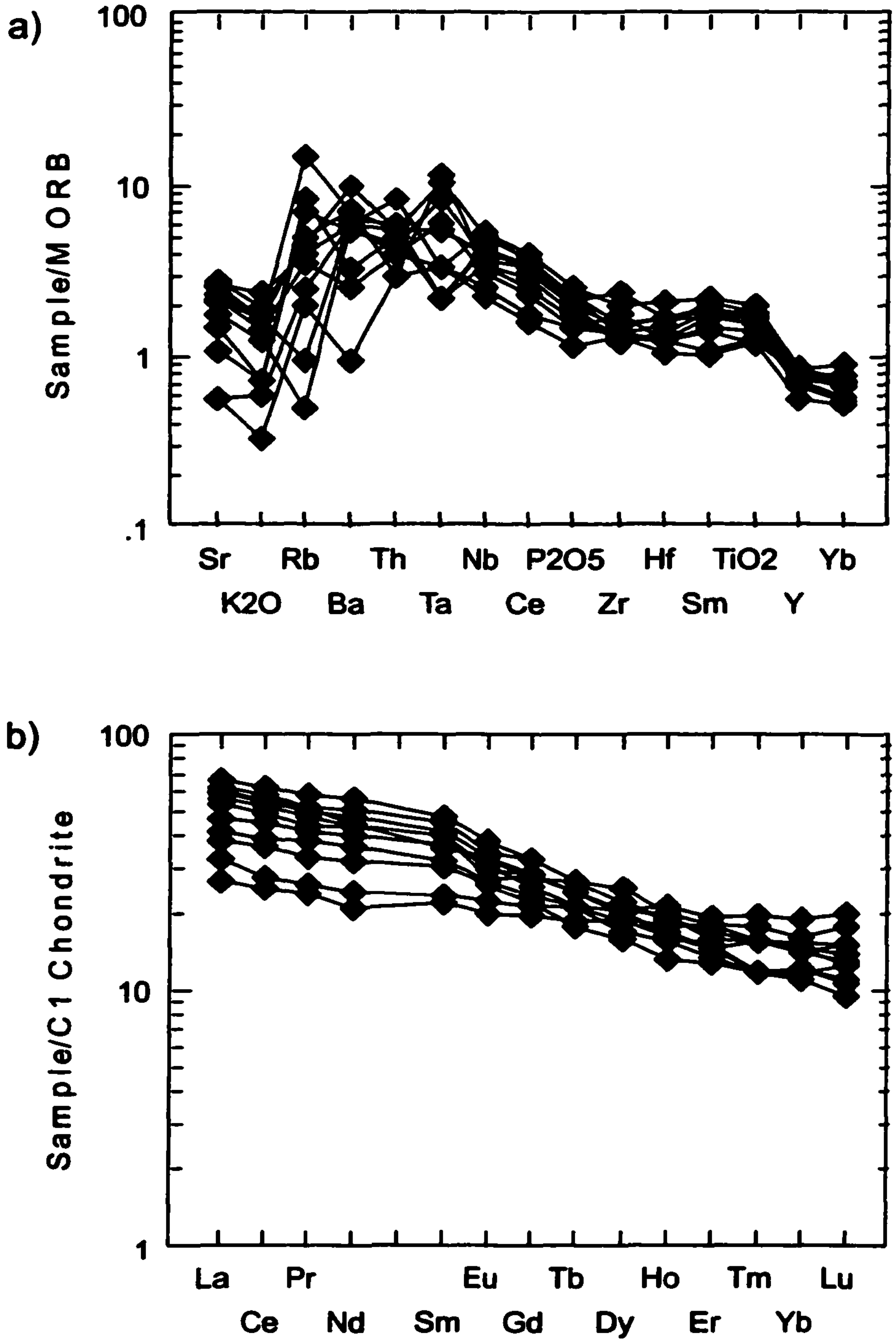


Figure 3. 6 (a) Multi-element diagram and (b) REE pattern of mafic rocks from Group B of the Tecomate Formation normalized to MORB and Chondrites values of Pearce (1983), and Evensen et al., (1978) respectively. Symbols as in Figure 3. 4.

Th-Nb discriminant diagram of Wood (1980) to better discriminate VAB setting from MORB, all mafic samples clearly plot in the VAB field (Figure 3. 7b). Despite the tholeiitic character of group A all samples plot in the calc-alkaline volcanic arc field. The reason for that could be the high fractionation of these rocks (low Mg #), which increases the Th abundance.

Taking into account the overall geochemical characteristics of group A (REE patterns, multi-element and discriminant diagrams), the mafic rocks seem to represent a volcanic arc setting. However, considering the geologic framework of the Tecomate Formation along with the geochemistry of group B, such a magmatic arc signature seems to represent an inherited feature in a post-collisional setting (Pearce et al., 1990).

Igneous rocks of group B are represented by eight volcanic and three intrusive mafic samples, with SiO_2 concentrations varying in a small range from 43.3 % to 51.5 %. TiO_2 concentrations are relatively high, ranging from 2.1 wt% to 2.98 wt%. However, there are some samples with an average of 1.8 wt% representing the transition between groups A and B that were included in the latter group. Using the Zr/Ti vs Nb/Y diagram of Winchester and Floyd (1977) to identify rock type, group B clearly plots in the alkali-basalt field (Figure 3. 4b). Nb/Y ratios vary in a small range between 0.56 to 0.75 suggesting the transitional tholeiitic-alkaline character of these rocks (Pearce and Cann, 1973). Mg # values range from 37 to 24, coincident with the low concentration reported in MgO, Ni, and Cr indicative of strongly fractionated magmas.

Multi-element patterns of group B, normalized to MORB using values of Pearce (1983), are quite homogeneous (Figure 3. 6a). The most mobile elements included in the LFS group (Rb, Ba, K, Sr) present extreme variations, possibly as a consequence of the metamorphic event that affected the Tecomate Formation. On the other hand, all samples show very similar HFSE spectra characterized by moderate to high enrichment in Nb, Zr,

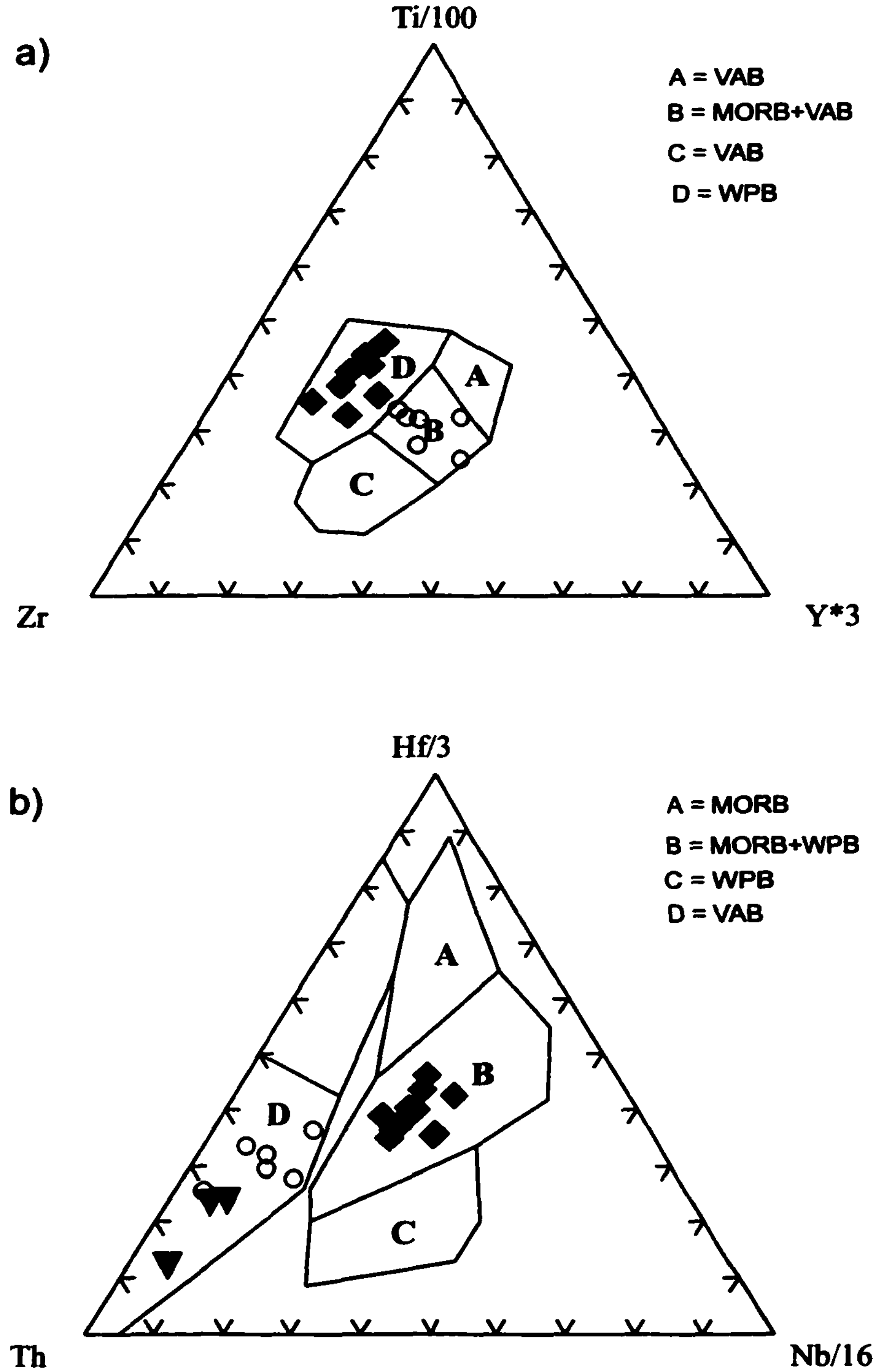


Figure 3.7 Samples of the Tecomate Formation, (a) Ti-Zr-Y discriminant diagram after Pearce and Cann, (1973), mafic rocks of group B plot in the WPB field. (b) Hf-Th-Nb discriminant diagram after Wood (1980), mafic and felsic rocks of group A plot in the VAB field. Symbols as in Figure 3.4.

Hf, and TiO_2, whereas the least incompatible elements Y and Yb are highly depleted.

The described patterns show the characteristic "humped" shape typical of within plate setting. Most samples also exhibit high Ti/Y ratios ranging from 5.3 to 7.0, similar to those values of within plate volcanic rocks. Depletion in Y and Yb could indicate the presence of residual garnet in the mantle source in addition to low degree of melting. Enrichment in most HFSE (Ti, Zr, and Nb) could also suggest either a deeper and enriched mantle source or the influence of sub-continental lithosphere (Pearce, 1983).

On the other hand, REE patterns are also very homogeneous and most samples show an important enrichment in LREE relative to HREE (Figure 3. 6b). Concentration in La varies from 38.4 to 66.2, whereas Yb content ranges from 10.9 to 15.1 times-chondrite respectively depending on the extent of fractionation (Evensen et al., 1978). This group has an average $(La/Yb)_N$ ratio of 4.0, characteristic of transitional to alkaline magmas in within plate settings.

In agreement with the described multi-element diagrams and REE patterns, mafic rocks of group B consistently cluster in the within plate field when using the Ti-Zr-Y, and Hf-Th-Nb tectonomagmatic discriminant diagrams of Pearce and Cann (1973) and Wood (1980) respectively (Figures 3. 7a-b). In conclusion, basalts from group B clearly represent a magmatic event produced by an extensional process in an intraplate setting.

Taking into account the whole geochemistry shown by all mafic rocks included in the Tecomate Formation, the transition displayed by all samples plotted in the Ti/100 vs V discriminant diagram of Shervais (1982) is noteworthy (Figure 3. 8). It seems that the lower group A displays an inherited VAB signature, which evolves during the tectonic evolution of the Tecomate Formation to the alkaline magmatism of group B typical of within plate settings. These data are coincident with the post-tectonic role played by the Tecomate Formation after the Silurian orogeny.

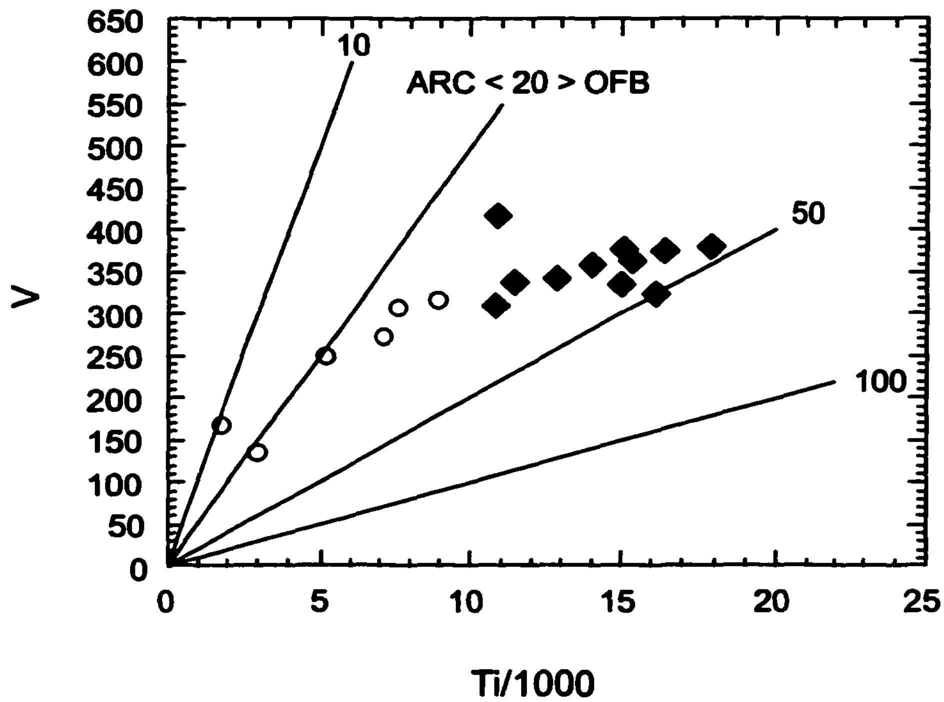


Figure 3. 8 Volcanic samples of the Tecomate Formation display an apparent evolution fom mafic rocks of group A with VAB signature to alkaline mafic rocks of group B with WPB signature, plotted in the Ti-V discriminant diagram of Shervais, (1982).

Nd-Isotope and Crustal Residence age

Nd-isotope analyses of the Tecomate Formation were carried out on six samples from group A and five samples from group B. As a whole, lavas from this formation display strong variation in initial ε_{Nd} values (assuming an absolute age of 400 Ma), ranging from moderatly positive to highly negative values (+ 5.58 to - 6.16) and present-day ε_{Nd} values range from + 3.48 to - 10.04 (Table 3. 4).

Group A present mainly negative values with $\varepsilon_{Nd(400\ Ma)}$ of basaltic rocks ranging from + 1.53 to - 6.16 and present-day values from +0.10 to -10.04, whereas rhyolites vary from -4.37 to 5.72 and - 7.8 to - 9.07 respectively. Similar values are common in continental margins (VAB setting), where mantle sources have been contaminated by sediments of sialic composition or directly by continental crust. Similar values are also reported in continental WP settings, where mantle sources could be contaminated either within the mantle or by subsequent assimilation of continental crust. Depleted mantle model ages $T_{(DM)}$ in group A are consistently high, ranging from 1239 to 1600 Ma., which suggests the strong contamination of this magmas with a very old crustal material.

On the other hand, samples from group B display very homogeneous ε_{Nd} values (assuming 400 Ma), ranging from + 4.22 to + 5.58, and present-day values ranging from + 2.02 to + 3.48. The small variation in ε_{Nd}, is supported by the reduced range of $^{147}Sm/^{144}Nd$ ratios present in this group. These values are similar to those isotopic compositions recorded in either IAB or WPB, which are lower than those registered in MORB settings (Whitford et al., 1981; White and Hofmann, 1982; Carlson et al., 1981). However, assuming that group B represents magmas erupted in an intraplate environment (WPB), Nd composition could be the consequence of magmas mainly derived from enriched sources in the mantle (EM) and low crustal contamination. Depleted mantle model ages $T_{(DM)}$ in group B are also more homogeneous, ranging consistently from 622 to 781 Ma, supporting the evidence of low extent of contamination.

Considering the reported ε_{Nd} data, samples from the Tecomate Formation graphically cluster into two different groups suggesting the contribution of different mantle source and extent of assimilation of crustal materials. According to the ε_{Nd} vs Ti/Yb diagram of Wilson (1994), basalts of group B plot towards the enriched source mantle direction, whereas basalts and rhyolites of group A plot towards the crustal contaminaton trend (Figure 3. 9).

Summarizing the available information of the Tecomate Formation, it is conclusive that its whole-rock geochemistry indicates a bimodal magmatic suite ranging from tholeiitic to transitional-alkaline composition. Mafic rocks are subdivided into groups A and B, the former displaying the influence of subduction-arc setting and the latter clearly representing intraplate volcanism with transitional-alkaline affinity. Notable is the evolutionary trend displayed by both magmatic groups as shown in the Ti - V diagram of Shervais (1982), which suggests a more complex tectonic setting varing from an inherited subduction related settings (VAG) to intraplate environmment (WPB) (Figure 3. 8).

Magmatic associations showing geochemical signatures of VAB to WPB transitions have been described in syn-and post-collisonal events (Pearce et al., 1990). Taking into account the post-orogenic role played by the Tecomate Formation, in addition to the transitional-alkaline composition of magmas from group B, the most appropriate setting for this formation is a post-collisional event where magmatism occurred during an extensional process in a zone with previous subduction history. The extensive magma-crust interaction reflected by melting of mantle beneath thickened crust in a post-collision setting is recorded by the low positive and high negative ε_{Nd} values of group A.

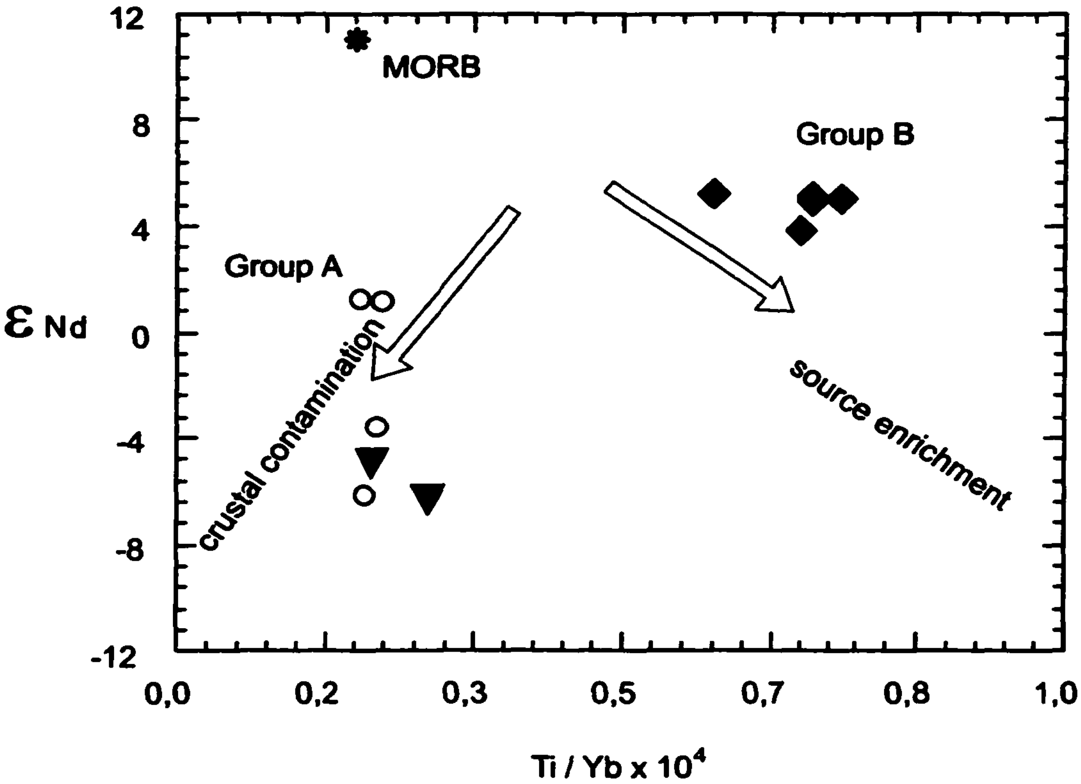


Figure 3. 9 e Nd - Ti/Yb diagram after Wilson (1994). Group A displays mainly negative values as a consequence of crusta contamination, while rocks of group B closter toward values representing an enriched mantle source. Symbols as in Figure 3. 4.

3. 4 Esperanza, Hornos and Teticic Granites

Petrography of granitoids

Based on the available isotopic data, in addition to the stratigraphic relationships described in Chapter II, the description of the different granitic suites is from the older to the younger intrusives: Esperanza, Hornos and Teticic granitoids of Late Ordovician-Silurian, Devonian and Carboniferous ages respectively.

Esperanza granitoids. Esperanza-type plutons are highly deformed displaying schistose and gneissic foliations, and all samples exhibit large variations in cataclastic textures ranging from protomylonites through mylonites to ultramylonites. As a whole, this group is clearly distinguished by the mineralogical association of biotite, muscovite and garnet. A very conspicuous and distinctive lithology within these granitoids is the augen-schistose phase, characterized by a porphyroblastic fabric where porphyroblasts, from 2 to 8 cm long, of pink K-feldspar (microcline) are enclosed in a fine, dark and spaced foliated matrix. Most porphyroblasts are aligned parallel to the foliation surfaces, but many of them have been rotated showing sigmoidal shapes. The parallel to anastomosed foliation exhibits thin bands and lenses of polycrystalline quartz, K-feldspar, plagioclase and isolated crystals of garnet, alternating with layers rich in muscovite and biotite. Chlorite is a very common alteration of garnet and biotite. In some petrographic descriptions of the type locality (km 261 along the Mexico City - Oaxaca highway), very scarce crystals of orthopyroxene have been reported (Fries and Rincón-Orta, 1965 and Ruiz-Castellanos, 1979).

The most widely distributed lithology is a mica-schistose phase displaying a planar anisotropic fabric, defined by bands of quartz, K-feldspar, and plagioclase alternating with thin, planar to anastomosed muscovite-rich bands. Even though garnet is an accessory mineral, it is widely distributed. Less abundant within the Esperanza granitoids is a quartzo-feldspathic gneiss that is partially migmatized where the mylonitic foliation is composed of alternating layers of recrystallized feldspar, muscovite and quartz ribbons. An even more rare phase within this group of granitoids corresponds to a more mafic group of rocks, including metadiorites and metagabbros. This phase is represented by 3 reported samples (139, 251 and 245) which display similar mylonitic deformation, although granular textures are also present. One sample shows an intrusive relationship with the augen-schistose phase, but presents the same mineralogical association of biotite-muscovite-garnet. However, 2 samples (251 and 245) differ mineralogically from the typical Esperanza granitoids in the presence of amphibole. Even though the relationship of this mafic phase within the Esperanza granitoids is not clear, it was considered as part of the unit based on the structural features.

Hornos-La Noria granites. These granites are characterized by a very distinctive megacrystic texture where large pink K-feldspar crystals are enclosed in a coarse-grained matrix made of quartz with undulous extinction, plagioclase, K-feldspar and biotite. Chlorite is present as a secondary mineral replacing biotite, whereas sericite replaces plagioclase and both minerals are well developed along shear planes and foliation surfaces. Scarce crystals of hornblende are also present (Figure 2. 15b).

These granitic bodies show different degrees of cataclasis, ranging from megacrystic texture to mylonitic foliations. An augen schistose fabric similar to that present in the Esperanza granitoids is also well developed in the Hornos granites, however, the latter differ mineralogically from the former in the absence of garnet (Figure 2. 16a). In addition, the presence of numerous enclaves within this type of granite is a noteworthy feature (Figure 2. 16b).

Teticic granites. This group includes microgranites, meta-aplites and muscovite schists. The very felsic Teticic granites also present all variations of cataclastic textures ranging from fine grained fabrics to mylonitic foliations, including foliated microaugen schist with white K-feldspar and quartz as the main porphyroids. Schists are characterized by a well-developed planar foliation defined by a granoblastic fabric. Microgranites display microgranular textures, ranging from relict heterogranular to micro-porphyroblastic. Mineralogically, this group is quite homogeneous including quartz, K-feldspar, plagioclase, muscovite, and chlorite. Dark minerals are rare (< 7%) indicating their leucocratic nature. Mineralogically, the Totoltepec pluton differs from this group of granites in the presence of biotite (altered to chlorite) that defines the mafic phase, as well as the abundance of plagioclase as the essential mineral, however the Totoltepec granite was also included within the Teticic group.

Whole-rock Geochemistry of granitoids

The following description is based on major elements. The analyzed samples from the Esperanza granitoids are predominantly felsic (SiO_2 = 66.5 to 73.5 wt.%), but minor plutons range in composition from mafic to intermediate (SiO_2= 46.7 to 56.4 wt.%). As a whole, this group displays an important gap in silica content from 57 to 66 wt. %. On the other hand, the Hornos and Teticic granites are more homogeneous with a silica average of 69 wt% in the former to very high values ranging from 77 to 81wt% in the latter.

According to the $Na_2O + K_2O$ vs SiO_2 diagram of Cox et al. (1979) adapted by Wilson (1989) for plutonic rocks, the overall samples fall in the sub-alkaline field (Figure 3. 10a). The felsic group of the Esperanza granitoids includes granite and granodiorite, whereas the mafic plutons vary from gabbro (samples 251 and 245) to diorite (sample 139). The Hornos and Teticic groups are mainly granites, but two samples of the latter group (samples 147 and 242) are trondhjemite based on major elements. On a normative diagram of Barker (1979), the Esperanza group mainly consists of granites, granodiorite

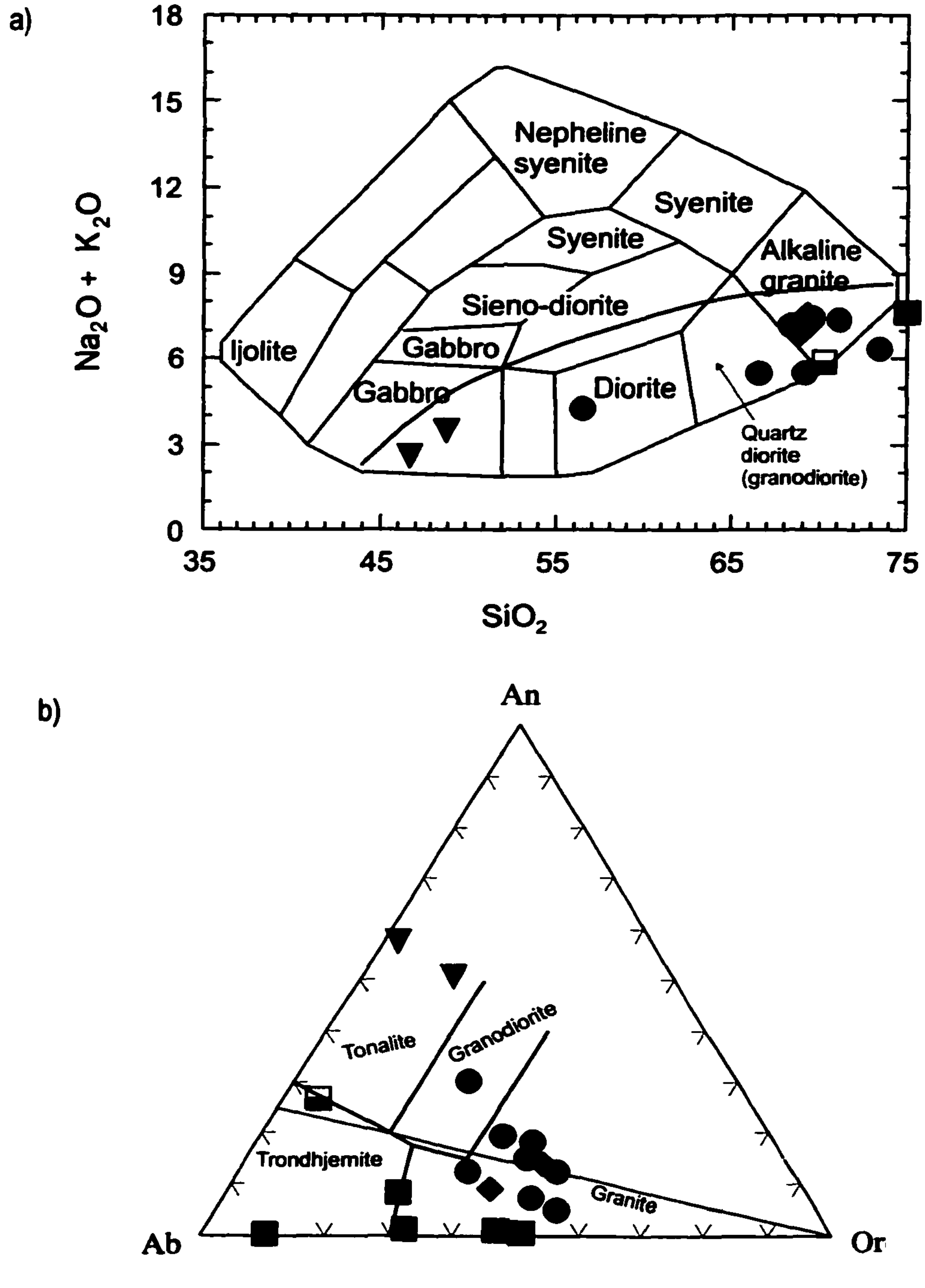
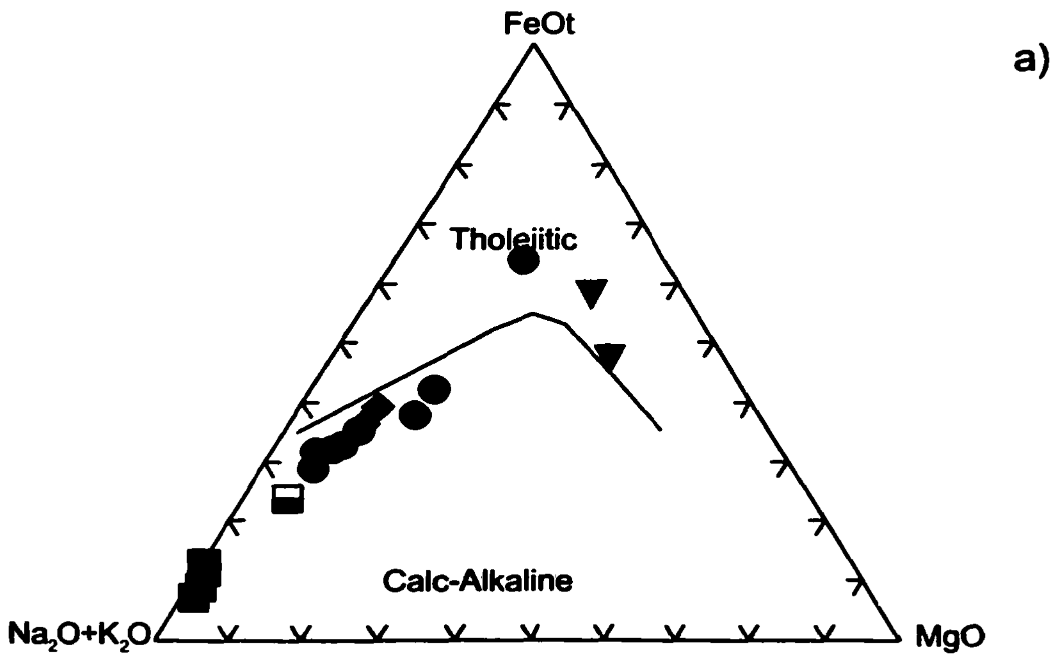


Figure 3. 10 (a) Na2O + K2O - SiO2 diagram after Wilson (1989), and (b) An-Ab-Or normative diagram after Barker (1979). Felsic Esperanza granitoids (black circles), Mafic Esperanza granitoids (black triangles), Hornos-La Noria granites (black diamonds), Teticic granites (black squares), Totoltepec granite (black and white squares).

(sample 139) and tonalites (samples 245 and 251). The Hornos group is represented by granitic rocks, whereas the Teticic group varies from granite to trondhjemite (Figure 3. 10b).

Even though the different groups of granitoids are not genetically related, they display an apparent fractionation trend through time (AFM diagram), ranging from more "mafic" granitoids in Silurian times (Esperanza group), to high-silica granites in Carboniferous times (Teticic group) (Figure 3. 11a). Based on the diagram proposed by Maniar and Piccoli (1984) to determine the Shand index, samples with SiO_2 higher than 56 wt.% range in molecular $Al_2O_3/(CaO + Na_2O + K_2O)$ from 1.08 to 1.44, classifying the overall rocks as peraluminous (Figure 3. 11b). According to the Shand index, only the most mafic samples (gabbros) show very low values (0.65 in average) indicating a metaluminous character (Shand, 1927). The wider variation (1.08 to 1.44) corresponds to the Esperanza granitoids, while the others range between 0.98 and 1.22. This factor was taken to indicate that almost all granitoids of the Acatlán Complex were produced by crustal anatexis during the continental collision of Laurentia and Gondwana (Ortega-Gutiérrez, 1978, 1993).

On the ACF diagram of Chappell and White (1992) which correlates chemical and mineralogical data with those granites derived from either, igneous or sedimentary source rocks (I and S type), most granitoids of the Esperanza group fall in the peraluminous field. This is coincident with the mineral association of muscovite, garnet and biotite, very characteristic of S-type or sediment-derived granitoids (Figure 3. 12a). Only the most mafic samples (gabbros) fall in the metaluminous field, therefore they are characterized by the presence of hornblende, which is a distinctive mineral of I-type plutons or mantle derived rocks. The intermediate rocks (diorite) fall in the peraluminous field where biotite characterizes both the S-type rocks and the most felsic I-type plutons (hornblende-free).



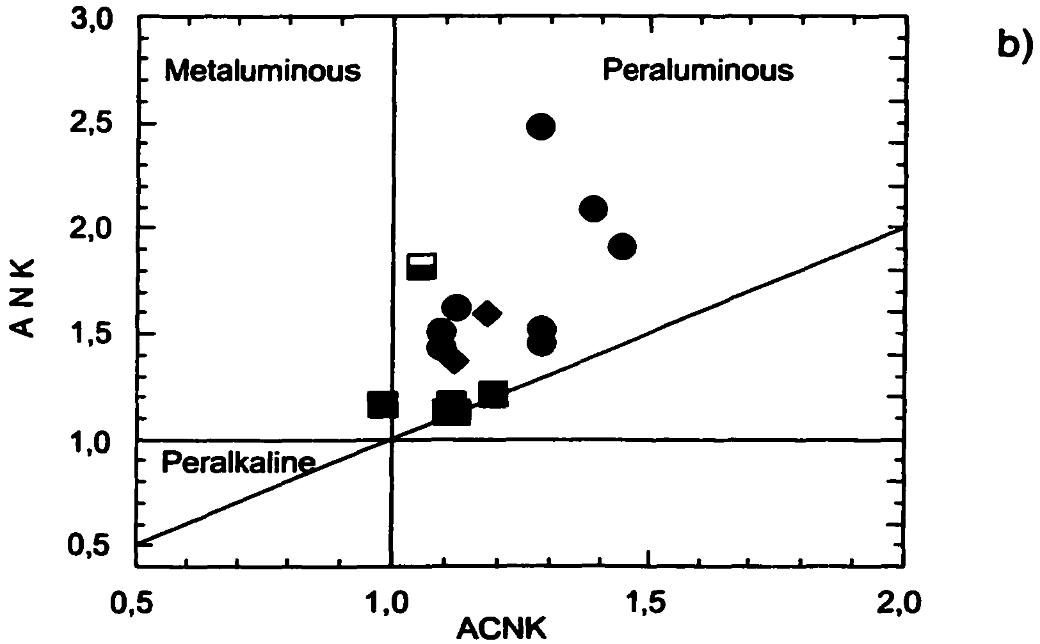


Figure 3.11 (a) AFM diagram of the Acatlán Complex granitoids displaying an apparent fractionation through time. (b) ANK-ACNKdiagram (after Maniar and Piccoli,1984) pointing out the main peraluminous character for the Acatlán Complex granitoids. Symbols as in Figure 3.10.

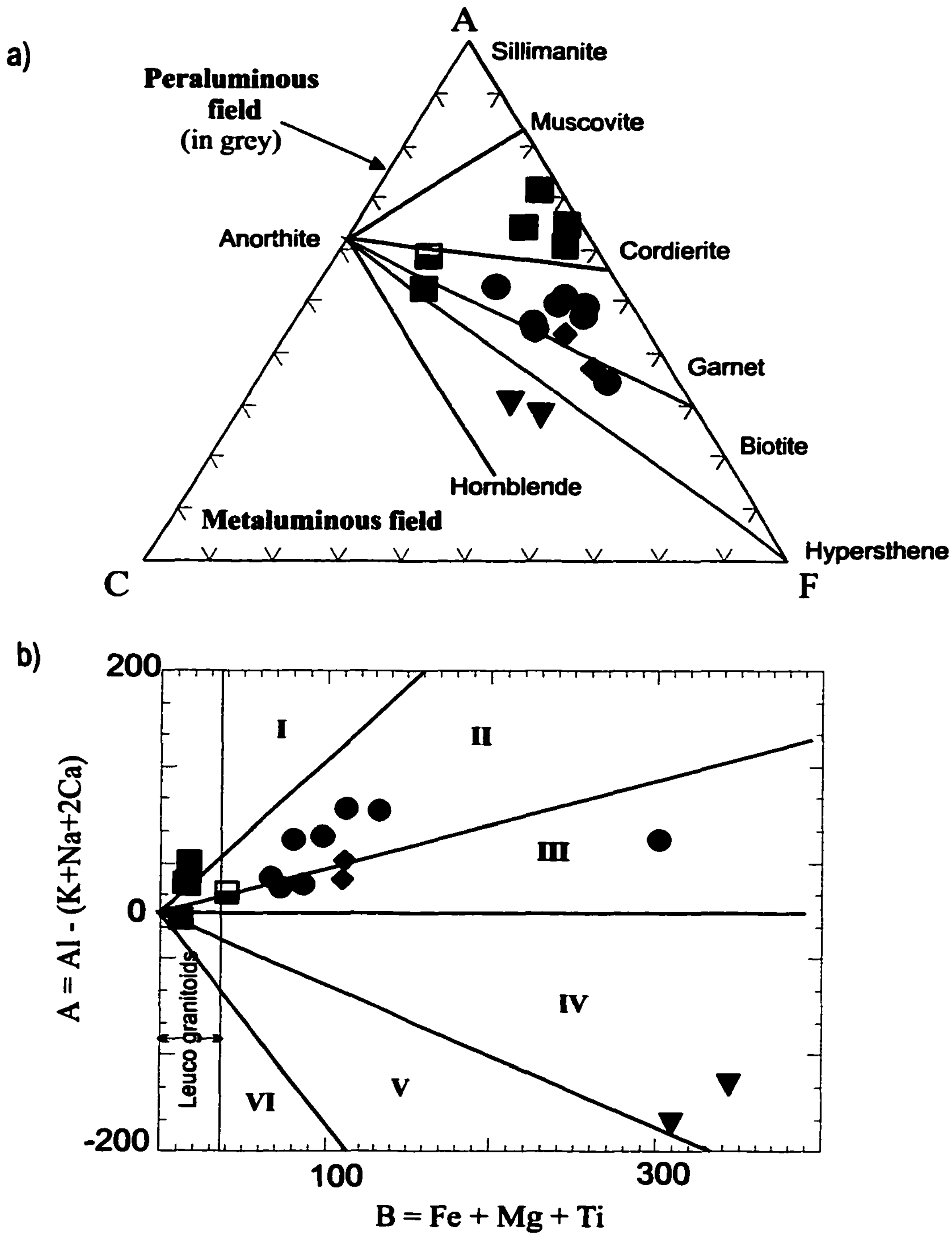


Figure 3. 12 a) ACF diagram after Chappell and White (1992). The grey area corresponds to peraluminous rocks composition. b) Chemical-Mineralogical diagram of Debon and Lefort, (1983) showing six sectors:I= Mu>Bi;II = Bi >Mu; III = Biotite; IV = Bi - Hb -Opx -Cpx; V = Cpx - Hb; VI = Cpx - Ep - Sph. Symbols as in Figure 3. 10.

In the same way, samples from the Hornos granites also fall in the peraluminous field where S-type and I-type fractionated rocks plot indistinctly. On the other side, Teticic granites also fall in the more peraluminous muscovite-cordierite field typical of fertile sediment source.

Similar information is derived from the AB diagram proposed by Debon and Le Fort (1983) to define chemical (peraluminous and metaluminous domains) and mineralogical characteristics, in addition to identify magmatic associations derived from the same process. According to this, the high-silica Teticic granites plot in the peraluminous leucocratic field with an almost vertical trend, whereas the felsic samples of the Esperanza granitoids display a rather positive slope within sector II of the peraluminous domain. Such tendencies are typical of aluminous associations where rocks are mainly or totally derived from the anatexis of sialic materials.

Only the mafic rocks of the Esperanza granitoids plot in sector IV, where rocks include mantle-derived magmas (Figure 3. 12b). Because of the existence of mafic samples and granites plotting close to the peraluminous and metaluminous limit, it seems conclusive that most Esperanza granitoids and Hornos granite are mainly derived from crustal material with little mantle contribution represented by the mafic intrusives. On the other hand, Teticic granites seem to be completely derived from anatexis of crustal rocks.

Analysis of rare-earth elements and selected trace elements from granitoids of the Acatlán Complex presents the following characteristics. Chondrite-normalized rare-earth-elements (REE) patterns for the more felsic Esperanza-type granitoids and samples of Los Hornos granites are illustrated in Figure 3. 13a. In general, these two groups of samples exhibit an overall enriched pattern ($\sum$ REE = 182-264), where the highly enriched LREE range from 136 to 187 and the HREE range from 20 to 29 times chondrite respectively. $(La/Yb)_N$ ratios for these two groups are relatively high, ranging from 5.9 to 8.14. In

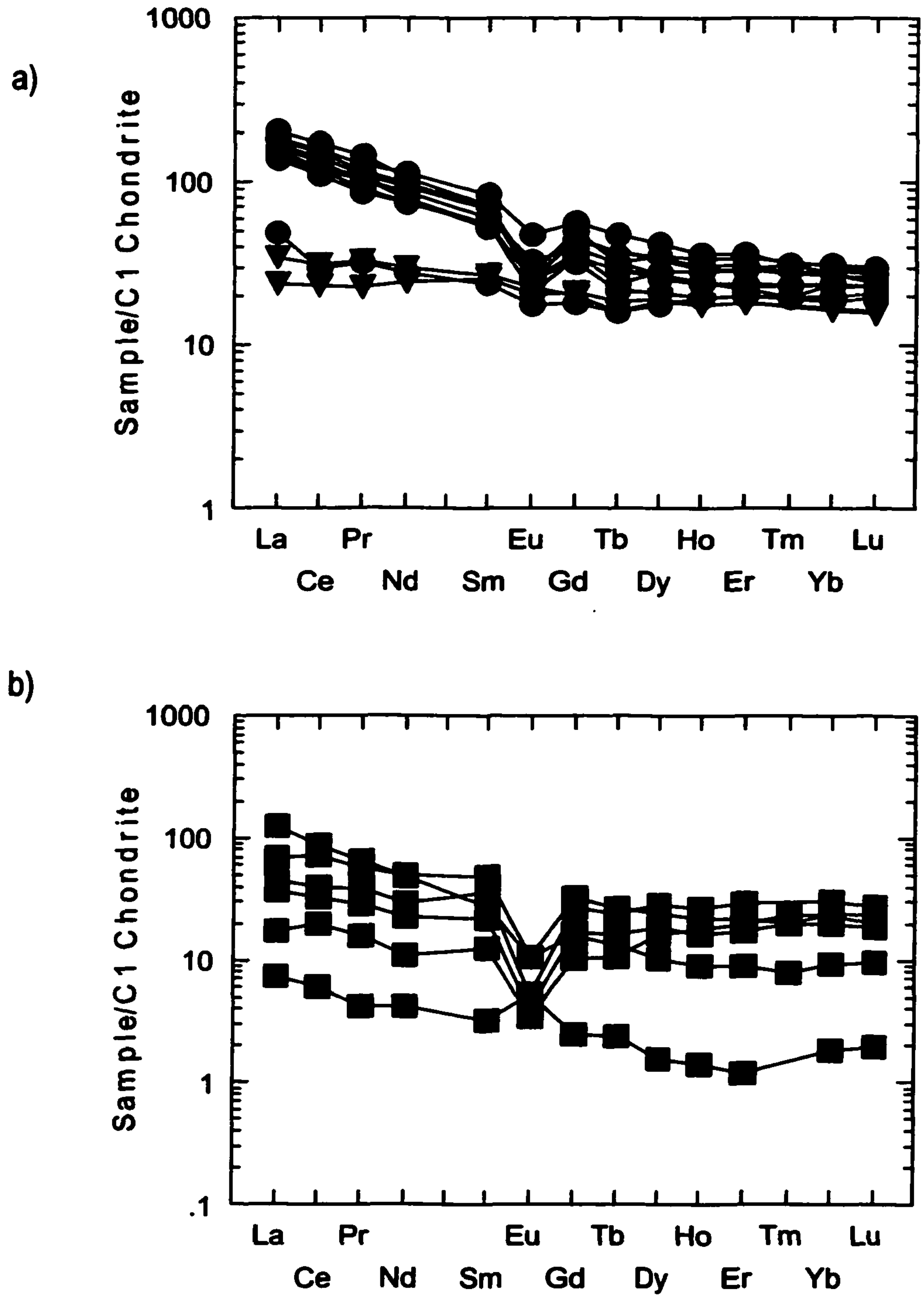


Figure 3.13 (a) REE patterns for the Esperanza and Hornos granites, and (b) REE patterns for the Teticic and Totoltepec granites normalized to Chondrite values, after Evensen et al., (1978). Symbols as in Figure 3.10.

addition, both groups have large negative Eu anomalies with an average of $Eu/Eu^{*} = 0.56$. Exceptions to this pattern are the most mafic samples of the Esperanza granitoids that present flatter patterns and lower ratios of $(La/Yb)_N$ ranging from 1.39 to 2.20 (Figure 3. 13a). On the other hand, the Teticic granites are characterized by much lower overall REE enrichment ($\sum$ REE = 63-130), with variable $(La/Yb)_N$ values ranging from 2 to 13.7, for individual bodies. This group displays the highest negative anomalies in Eu ($Eu/Eu^{*} = 0.2$ to 0.5) in coincidence with its higher silica content and extent of fractionation (Figure 3. 13b).

Additionally, multi-element patterns normalized to ORG (Pearce et al., 1984) are similar for both the felsic plutons of the Esperanza group and the Hornos granites. These groups of plutons display important enrichment in the most incompatible Rb, Ba, K_2O and Th elements, and in a lesser extent in Ce and Sm, indicative of crustal involvement (Pearce et al., 1984), whereas Ta, Nb, Hf, Zr, Y and Yb show significant negative anomalies (Figure 3. 14a). Chondrite normalized patterns (not shown) also exhibit an overall enrichment in most elements, but important negative anomalies in Ta and Nb. Accordingly, multi-element patterns from the Esperanza granitoids and the Hornos granites are more similar to those patterns present in VAG settings. However, similar behavior has also been reported in WPG and COLG settings (Pearce et al., 1984).On the other hand, Teticic granites shows multi-element patterns enriched in Th, Ba, Rb, and K_2O, but slightly depleted in Nb, Ta and Zr similar to VAG settings (Figure 3. 14b). The general enrichment of LFS elements contrasting markedly with the very low HFSE concentrations, is more typical of those patterns described in collisional settings (Pearce et al., 1984).

Contrary to the strong peraluminous character, mineral association and trace element geochemistry, none of the analyzed samples plot within the collisional setting

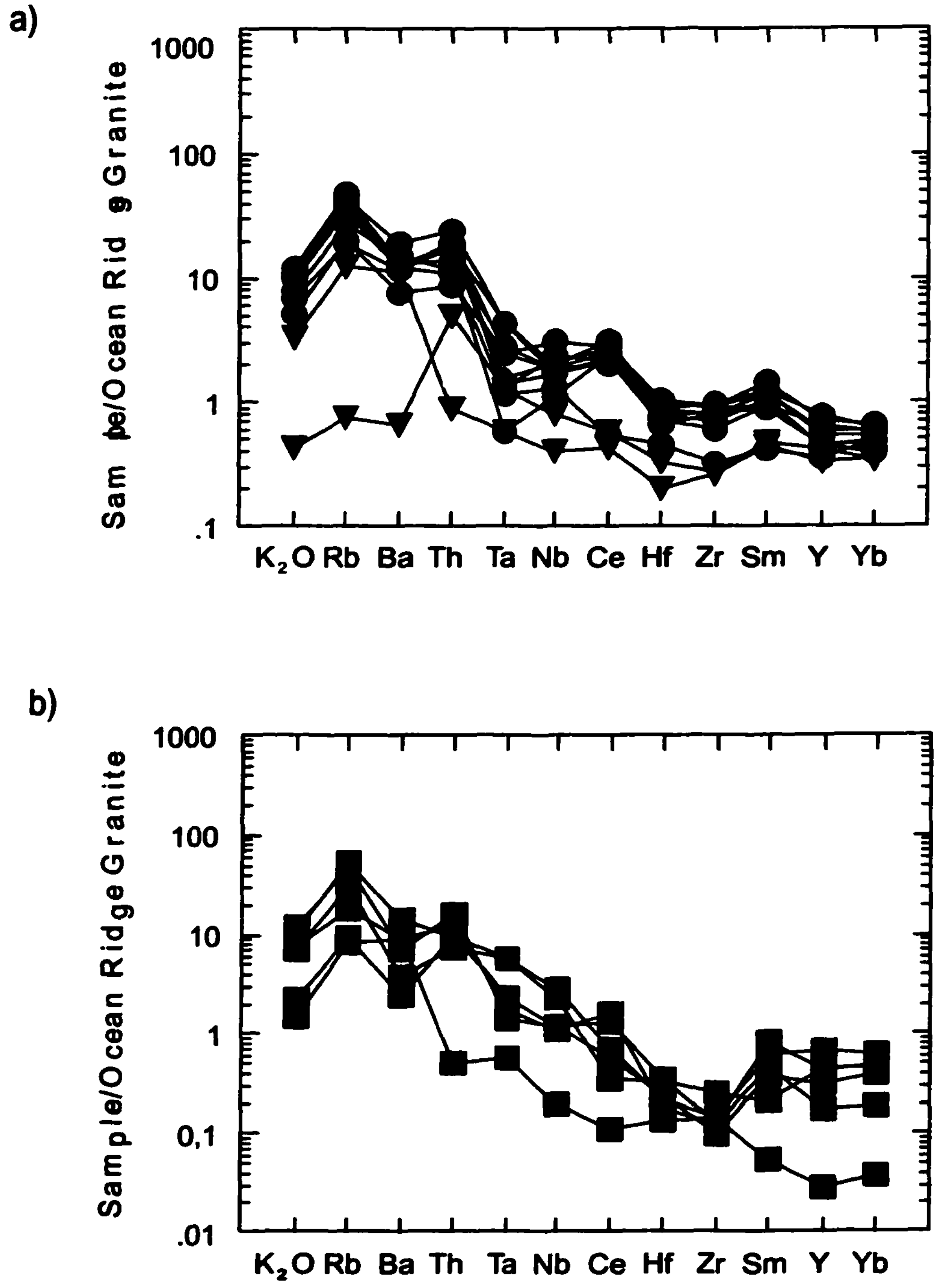


Figure 3. 14 (a) Muli-element diagram for the Esperanza and Hornos granites, and (b) Multi-element diagram for the Teticic and Totoltepec granites normalized to ORG values, after Pearce et al., (1984). Symbols as in Figure 3. 10.

when using the tectono-magmatic discriminant diagrams of Pearce et al. (1984).

Based on the immobile elements Y-Nb, most samples of the analyzed granitoids (Esperanza, Hornos and Teticic) cluster close to, or straddle the boundary between VAG + Syn-COLG and WPG settings, whereas the mafic samples of the Esperanza group clearly fall within the VAG environment (Figure 3. 15a). Using the Yb vs Ta or Rb vs Y+Nb diagrams to better discriminate VAG from Syn-COL granites, the overall samples fall overlapping the VAG and WPG tectonic settings, but none within the Syn-COLG field (Figure 3. 15b).

Using the discriminant criteria of Whalen et al. (1987) to identify A-type granites (WPG), the Hornos granites and some samples of the Esperanza granitoids (samples 139 and 185) tend to fall within the A-type setting (not shown in figure). However, none of the samples are peralkaline and the abundances of Nb (4-31 ppm), Y (23-50 ppm), Zr (90-315), and Ga/Al (1.7-3.48) ratio are lower than the typical A-type or within plate granites suggested by Whalen et al. (1987).

In summary, considering most geologic, mineralogic and geochemical parameters, the Esperanza group seems to represent syn-collisional granitoids mainly derived from crustal melting with a small imput from the mantle registered by the mafic intrusives. The geochemical characteristics of the Hornos granites are similar to those shown by the Esperanza granitoids, but geologically they seem to be post-collisional or within plate granites. On the other hand, most geochemical and mineralogic features of theTeticic granites would also represent collisional granites (Figure 3.12a). However if they were collisional, their assumed age (287 Ma) would be incorrect or they could also represent the syntectonic magmatism of the Alleghenian orogeny.

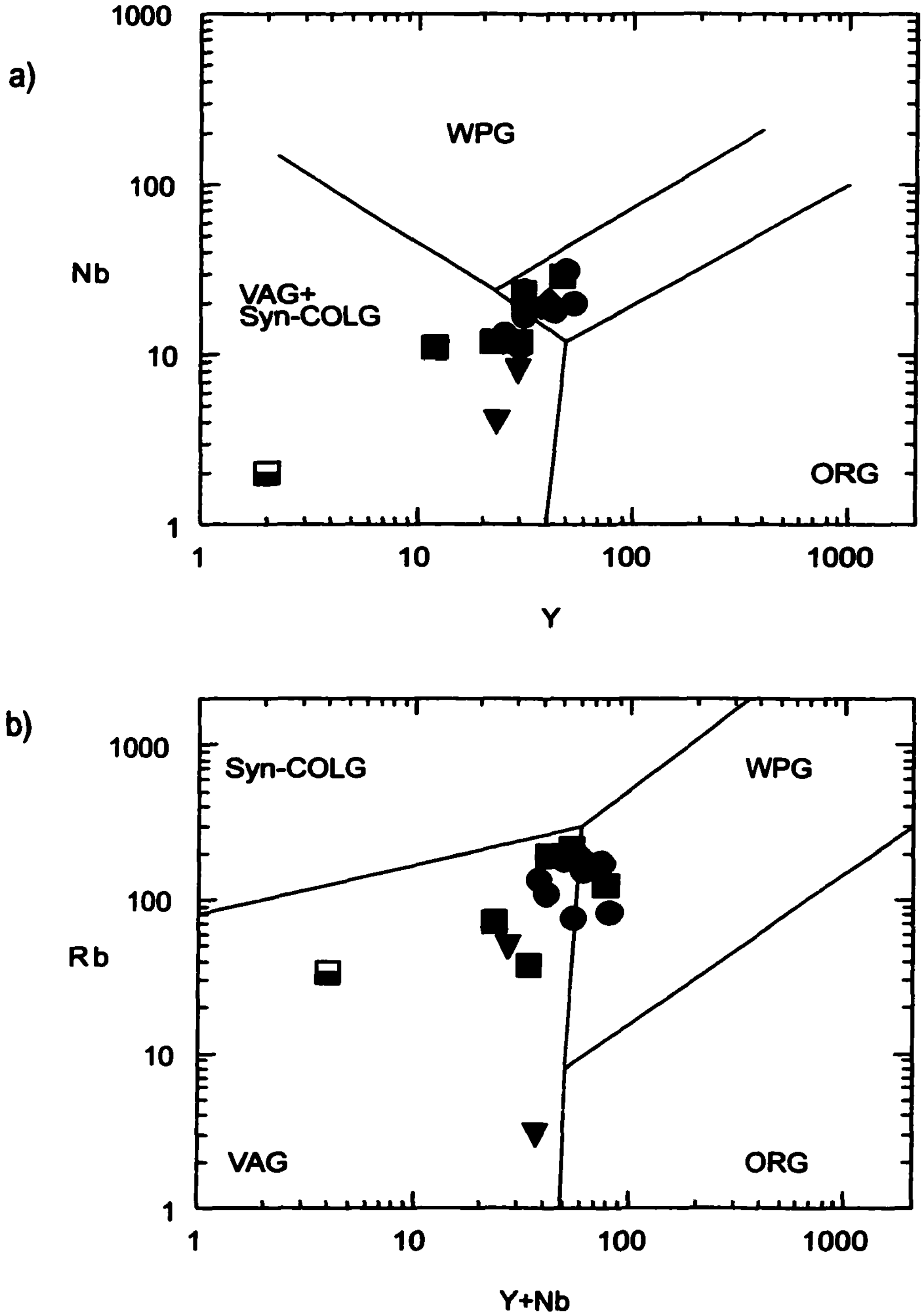


Figure 3. 15 (a) Nb-Y and (b) R- Y+Nb diagrams of Pearce et al., (1984) summarizing the tectonic setting for the different granitoids of the Acatlán Complex. Symbols as in Figure 3. 10.

Nd isotope and Crustal Residence ages

This stuty presents ten new Nd isotopic analyses obtained from the different groups of granitoids included in the Acatlán Complex (Table 3. 4). Felsic samples of the Esperanza granitoids exhibit remarkably homogeneous $\varepsilon_{Nd(440\ Ma)}$ values ranging from - 4.05 to - 4.96 or present-day values ranging from - 7.97 to - 9.64 and crustal residence ages between 1.1 and 1.7 Ga. The more mafic rock of this group displays an $\varepsilon_{Nd(440\ Ma)}$ of + 3.35 or present-day value of + 2.3. However, this sample also shows similar T_{DM} values of 1.2 Ga. Similar results were calculated for the Hornos granites with an $\varepsilon_{Nd(370\ Ma)}$ average of - 5.2, present-day ε_{Nd} values of - 8.7 and crustal residence ages of 1.4 Ga. These results are similar to those reported by Yañez et al. (1991), for the Esperanza group (sample 06) and the Hornos granites (sample 18). The obtained data for the Esperanza granitoids and Hornos granite indicate that these rocks could be the consequence mostly of simple crustal melting and/or mixing of mantle-derived magmas with an older crustal source, as suggested by the > 1.0 Ga T_{DM} of the mafic rock.

On the other hand, the leucocratic rocks of the Teticic group exhibit two different ranges of values. The strongly foliated samples have the more negative $\varepsilon_{Nd(287\ Ma)}$ and present-day average values of - 4.8 and - 6.2 respectively in addition to the older T_{DM} values averaging 2.0 Ga. The reported crustal residence ages for this group represent the oldest ages registered in the Acatlán Complex and probably indicate the presence of a Early Proterozoic crustal source besides the previously suggested Grenvillian rocks (Yáñez et al., 1991). However, it has also been suggested that rocks with $^{147}Sm/^{144}Nd$ ratio higher than 0.14 display anomalous older crustal residence ages (Kerr et al., 1995). Within this group, one sample (182), displays a slightly positive $\varepsilon_{Nd(287\ Ma)}$ and present-day values of + 1.56 and + 0.4 respectively, and T_{DM} of 1.2 Ga, similar to the other analyzed granitoids of the Acatlán Complex. Nevertheless, the strongly foliated Teticic granites were correlated with the Totoltepec granite (287 Ma), the latter shows more positive $\varepsilon_{Nd(287\ Ma)}$ of + 2.5 and lower T_{DM} ranging between 0.66 to 0.8 Ga (Yáñez et al., 1991) suggesting different melt sources.

In summary, the obtained isotopic data in addition to the peraluminous composition of most granitoids of the Acatlán Complex are indicative of a major supracrustal component and a minor imput of mantle-derived magmas. Our data strongly support previous Nd isotopic results and interpretations that most of the Acatlán Complex is composed predominantly of recycled crust (Yáñez et al., 1991). It is noteworthy that only the Totoltepec stock seems to represent Paleozoic additions of juvenile magmas to the crust as suggested by its younger model ages. It is also possible that the suggested correlation between the Teticic and Totoltepec granites is incorrect. In this case, the Teticic granites would be older and possibly late collisional (post Esperanza granitoids) but pre-dating the intra plate activity (pre-Hornos granites).

CHAPTER IV. TECTONIC CORRELATION AND EVOLUTION OF THE ACATLAN COMPLEX

4. 1. Introduction

One of the most important steps in analyzing orogenic systems has been the identification of different tectonostratigraphic terranes, the characterization of tectonic affinity and the type of relationship between terranes. However, a more rigorous analysis based on geologic, stratigraphic, structural, geochemical and geochronologic data should lead to the reconstruction of the different tectonic events that produced such an orogenic system. Therefore, the following analysis is an attempt to understand the tectonic evolution of the Acatlán Complex, within a regional context determined in part by the evolution of the Paleozoic systems where the Acatlán Complex may have formed, including the Appalachian system and the proto-Andean margin of Gondwana.

The Paleozoic systems have traditionally been defined as the final product of continental collision between Laurentia, Baltica, and Gondwana in late Paleozoic time, to generate the supercontinent Pangea. The whole evolution of this Paleozoic orogen has been called the "Wilson Cycle" (Wilson, 1966), where the Eocambrian supercontinent, Rodinia, broke up during late Precambrian time to form the ancestral Iapetus Ocean, which in turn was closed in late Paleozoic time to unify once again all the continental masses.

Even though the Appalachian-Caledonian system was spread out during the break up of Pangea, major pieces have been identified in Europe, Greenland, North America, Africa and South America, and its general reconstruction is well constrained. But, because of the proposed continuous interactions between continents, the initial position of individual terranes has been controversial. Therefore, models suggesting the translation of terranes by continental collisions, (Dalla-Salda et al., 1992; Dalziel et al., 1994; Dalziel, 1997), or the multiple accretion of oceanic island arcs and back-arc systems to Laurentia and Andean-Gondwanan margins (van Staal, 1996; Niocaills et al., 1997; Ramos et al., 1998; Dalla-Salda et al., 1998), are in dispute.

Since most paleogeographic reconstructions and geologic correlations place the Mixteco (Acatlán Complex) and Oaxaca terranes adjacent to the northern Appalachians and the northwestern South American margin during early to middle Paleozoic time, the present analysis focuses on the evolution of the Appalachian system and the proto-Andean margin of Gondwana as a reference framework (Yáñez et al., 1991; Ortega-Gutiérrez et al., 1999; Keppie and Ramos, 1999). Central and southern South America are excluded in this analysis because the geologic framework of these regions differs from the geologic framework of southern and eastern Mexico (Ramos et al., 1986; Ramos, 1988; Dalmayrac et al., 1980; Bahlburg and Breitkreuz, 1991; Dalla-Salda et al., 1998, Ramos et al., 1998; Keppie and Ramos, 1999).

Concerning the area of interest, there is a lot of information summarizing the evolution of the eastern margin of Laurentia, such as those by Williams and Hatcher (1983); Rast (1989); Horton et al. (1989); Hibbard and Samson (1995); and Williams (1995) among other important works. Additionally, there are hundreds of scientific papers analyzing diverse and specific topics of the Appalachian orogen in Canada and U. S. A. In contrast, the available information about northwestern South America is more scarce. Important papers have been written by Restrepo and Toussaint (1988); Forero-Suárez (1990); Restrepo-Pace (1992, 1995); Restrepo-Pace et al. (1997); Benedetto and Ramírez (1982); and Feininger (1987), and many others.

The scope of this chapter includes, primarily, the tectonic significance and characterization of every unit of the Acatlán Complex. Subsequently, a broad correlation is made with the principal terranes and orogenic events recorded along the eastern margin of Laurentia and the Proto-Andean margin of Gondwana. Finally, a whole evolution model of the Acatlán Complex is proposed within the global context suggested for the evolution of the Paleozoic orogenic systems.

The litho-tectonic units that make up the Acatlán Complex are considered to form a composite terrane, where two tectonically juxtaposed sequences constitute the basal terranes: a) the Petlalcingo Group, which represents the structural lower plate; and b) the allochthonous Piaxtla Group which is thrust over the former group. Finally, the slightly metamorphosed Tecomate Formation represents the postcollisional unit that unconformably overlies the other groups (Figure 2. 6). Basically, the metamorphic Acatlán Complex records the evolution of the Paleozoic Appalachian-Andean systems. This complex seems to have originated by the opening and closing of an oceanic basin following the general concept of the Wilson Cycle (Wilson, 1966). However, the Acatlán Complex seems to represent only one of the two-sided margins of the Iapetus Ocean.

4. 2. Tectonic significance of the Petlalcingo Group.

Cosoltepec Formation

Because the Cosoltepec Formation represents 70% of the whole group, the main features of this formation should be considered as the reference in this analysis. The tectonic interpretation of this unit is based on its predominant lithology, geochemical affinity of its metavolcanic rocks, the high negative ε_{Nd} values and relative old model ages obtained for the sediments, as well as the relationship with the adjacent Oaxaca terrane. According to this information, the Petlalcingo Group is interpreted as a siliciclastic sequence deposited on a distal passive margin wich is floored by oceanic crust as part of an ancient continental terrace.

Sedimentologically, the very fine grained siliciclastic sediments (quartzite) interbedded with silstone and black shale, which does not contain limestone, suggest that this group was deposited on a deep ocean floor, with sediment derived from a Proterozoic continental source. The noteworthy absence of limestone in this formation could also suggest that it was deposited in a geographic region located in high latitudes.

Very preliminary data of detrital zircons from quartzite yielded a Transamazonian age of 1800 Ma (Robinson, 1990). However, it is provable that detrital zircons from the Cosoltepec Formation have a late Proterozoic provenance source similar to the Grenville rocks. It is supported by the more systematic and accurate T_{DM} ages from finer sandstone and schist of the Cosoltepec Formation averaging 1.5 Ga, identical to the Grenvillian rocks of Oaxaca (Yáñez et al., 1991; Patchett and Ruiz, 1987). In addition, the middle to highly negative ε_{Nd} values ranging from - 5.6 to - 8.0 strongly suggest the Proterozoic provenance of these sediments. Even more, recent analyses in Ordovician and Carboniferous sediments overlaying the Oaxaca terrane have also yielded 1.0 to 1.1 Ga ages for detrital zircons (Gehrels 2001, written communication). This scenario strongly suggests a Grenville source for sediments of the Cosoltepec Formation however, the presence of detrital zircons of older ages could also be possible taking into account that these terranes (Mixteco and Oaxaca) might have been located sorrounding the Amazonian craton. This hypothesis will be tested in further studies.

Geochemistry of massive and pillowed basalts defines an oceanic setting mainly characterized by MORB and OIB lavas, which suggest the presence of an ocean floor of possible Ordovician age (452 Ma according to Armstrong *in* Ortega-Gutiérrez et al., 1999). The widespread presence of basaltic rocks all over the Cosoltepec Formation suggests that these rocks constitute the substrate of the siliciclastic sediments, supporting the interpretation of a very distal passive margin. In addition, there are no reports of continental basement beneath the Petlalcingo Group.

Even though there are no paleontologic constraints on the age of the Cosoltepec Formation, it would be pre-Devonian considering the unconformity with the overlying Tecomate Formation. More specific, it should be considered pre-Silurian base on the time of thrusting and emplacement of the Piaxtla Group over the Cosoltepec Formation. Furthermore, if the age of 452 Ma from basaltic rocks included in the unit is correct, an age of Cambrian (?) to Early - Middle Ordovician for the Petlalcingo Group seems likely.

The presence of Grenvillian rocks of the Oaxaca terrane located to the east of the Acatlán Complex displaying similar isotopic values to sediments of the Cosoltepec Formation, has led to the interpretation that these terranes were linked early in their history. In addition, because of the presence of platform limestones of Tremadocian age with fauna of Gondwanan affinity overlying the Oaxaca terrane, the Mixteco-Oaxaca block is considered to be of Gondwanan provenance.

Another interpretation suggests that the Cosoltepec Formation represents an enormous trench-fill deposit (Ortega-Gutiérrez, 1993; Ortega-Gutiérrez et al., 1999). However, in spite of the enormous outcropping area there are no previous reports of any chaotic melange with exotic blocks within the unit, which would characterize a subducction complex. Besides, the sedimentary sequence does not contain any volcanic influence of a proximal volcanic arc.

4. 3. Regional Correlation of the Petlalcingo Group.

Since the Paleozoic Appalachian system was considered to evolve from the opening and closing of an ocean between two continental plates (Wilson, 1966), the western margin of the Iapetus Ocean was largely identified in the eastern miogeocline of Laurentia. However, the eastern margin of Iapetus is not clearly defined, thus it has been proposed to be along the Baltica, Africa and Amazonian cratons in very different positions

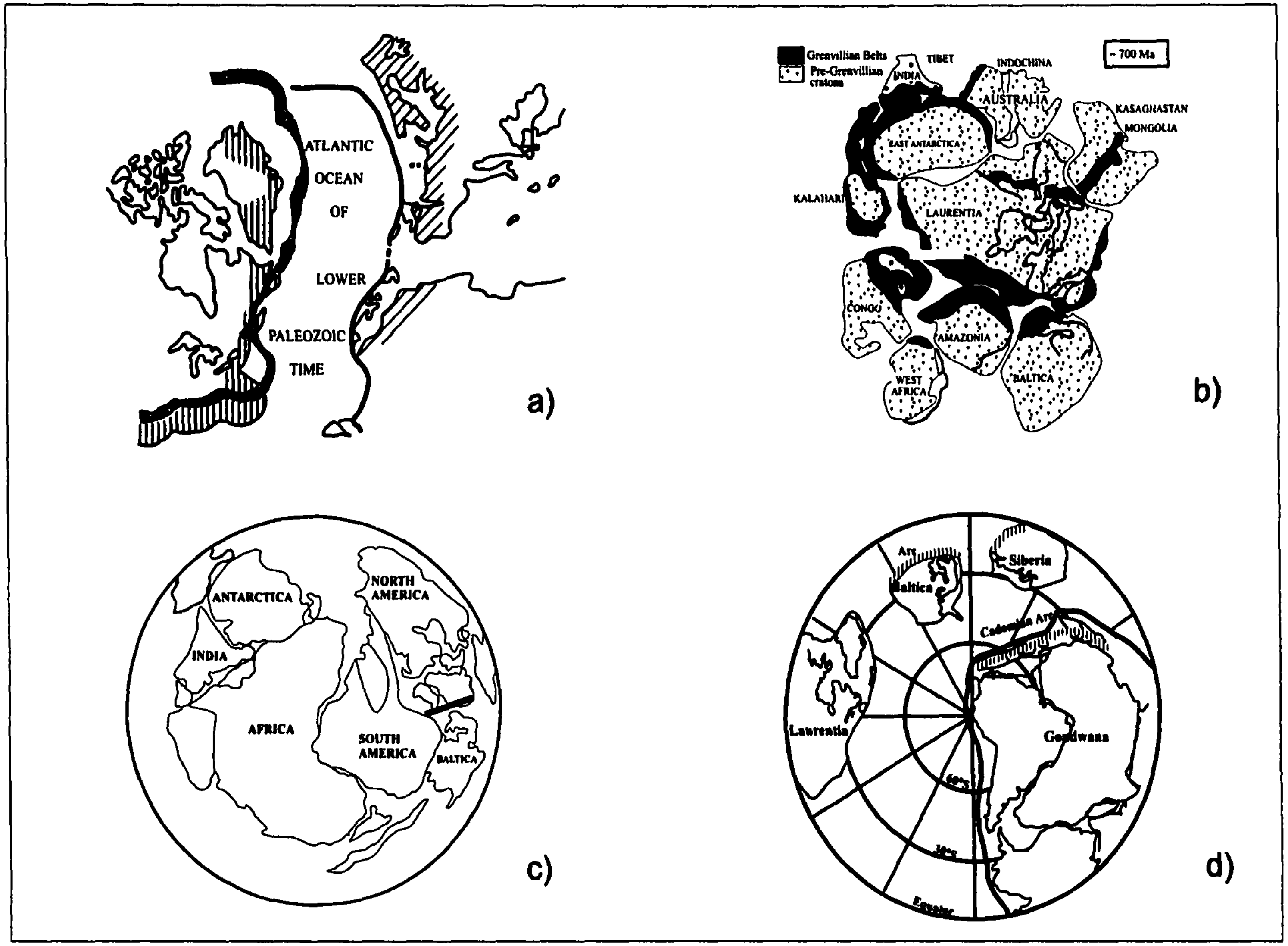


Figure 4. 1 Paleogeographic reconstructions showing different continental blocks as part of the eastern Iapetus margin a) after Wilson (1966); b) after Hoffman (1991); c) after Dalziel (1994), and d) after Torsvik (1996).

and variations (Wilson, 1966; Hoffman, 1991; Torsvik et al., 1996; Dalziel, 1994) (Figure 4. 1). In order to establish a possible paleogeographic correlation for the Cosoltepec Formation, stratigraphy of the Laurentian and Gondwanan margins were studied. In general, there is a major stratigraphic difference between those margins, on one side of the Iapetus, the Laurentian passive margin mainly developed a carbonate sequence, while the Gondwana margin was characterized by a siliciclastic sedimentation more akin to the Petlalcingo Group. A brief description of the Laurentian and Gondwanan margins is given below.

Laurentian Margin

The rift-drift stages recording the evolution of the Iapetus Ocean are well documented along the entire eastern margin of Laurentia, ranging from about 700 Ma for the former to 460 Ma for the end of the latter stage. Rifting occurred parallel to the Grenville orogen, therefore, the basement rocks along this passive margin belong to this belt. During the drift period (Cambrian-Early Ordovician) the passive margin developed a thick platform sequence of carbonate sediments because of the equatorial position of Laurentia. Offshore slope-rise facies sediments are represented by deep-water limestone, carbonate breccia, shale, chert and some basalts representing the ocean floor. Deep-water sediments of the slope-rise prism would represent abyssal submarine fan deposits coeval with the carbonate sequence of the platform.

In the northern Appalachians, the passive margin is represented by the external and internal Humber terrane (Williams and Hatcher, 1983, Williams, 1995). On the other hand, in central and southern Appalachians, the passive margin is constituted by the Native terranes including the Laurentian Platform (Blue Ridge Province) and the Allochthonous-offshore sequences (e.g. Hamburg, Westminster and Taconic Range

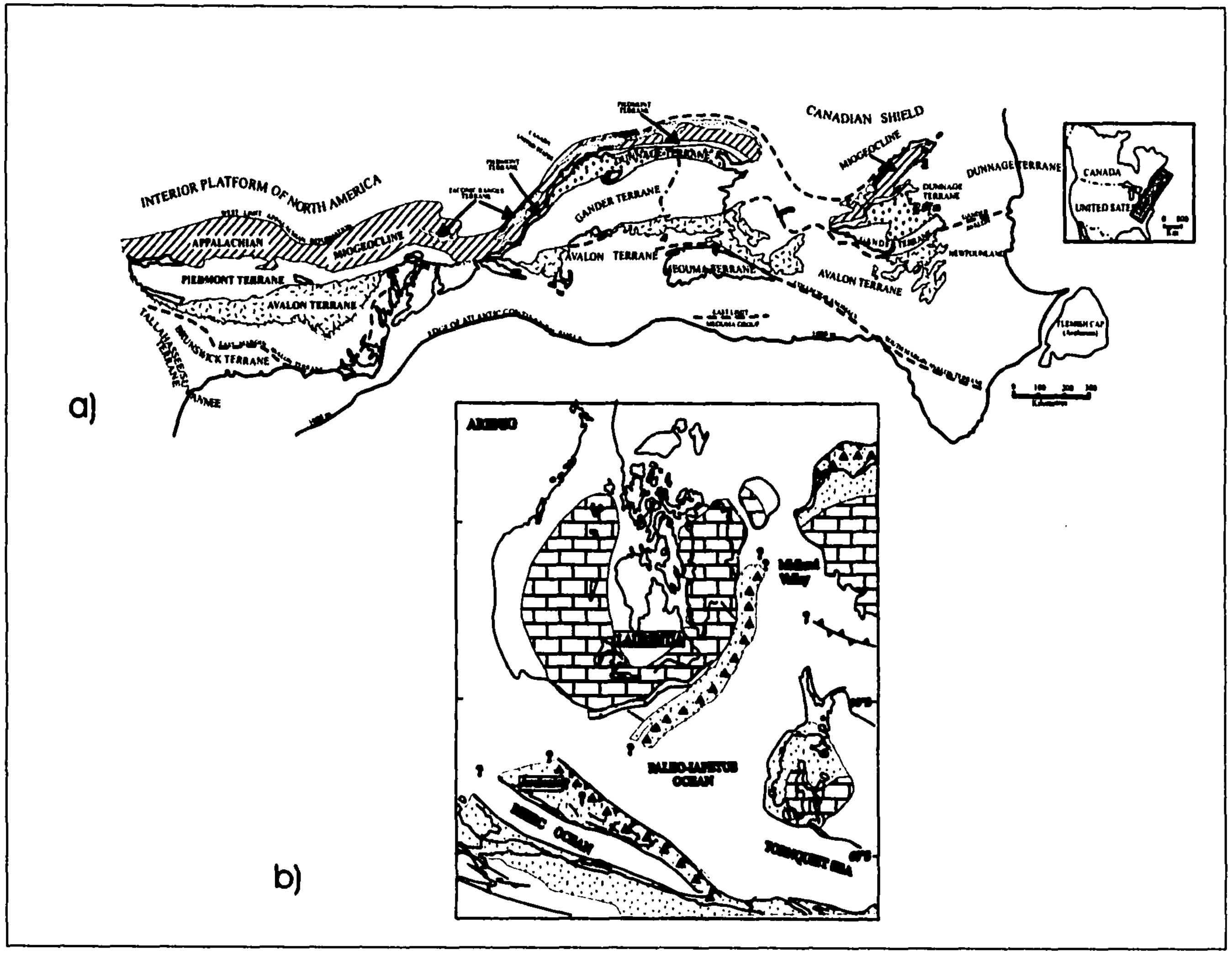


Figure 4. 2. A)Eastern margin of North America showing the main distribution of the Laurentian miogeocline (after Williams and Hatcher, 1983), b) Early Ordovician Paleogeographic reconstruction of Laurentia displaying its carbonate platform and the surrounding intra-oceanic arc (after Torsvik, 1996).)

terranes) (Horton et al., 1989; Rast, 1989) (Figure 4. 2). The deposition of carbonates in this miogeocline was interrupted in the Middle Ordovician by the emplacement of oceanic terranes during the Taconic orogeny.

The well developed early Paleozoic carbonate platform-offshore-rise sediments deposited along the Laurentian passive margin strongly contrast with the siliciclastic sediments of the Petlalcingo Group. Therefore, a possible correlation of this group with the eastern passive margin of Laurentian (or western margin of the Iapetus Ocean) should be excluded.

Gondwanan Margins

On the other hand, the location and evolution of the eastern margin of the Ipetus Ocean is not well defined, because it has been correlated with regions that are now geographically dispersed (e.g., Baltica, northwest Africa, northwestern South America or southwest South America) (Figure 4. 1). According to such models, possible correlations of the Petlalcingo Group with any passive margin sequence from those regions are more difficult to establish. However, if we consider that the Acatlán Complex was tied to the Grenville belt of western Gondwana, represented in Mexico by the Oaxaca terrane, the best area for correlation is the northwestern region of South America where similar rocks are exposed. This is supported by additional correlation between Ordovician and Silurian sequences from Oaxaca and Tamaulipas in Mexico with similar units in western Venezuela and Colombia.

Taking into account the high-pressure rocks (Piaxtla Group) that tectonically over thrust the Petlalcingo Group, the best correlations should also consider the siliciclastic sequences of passive margins and associated high-pressure units surrounding peri-

Avalonian terranes in the present northern Appalachians (van Staal et al., 1990, van Staal, 1994). High-pressure sequences of Paleozoic age have not been reported in the Andean region of Venezuela, Colombia and Ecuador.

Proto-Andean margin of Gondwana. According to Forero-Suárez (1990), and Restrepo and Toussaint (1988), the Paleozoic evolution of the northwestern proto-Andean margin of Gondwana is represented by the Eastern South American and Central Andean Provinces (Figure 4. 3).

The Eastern South American Province consists of a metamorphic basement and lower Paleozoic sedimentary rocks forming the cover. This autochthonous terrane includes the Guyana shield, which consists of Archean high-grade gneiss, migmatites and granites ranging between 3.7 and 3.4 Ga. It is surrounded by three high-grade metamorphic belts considered to be mobile zones produced during the Transamazonian (1.8 to 2.0 Ma), Paraguayan (≈1.6 Ga) and Nickerian-Orinoquian-Grenvillian (1.1 to 1.3 Ga) orogenies (Trompette, 1994). Noteworthy is the scarcity of Pan-African rocks, only in Venezuela and Peru exist reports of this event (Burkley, 1976 *in* Stewart et al., 1999 ;Cobbing et al., 1977). Basement rocks of cratonic Guyana do not outcrop in Ecuador, but considering that granulites have been reached in subsurface, this region could be the continuation of the eastern Cordillera terrane of Colombia and Venezuela (Feininger, 1987).

The northwestern South American craton is covered by a sequence of clastic platform sediments deposited mainly during the Middle Cambrian to Middle Ordovician and consists of fine conglomerates, sandstones, siltstones and shales with minor carbonates (Harrington and Kay, 1951; Benedetto and Ramírez, 1982). Latest Cambrian-Ordovician sediments of this region (e.g., El Baúl in Venezuela and La Macarena in Colombia) show major similarities with those sediments in Oaxaca having faunas of

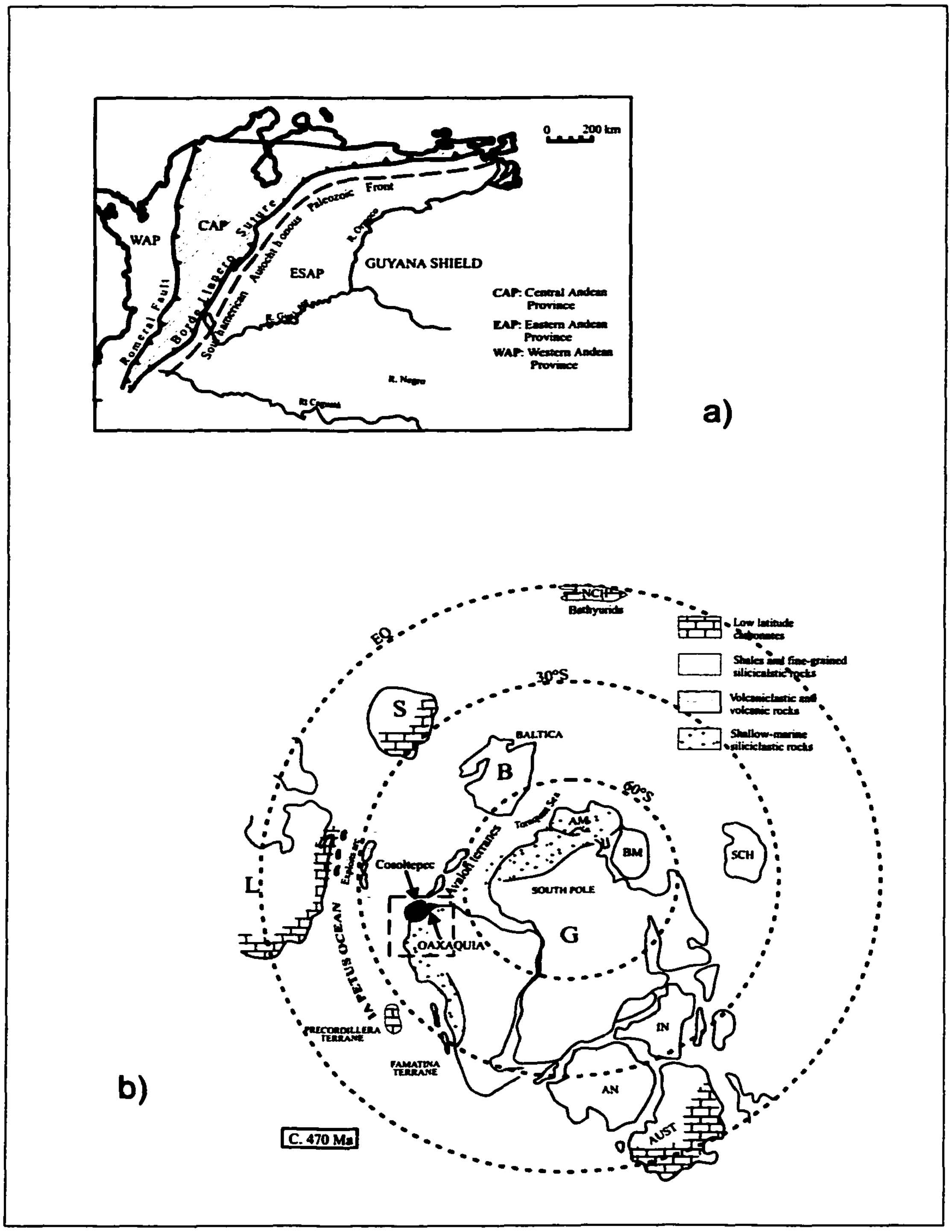


Figure 4. 3. a) Geologic provinces of northern South America according to Forero-Suárez (1990), b) Early Paleozoic reconstruction of Gondwana showing its western clastic miogeocline. The Oaxaca terrane and Cosoltepec Formation placed adyacent to the northwestern Amazonian craton (after Astini, 1998).

Gondwanan affinity and simliar also with sediments in Argentina, Bolivia, Avalon, and Baltica (Benedetto, 1998). Middle to upper Paleozoic rocks in this region are absent.

The Central Andean Province is divided into eastern and western regions. The eastern region is located west of the Guyana craton forming a discontinuous belt of Late Precambrian basement represented by granulitic, anorthositic and amphibolitic gneisses. These Precambrian rocks crop out in Santa Marta, Santander, and Garzón massifs in Colombia, and Cordillera de Mérida massif in Venezuela, and altogether, are considered to be part of the Grenville belt of the Gondwana continent. In contrast, Forero-Suárez (1990) considers these outcrops as parts of an allochthonous belt.

Recent isotopic data from the Garzón Massif show great similarities with those Grenvillian rocks of Oaxaca (Patchett and Ruiz, 1987; Restrepo-Pace et al., 1997; Ruiz et al., 1999). An important fact is that in Cordillera de Mérida and Arequipa massifs of Venezuela and Perú respectively, there are scarce plutonic and metamorphic events with ages ranging between 620 and 650 Ma affecting the Grenvillian rocks (Cobbing et al., 1977; Burkley, 1976 *in* Stewart et al., 1999). However, similar events have not been repported in the Grenvillian rocks of Oaxaca.

In the eastern Central Andean Province, the Grenvillian basement is unconformably overlain by low to medium-grade metamorphic rocks, which consist of phyllite, quartzite, graphitic schists and greenschists coeval with the Cambrian-Ordovician rocks of the eastern platform (Benedetto and Ramírez, 1982). The presence of any Silurian sediments in Colombia is poorly documented, but some fossils have been reported in low-grade metamorphic rocks of the Santander massif (Forero-Suárez, 1990). However, unmetamorphosed Silurian sediments are well represented in the Cordillera de Mérida region (Boucot et al., 1972). These Silurian rocks of Venezuela have been correlated with a small outcrop located in Ciudad Victoria, Tamaulipas in northeastern Mexico (Boucot et al., 1997; Stewart et al., 1999). Upper Lower to Upper Devonian epicontinental sediments made up mostly of conglomerate, sandstone, and shale unconformably cover the previous units suggesting a Silurian orogeny. Magmatic events affecting these metamorphic rocks have been reported from granitoids of mainly Middle-Ordovician (460 Ma) and Silurian ages (420 Ma) (Forero-Suárez, 1990; Restrepo-Peace, 1995). In addition, very scarce Devonian intrusions have been reported in this region.

On the other hand, the western region of the Central Andean Province is an Early Paleozoic belt that extends from northern Colombia to northern Perú (Restrepo-Pace 1992). Lithologically, this terrane is very homogeneous and consists of quartzite, phyllite, graphitic schist, greenschist and amphibolitic greenschist. Some Ordovician graptolites have been reported from these rocks in Colombia (Botero, 1940 and Mojica et al., 1989 *in* Forero-Suárez, 1990). Geochemical analyses carried out in sediments of this belt reveal two distinct sources for the protolith: the amphibolitic schists display an oceanic island arc affinity, while the graphitic schists suggest rocks derived from a continental source (Restrepo-Pace, 1992). The metamorphic event affecting these rocks is suggested to be Late Silurian-Early Devonian because of the unmetamorphosed Devonian sediments overlying this region (Forero-Suárez, 1990; Restrepo-Pace, 1992). In Ecuador, this metamorphic belt is exposed along the eastern slope region, but there is not any report constraining its age (Feininger, 1982).

In summary, the northwestern proto-Andean margin of Gondwana was surrounded by siliciclastic platform sediments developed during the Middle Cambrian to Middle Ordovician overlapping metamorphic rocks of Grenville age. To the west, the coeval slope-rise sediments would apparently be represented by those low-grade-fine-grained metamorphic sediments of the Central Andean Province (Restrepo-Pace, 1992). Contrary to what is well known in eastern Laurentia, the absence of rift-type sedimentation and related magmatism in this northwestern margin of Gondwana is noteworthy.

There are two main Paleozoic orogenic events in northwestern Andes. The Middle Ordovician orogeny is represented by the widespread ≈ 460 Ma magmatism that affects the eastern part of the Central Andean Province, and recorded by the stratigraphy in the Mérida Massif where unmetamorphosed sediments of Late Ordovician-Silurian age unconformably overlie low-grade metamorphic rocks of Cambrian-Ordovician age. This Ordovician event has been termed Caparonensis orogeny (Restrepo-Pace, 1995). On the other hand, the Silurian orogeny is well documented in the Perijá and Santander massifs where late Lower Devonian sediments cover the low-grade metamorphic rocks of Ordovician-Silurian age in addition to the mainly Silurian (420 Ma) plutonic event recorded in the eastern Central Andean Province. This orogeny has been described as a Caledonian event by Forero-Suárez (1990).

Peri-Gondwanan margin in the northern Appalachian. According to Williams and Hatcher (1983), those sequences representing the eastern counterpart of the Iapetus Ocean are distributed within the Exploits zone of the Dunnage terrane, as well as the Gander, Avalon and Meguma terranes of the northern Appalachians. In the central and southern Appalachians, the Gondwanan margin is represented by the Carolina and Piedmont terranes. In this analysis the possible correlation of the Petlalcingo Group would be focused on the Gander terrane since the lithology and tectonic setting are more similar. Terranes in central and southern Appalachians are not considered here because the Piedmont and Carolina do not exhibit correlative units to the Petlalcingo Group, or at least they are not well represented (Horton et al., 1989).

Gander Terrane. This terrane is distributed along the northern Appalachian system, from Newfoundland in Canada to New England in U.S.A. (Williams and Hatcher, 1983), with some possible correlatives in the British Caledonides (Winchester and van Staal, 1995). The Gander terrane has been considered for a long time to be the eastern passive margin of the early Paleozoic Iapetus Ocean (Williams, 1979; Colman-Sadd, 1982). Lithologically, this terrane is characterized by a thick sequence of quartzite, siltstone and shale that grades into psammitic schist, gneiss and migmatite (Williams, 1995). Similar Barrovian-type metamorphism is also present in the Petlalcingo Group where phyllite and quartzite of the Cosoltepec Formation grade into biotite-schist of the Chazumba Formation and finally into gneiss and migmatite of the Magdalena Formation (Ortega-Gutiérrez, 1975).

The age of the Gander terrane is poorly constrained but an Early Cambrian to Early Ordovician age has been considered based on its stratigraphic relationships, very scarce non *in situ* fossils, sediment provenance and isotopic studies (Neuman, 1984; Dunning et al., 1990; David et al., 1991; Colman-Sadd et al., 1992; van Staal et al., 1996).

According to van Staal et al. (1996), the more reasonable sedimentary sources for the Gander terrane are the Avalon terrane and the Amazonian craton because of the presence of detrital zircons of highly variable ages including Early Cambrian (0.54-0.55 Ga), Neoproterozoic (0.6-0.8 Ga), Mesoproterozoic (1.0-1.5 Ga), and Transamazonian (1.9-2.0 Ga) and Archean rocks (2.5-2.7 Ga). Although the basement-cover relationship between the Gander and Avalon terranes is not exposed, deep seismic data (Stockmal et al., 1990), Nd isotopic signatures (D' Lemos and Holdsworth, 1995; Whalen et al., 1994, 1996), and sediment provenance (David et al., 1991; van Staal et al., 1996), strongly suggest a close basement-cover relationship between them.

Nd studies carried out in metasediments and granitoids of the Gander terrane indicate a mean T_{DM} of 1.5 Ga and very negative ε_{Nd} values ranging between - 9 and - 7 (D'Lemos and Holdsworth., 1995). It is important to point out that model ages and isotope

Nd values from sediments of the Gander terrane are very close to those values obtained from sediments of the Cosoltepec Formation (Yañez et al., 1991).

In Newfoundland, the upper contact of the Gander terrane is clearly tectonic where ophiolitic rocks of the Exploits zone (Dunnage terrane) overlie sediments of the Gander Group. In the Mount Cormack zone (central Newfoundland) sediments of the Gander terrane surrounded by ophiolitic rocks of the Exploits zone are interpreted to be a structural window, with the Exploits zone as an allochthonous terrane (Colman-Sadd and Swinden, 1984). In New Brunswick, the Gander terrane is overlain by a Middle Ordovician volcanic sequence considered to be part of the Dunnage terrane (van Staal, 1987, van Staal et al., 1990). Similar relationships are found in Maine and New England (Williams and Hatcher, 1983). Tectonic relationships between the Gander and Dunnage terranes resulted from the Taconic orogeny.

The oldest period of plutonism within the Gander terrane was *ca.* 480 Ma and is related to the emplacement of the allochthonous Dunnage terrane (Colman-Sadd et al., 1992). However, the main phase of metamorphism and plutonism of this terrane has been dated as Silurian, with syntectonic intrusives yielding 430 to 415 Ma (Dunning et al., 1990). Finally, this terrane was affected by another granitoids with ages averaging 380 Ma.

According to the analyzed information about the western proto-Andean margin of Gondwana and the eastern Gondwanan margin in the northern Appalachians, the Petlalcingo Group presents great similarity with the Gander terrane, but there are also strong differences. The Andean region apparently presents the more appropriate tectonic framework with respect to the Oaxaquian links to South America and the suggested relationship between Oaxaca and the Acatlán Complex (Ortega-Gutiérrez et al., 1995)

Main similarities with the Gander terrane are: thick and widespread sequence of quartzite and phyllite, similar model ages and ε_{Nd} values, Proterozoic continental source, tectonic association with overlying oceanic terranes (Dunnage in the northern Appalachian, Xayacatlán Formation in Acatlán), and possibly similar age. Main differences are: type of basement (continental in the Gander terrane, oceanic in the Petlalcingo Group), and different continental blocks bordering to the east (Grenvillian rocks in Acatlán, Avalonian rocks in northern Appalachians).

Main similarities with the northwestern proto-Andean terranes are: siliciclastic sediments, Proterozoic source of the sediments, possibly similar age, and apparent close relationship with the "Oaxaquia block" (Grenville basement overlain by Tremadocian and Silurian sediments) similar to the Eastern Andean terrane. Although the substrate of western Central Andean terrane in Colombia is unknown, main difference with the Petlalcingo Group is the absence of early Paleozoic oceanic terranes tectonically overlying the Central Andean Province (e.g., the Piaxtla Group overthusting the Petlallcingo Group).

Based on this data, the best position of Oaxaquia and the Acatlán Complexes during the Early Cambrian to Middle Ordovician, as part of the eastern passive margin of the Iapetus Ocean, would be located southwest of the Avalon-Gander block along the northwestern proto-Andean margin of Gondwana, following the paleogeographic reconstruction proposed by Dalziel et al. (1994), van Staal et al. (1996), Murphy et al. (1999), and Keppie and Ramos (1999) (Figure 4. 3).

4. 4. Tectonic significance of the Piaxtla Group

The Piaxtla Group represents an allochthonous terrane made up of two very distinctive and contrasting units that are affected by the same high-pressure eclogite facies metamorphism. The older Xayacatlán Formation consists mainly of metabasites, metasediments and minor metagabbros and serpentines representing the only outcrop of eclogites in Mexico. The younger unit, the Esperanza granitoids, consists of highly deformed and high grade metamorphosed plutonic complexes, which intruded the Xayacatlán Formation during a mainly Silurian magmatic event (440 to 425 Ma). This group tectonically over thrusts the Petlalcingo Group, and in turn, is unconformably overlain by the Tecomate Formation of Devonian age.

Xayacatlán Formation

The most distinctive feature of this formation is the high-pressure mineralogical assemblage displayed in metasediments and metabasalts, which in the latter rocks consist basically of omphacite-quartz-garnet-barroisite-phengite-rutile (Ortega-Gutiérrez, 1975, 1991). Petrologic studies consider that the eclogitic phase reach its peak metamorphic conditions at temperature and pressure up to 550^{o} C and 12 to 15 kb, respectively, suggesting the subduction of these rocks to deep levels in the crust (Ortega-Gutiérrez, 1995; Ortega-Gutiérrez and Reyes-Salas, 1997; Meza-Figueroa, 1998).

Geochemical analyses from metabasites indicate an exclusive oceanic affinity, displaying MORB and OIB settings as well as IAT environment (Meza-Figueroa, 1998). In any case, metabasaltic rocks suggest a complex oceanic setting, where a normal or enriched oceanic floor and island arc rocks are closely related. Even though there is no crystallization age for these metabasites, a pre- 440 Ma, Ordovician to Cambrian age is reasonable because of the intrusive relationship with the Esperanza Granitoids. The Devonian age of 388 Ma for this formation (Yañéz et al., 1991) is apparently associated with the Acadian orogeny.

Esperanza Granitoids

One of the most intriguing features of the Acatlán Complex is the presence of the

Esperanza Granitoids, which are highly deformed (mylonites) and metamorphosed to eclogitic conditions (Ortega-Gutiérrez, 1991). These plutonic complexes range in composition from mainly granites to scarce granodiorites, diorites and gabbros. Most of these bodies have a preferential felsic and intermediate composition, are peraluminous, and classify as S-type granites according to Chappell and White (1992). However, there are some peraluminous plutons that fall in the most fractionated field of the I-type plutons characterized by the abundance of biotite. According to the Pearce et al. (1984) diagrams, most granitoids cluster close to or straddle the boundary between VAG+Syn-COLG and WPG settings and only the most basic fall in the VAG field. In addition, ε_{Nd} values are mildly negative at around - 4 and model ages range from 1.1 to 1.7 Ga. The age of this magmatic event has been dated in between 440 to 425 Ma (Robinson, 1990; Ortega-Gutiérrez et al., 1991). Based on geochemical analyses, mineralogical association, geologic relationships and isotopic age, the Esperanza granitoids represent a magmatic event derived from crustal melting (crustal recycling) with very little input from the mantle produced by major crustal and lithospheric thickening during a continental collision in Silurian time.

In summary, the Piaxtla Group represents a complex subduction region where an ocean floor, but possibly oceanic arc-trench system, was involved in the consumption of ensimatic rocks to deep levels reaching eclogite facies. During the consumption of these oceanic rocks by continental collision, the subsequent lithospheric thickening enabled crustal melting to produce the Esperanza Granitoids. This collisional event also allowed the overriding of the Piaxtla Group on top of the Cosoltepec Formation.

4. 5. Regional correlation of the Piaxtla Group.

Proto-Andean margin of Gondwana

At the present time, there is no record of any early to middle Paleozoic high-pressure rocks in the northwestern proto-Andean margin of Gondwana. The only reported subduction complex in the South American continent is located in southern and central Chile associated with the Chilenia terrane (Hervé, 1988). It is characterized by the presence of melanges with exotic blocks of different ages and lithology. High- pressure metamorphic rocks are found as boulders associated with serpentine bodies. Silurian, Devonian and Permian fossils have been reported in different lithologies, however radiolarian cherts in the matrix have yielded Late Carboniferous to Permian ages. Therefore, no correlatives to the Piaxtla Group exist in the Proto-Andean margin of Gondwana.

Eastern margin of Laurentia. In the Appalachians, there exists a well documented record of high-pressure rocks extending the length of this orogen from the Blue Ridge province in North Carolina to the Baie Verte-Bronton line in Newfoundland (Adams et al., 1995). All these outcrops, which include blueschists or eclogitic rocks, seem to be associated with the suture zone between rocks of different oceanic affinity and the passive margin of the eastern Laurentia floored by Grenville basement (Figure 4. 4). The consumption of oceanic crust tied to the Laurentian margin in an east-dipping subduction zone, followed by continent-continent or continent-ocean arc collision during the Taconic orogeny produced the high-pressure rocks of this belt. Besides, this orogeny also produced large folding, thrust faulting and the emplacement of oceanic allochthonous above the miogeoclinal rocks of Laurentia. Folding and thrusting in this process had a main northwest displacement (Adams et al., 1995; Williams and Hatcher, 1983). Obviously, these regions are unrelated to eclogites of the Piaxtla Group because of the regional geologic context, however it is important to note the abundance of high-pressure rocks in this region, in contrast to the Andean margin.

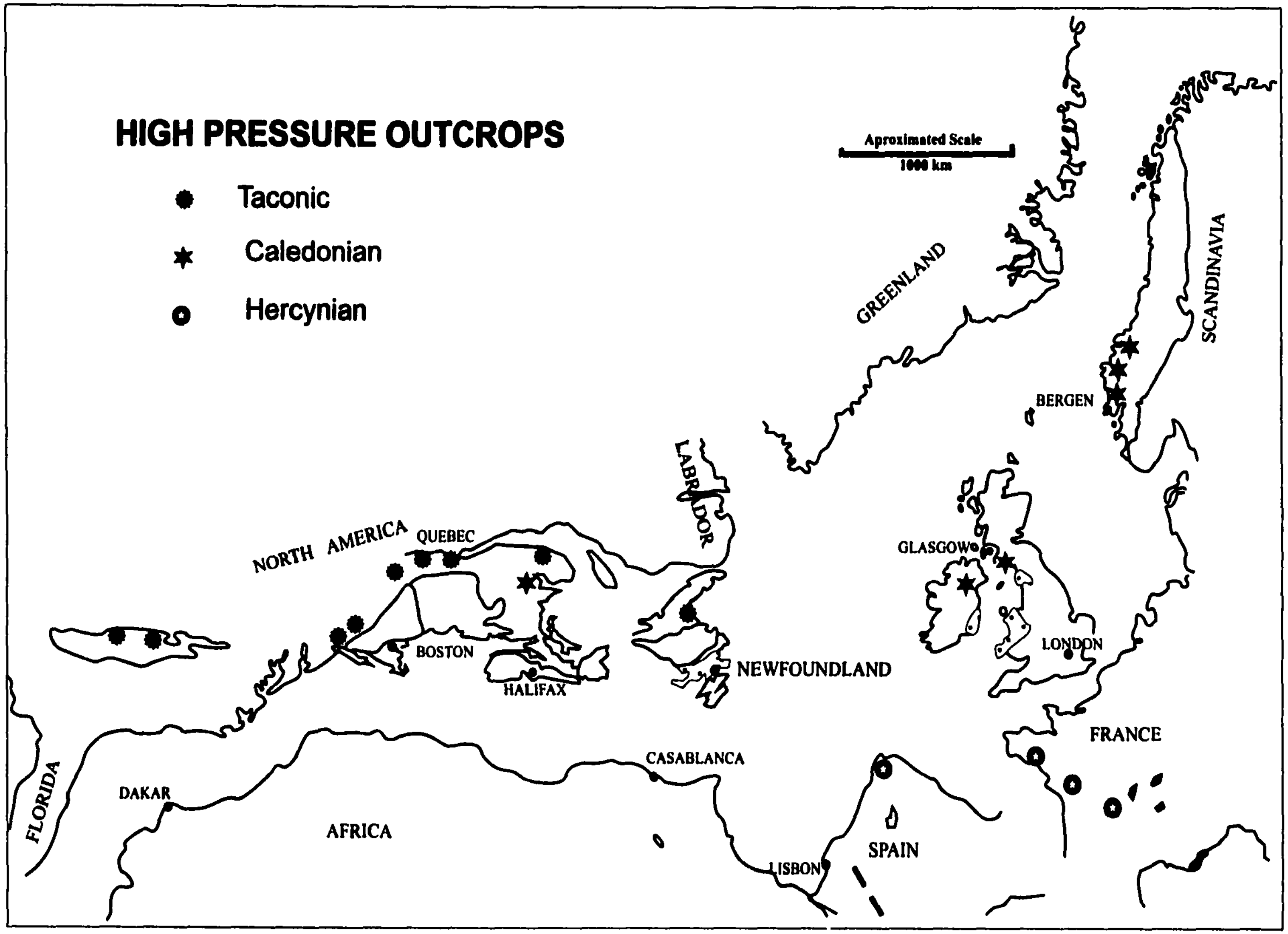


Figure 4.4 Distribution of high- pressure rocks along the Appalachian-Caledonian-Variscan systems. The high-pressure rocks of the Piaxtla Group would be included somewhere in the southern Caledonian belt. (After Robinson,1988).

Peri-Gondwanan margin in the northern Appalachians. High-pressure rocks associated with the peri-Gondwanan terrane in the northern Appalachians have only been reported in New Brunswick (Brunswick subduction complex), where the Gander and Dunnage terranes continuously interact to develop a complex evolution (van Staal et al., 1990; van Staal, 1994, van Staal and de Roo., 1995). High-pressure rocks of the Brunswick complex consist of blueschist. According to van Staal and de Roo (1995), the Early to Late Ordovician evolution of the oceanic Dunnage terrane (Figure 4. 2) was characterized by several collisions of different oceanic arcs which were developed in a subduction zone with constant arc-polarity reversals.

Within the Dunnage zone a back-arc basin was generated during the Middle to Late Ordovician, to finally produce a west-dipping subduction zone because of the closure of this back-arc basin at the end of the Late Ordovician. The west-dipping subduction zone allowed the emplacement of the Dunnage terrane over the Gander zone in a southeastward direction. The closure of this Late Ordovician arc-back-arc-basin system was produced by a continental collision (continent against peri-Gondwanan micro continents) in Silurian time. Based on Ar/A data, van Staal et al. (1990) considers that this event lasted from Late Ordovician (447 Ma) to Late Silurian (416 Ma). 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.

The presence of high-pressure rocks within the same period of time also extends to the Caledonides in Greenland and Scandinavia (collisional process between Laurentia and Baltica). There are also high-pressure rocks in Iberia and France, but they are related to the collision of Armorica against Baltica during the late Paleozoic (Strachan et al., 1995; Dallmeyer et al., 1986; Robinson et al., 1988) (Figure 4. 4).

The described geologic framework of the peri-Gondwanan region of the northern

Appalachians is a very attractive model, which matches some of the main features and relationships between the Piaxtla and Petlalcingo groups. Thus, a possible location of the Acatlán Complex could be southward of the Brunswick complex following the Caledonian high-pressure trend. However, the main difference with this region is the vergence of folding and thrusting and therefore the sense of transport: southeastward in the Dunnage-Gander relationship and northwestward in the Acatlán Complex units.

Silurian granitoids have been reported along the length of the Appalachians and extending to the Caledonian orogen, therefore the correlation with the Esperanza granitoids could be possible. This magmatic event has been bracketed between *ca.* 440 and 410 Ma and related to the Silurian orogeny (Dunning et al., 1990; Hibbard, 1994). In northwestern South America there is litlle information about this Silurian magmatism. However, the available data place this event along the boundary between the eastern Central Andean Province and the Guyana Shield (Forero-Suárez, 1990).

Even though it is well known that characterization of magmatic rocks largely depends on tectonic and petrologic factors and these vary from place to place, in the northern Appalachians, Silurian granitoids broadly represent magmas generated in late collisional to postcollisional settings (Whalen et al., 1994; Kerr, 1997). Most granites were in part derived from older crust during crustal thickening accompanying the closure of Iapetus. However, there is little vidence that mantle-derived magmas were also involved possible as the consequence of postorogenic lithospheric delamination (Colman-Sadd, 1982: Whalen et al., 1994; van Staal and de Roo, 1995). Following this type of ideas, the origin of the Esperanza Granitoids may be similar to this tectonic process.

4. 6. Tectonic significance of the Tecomate Formation

The slightly metamorphosed Tecomate Formation is made up mainly of fine-grained marine pelitic to psammitic sediments with minor conglomerates and scarce but very distinctive horizons of limestones. At the bottom, this sequence presents a bimodal suite with abundant basaltic flows, some rhyolites and scarce gabbros. Geochemistry of these magmatic rocks displays a transition from inherited island arc affinity to within plate setting. Magmas also record a transition from tholeiites to clearly transitional-alkaline rocks. In addition, the lower part of the volcanic suite involves crustal recycling because of its mildly negative or slightly positive ε_{Nd} values (- 6 to + 1.5) and late Proterozoic model ages (1.2 to 1.6 Ga). In comparison the transitional-alkaline magmas display more homogenous positive ε_{Nd} values (+ 5) and model ages averaging 0.65 Ga, which suggest the input of new mantle material.

According to the information above, the Tecomate Formation represents magmas produced in a extensional setting following a previous collisional event. It is supported by the presence of cobbles and pebbles of granitoids and metabasites of the Piaxtla Group in conglomerates of this unit. Even though the lower contact of this unit is highly mylonitic, the relationship with respect to the Piaxtla and Petlalcingo groups is unconformable. The age of this formation is Devonian by means of its stratigraphic position (overlying the 425 Ma granite and underlying unmetamorphosed Mississipian rocks), therefore it represents the postcollisional deposit of the Silurian event, which in turn was metamorphosed by the Acadian orogeny in Late Devonian.

4. 7. Regional correlation of the Tecomate Formation

Devonian rocks with a postcollisional signature are found in both the Appalachian orogen and the Andean margin. In northwestern South America, Devonian rocks are mainly distributed in Venezuela and Colombia and most are dated as late Emsian to Givetian (late Early Devonian to late Middle Devonian). Sediments in this region consist of shale and sandstone and scarce limestone deposited in a shallow marine setting (subtidal), displaying very similar faunas to those reported in the Eastern Americas Realm (Barrett and Isaacson, 1988). Throughout this region, the lowermost Early Devonian is missing and Emsian sediments overlie Cambro-Ordovician and even Silurian sedimentary or metasedimentary rocks, thus, these sequences in northwestern South America are considered to be postcollisional, deposited after the Caledonian orogeny of the Late Silurian-Early Devonian (Forero-Suárez, 1990). In turn, the top of Devonian rocks appears to be conformable with Lower Carboniferous units suggesting the absence of the Acadian orogeny.

In central and southern South America, Devonian sedimentation is related to higher paleo-latitudes and characterized by endemic Malvinckaffric- Real faunas, which are different to those in northern South America. In addition, southward of central Chile an active margin may have still existed (Barrett and Issacson, 1988, Hervé, 1988, Ramos et al., 1986).

In the southern Appalachians, the Talladega group in Alabama represents a successor basin deposit, which includes a thick bimodal volcanic package of a probable extensional setting, this sequence is affected by a later metamorphic event produced during the Acadian orogeny (Tull and Telle, 1988).

In the northern Appalachians, most Late Silurian-Early Devonian units in the Gaspé Belt and Mascarene Belt of Quebec, New Brunswick and Maine are characterized by bimodal volcanic rocks associated with marine siliciclastic and carbonate shales (Bourque et al., 1995; McLeod and McCutcheon, 1995). These coeval bimodal volcanic rocks and sediments cover all previous terranes (Humber, Dunnage, Gander and Avalon), and are attributed to an intraplate setting dominated by wrench tectonics. In these regions the Silurian orogeny is dated in the Late Silurian (Late Ludlow-Early Pridoli).

Additionally, these sequences were subsequently deformed in the Middle to Late Devonian during the Acadian orogeny, thus Lower Carboniferous sediments unconformably overlie Devonian units. Also in the Cape Ray Belt of Newfoundland, the late Early to Middle Devonian Windsor Point Group consists of terrestrial sediments and bimodal volcanics considered to have originated in an extensional setting which is supported by the volcanic geochemistry suggesting a within plate affinity (Chorlton et al., 1995).

Contrary to the Lower-Middle Devonian rocks of northern South America, in the Appalachians, rocks of this age are made up of bimodal volcanics interbedded with sediments of shallow marine or continental environment, which display within plate affinity. This lithologic association clearly defines an extensional event carried out during its deposition, therefore, the Tecomate Formation, which displays similar characteristics and age is considered to be part of this event. Later on, these rocks were intruded, deformed and in many cases metamorphosed during the Acadian orogeny.

4. 8. Tectonic evolution of the Acatlán Complex

At this time, there are many concerns about the stratigraphic control of the Acatlán Complex because no fossils have been found in the low-grade metasedimentary rocks, no isotopic ages exist in metavolcanic rocks and just a few reliable U/Pb ages from some plutons have been obtained. Nevertheless, by using the available isotopic ages, in addition to tectonic and stratigraphic relationships between units, the general evolution of the Acatlán Complex could be established, at least identifying the most important events.

Once the units have been tectonically characterized (petrologically and geochemically) and related to a specific period of time (structural and/or stratigraphically), the evolution of the Acatlán Complex is integrated to the world wide plate tectonic framework during Paleozoic time. It is evident, however, the evolution of the Acatlán Complex represents just a small region of the eastern Iapetus Ocean and the western counterpart is missing.

In order to portray the whole evolution of the Acatlán Complex a synthetic description is given for every period of time, in addition to diagrams and paleogeographic maps.

Pre-Silurian reconstruction. In the Acatlán Complex, rocks older than Silurian in age are represented by the Xayacatlán 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. Both sequences include metavolcanic rocks with MORB and OIB affinity closely associated with quartzite and schist. These 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 for the case of the Xayacatlán Formation (Meza-Figueroa, 1998).

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 northern Appalachians, the oceanic Notre Dame arc was adjacent to the Laurentia craton, whereas in the eastern margin of Iapetus another intraoceanic arc, represented by the Exploits region, was close to the Gondwanan margin (van Staal., 1994, van Staal and Roo, 1995). A similar tectonic framework could be suggested for the two older sequences of the Acatlán Complex prior to the Silurian orogeny (Figure 4. 5).

The passive margin, represented by the continental Oaxaquia microplate and the associated rocks of the Petlalcingo Group (Gondwanan margin), would be moving northwest toward a west-dipping subduction zone similar to the Gander-Avalon margin.

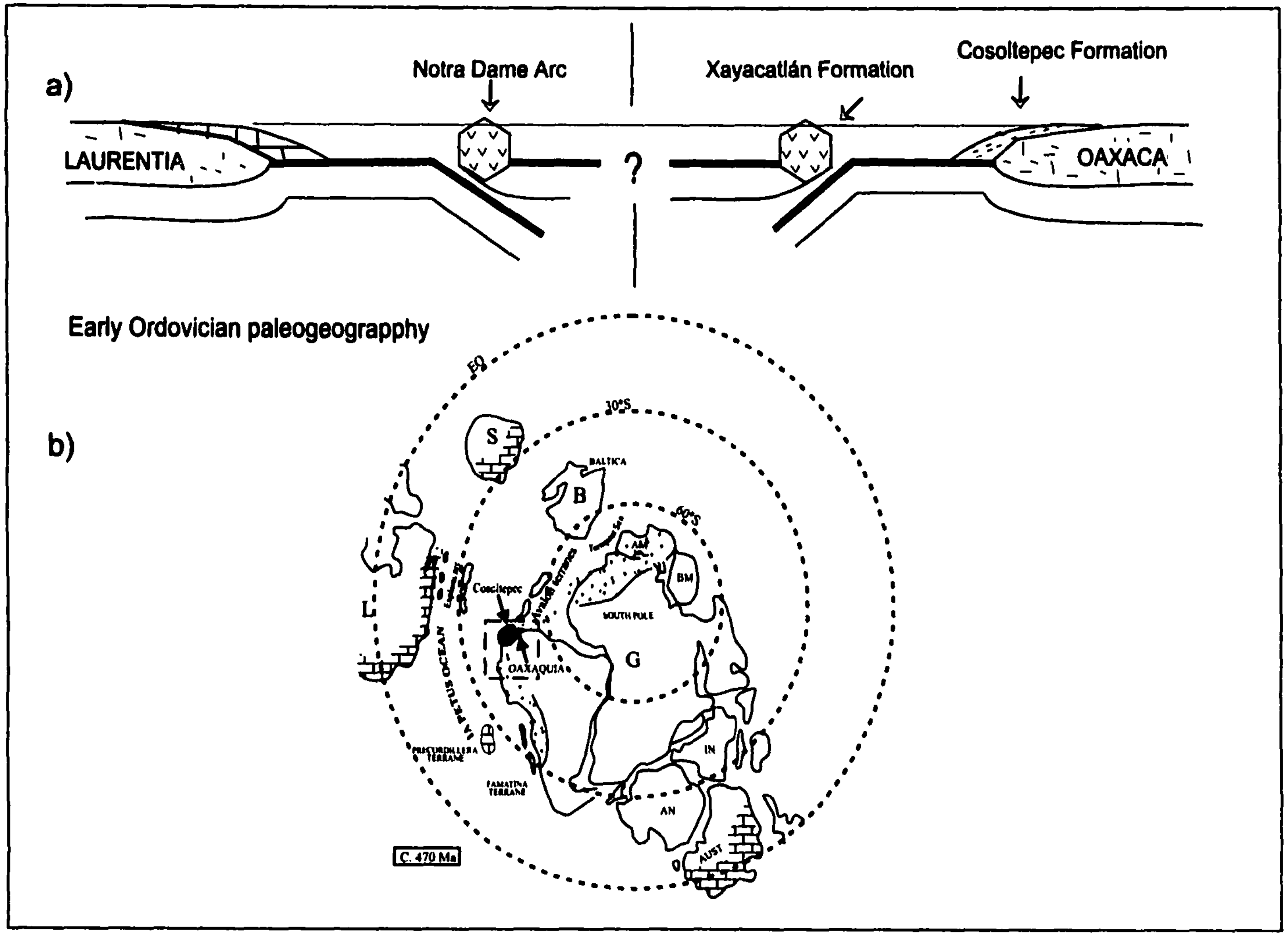


Figure 4.5 a) By the Early Ordovician the Xayacatlán Formation represented an arc-trench system, whereas the Cosoltepec Formation represented the distal deposit of the passive margin of Gondwana (Oaxaca); b)Pelogeography after Astini (1998).

Whereas the Xayacatlán Formation would represent the arc-trench system similar to the intraoceanic Exploits region.

Late Ordovician-Late Silurian reconstruction. This period of time is represented by the Esperanza Granitoids which have yielded ages ranging from 440 to 425 Ma (Robinson, 1990; Ortega-Gutiérrez et al., 1999). Although these intrusives only affected the Xayacatlán Formation, they cannot be considered arc type because of their mineralogical association, peraluminous character and crustal source, instead, they suggest a syncollisional setting. Both metabasite and metasediments of the Xayacatlán Formation and granitoids of the Esperanza group underwent an eclogite facies metamorphism and constitute the Piaxtla Group. The process of plutonism and metamorphism was accompanied by intense thrusting and the emplacement of the Piaxtla Group over the Petlalcingo Group (Figure 4. 6). Later on, the exhumation of plutonic and high-pressure rocks was carried out in a short period of time, before the deposition of the Tecomate Formation during the Early to Middle Devonian.

In the Brunswick Complex of the northern Appalachians the blueschist metamorphism concomitant with plutonism and thrusting also occurred between 447 to 416 Ma (van Staal et al., 1990) during the Silurian orogeny. This orogeny is well documented in northern Appalachians and termed Salinian orogeny by Dunning et al. (1990) or early Acadian by Hibbard (1993) to distinguish it from the classic Acadian orogeny *(sensu stricto)* of Middle to Late Devonian, in any case both authors correlate this event with the Caledonian orogeny. In northwestern South America this Silurian event is also correlated to the Caledonian Orogeny (Forero-Suárez, 1990). According to this, the Caledonian denomination is used in this analysis.

In the Appalachians, the Caledonian orogeny has been attributed to the final collision of the Avalonian terranes against Laurentia and the consequent telescoping of the entire orogen. It is supported by available paleomagnetic data (Torsvik et al., 1993). It has also been suggested that the oblique convergence of these two plates produced additional sinistral shearing (Hibbard, 1993; van Staal and de Roo, 1995). By this time, the Iapetus Ocean would be closed by the collision of the peri-Gondwanan terranes (Dunnage-Exploits, Gander, and Avalon). On the other hand, the Laurentia and Gondwana cratons would still be separated by the wide Rheic Ocean (Torsvik et al., 1993; Dalziel, 1997; Keppie and Ramos, 1999). However, this model does not explain the Caledonian event in northwestern South America.

Taken together, these data suggest that high-pressure metamorphism, thrusting and plutonism of the Piaxtla Group were part of the Silurian orogeny. No data exist, at this time, to suggest an older event to be correlated with the Taconic or Caparonensis-Ocloyic orogenesis of eastern Laurentia and western Gondwana. In order to explain the Silurian orogeny recorded by the Acatlán Complex, it has been suggested that the Petlalcingo group along with Oaxaca, were separated from the Amazonian craton drifting toward the Laurentia margin as a crustal block following the same northwest direction as the Avalonian terrane (Figure 4. 6).

This interpretation conflicts with the similarity of Silurian faunas found in Ciudad Victoria in northeastern Mexico with those faunas found in Cordillera de Mérida in Venezuela, which suggests that Oaxaquia was part of the South American continent by this time. However, the same problem is also present with Silurian fossils found in Massachusetts, Maine and New Brunswick in the United States and Canada correlatives to the same European Province of Mexico and Venezuela. Stewart et al. (1999), suggest that those terranes with Silurian faunas of European affinity (South America, Armorica, Mexico, Avalon) would be located close to Gondwana and far away from Laurentia,

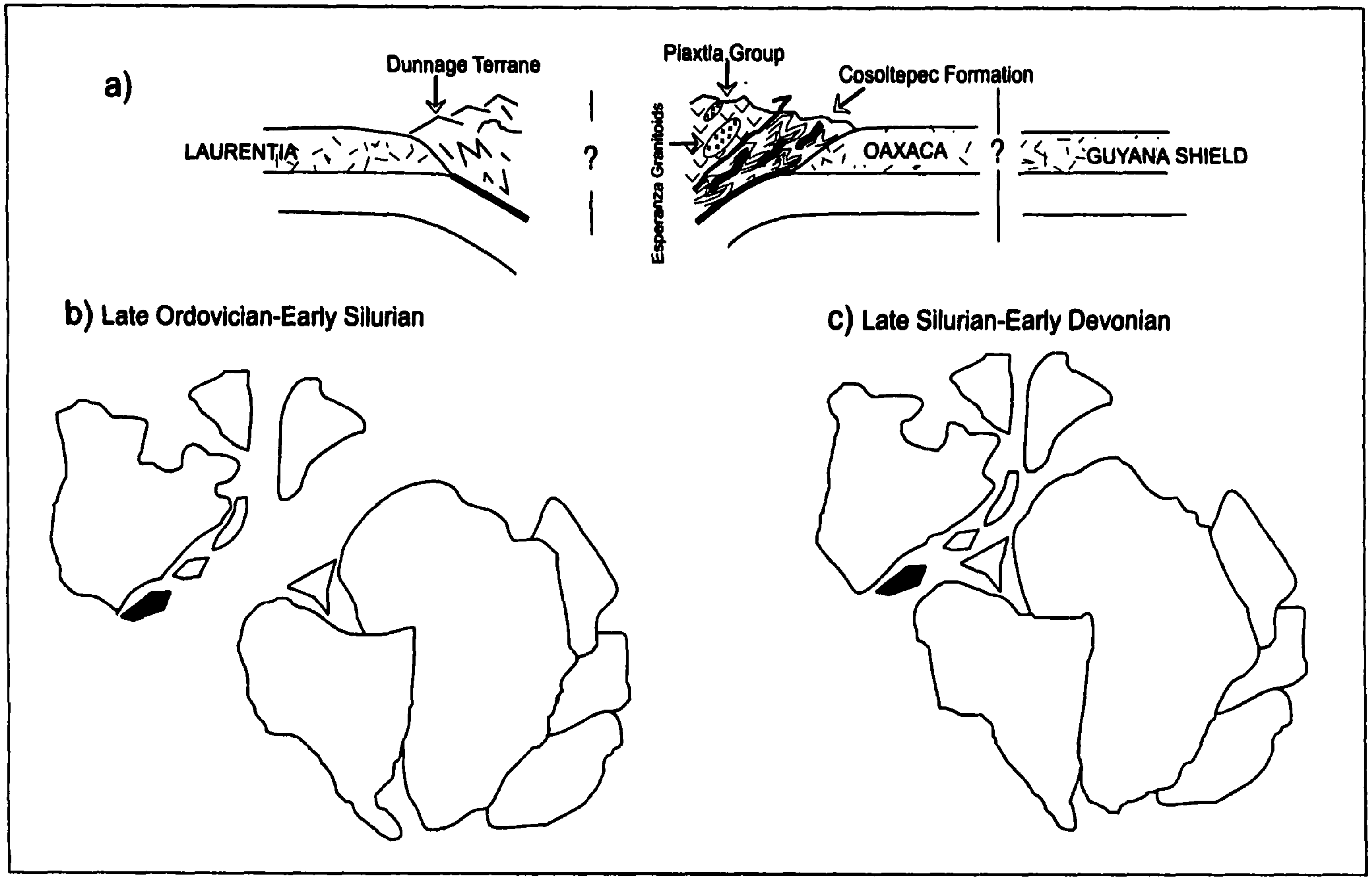


Figure 4.6 a) The Late Ordovician-Late Silurian orogeny (Caledonian) produced the emplacement of the Esperanza Granitoids, eclogitizacion and emplacement of the Piaxtla group over the Cosoltepec Formation; b) Collision of the Avalonian terranes during the Late Ordovician (after Torsvik et al., 1993), and c) Collision of Gondwana during the Late Silurian (after Van der Voo, 1988).

separated by the vast Rheic Ocean. This interpretation is inconsistent with geologic information.

If Laurentia and Gondwana were separated by this time, it could be possible that the supposed amplitude of the Rheic ocean would be narrower to allow the collision of both continents in the Late Silurian-Early Devonian as was already proposed by Van der Voo (1988). According to this model, by the Middle to Late Silurian, Oaxaquia would still be tied to the Amazonian craton or in a proximal position to share the same fauna (like the Avalonian terranes), but moving toward Laurentia to finally collide during the Caledonian orogeny in the very Early Devonian. This interpretation is very attractive because it would explain the presence of similar Silurian faunas in the Gander-Avalon terranes, Oaxaquia and Venezuela, in addition to the Silurian orogeny recorded in northern South America where Lower Devonian sediments unconformably overlie low-grade Ordovician and Silurian metamorphic rocks (Forero-Suárez, 1990). This would also explain the subsequent occurrence of similar Early Devonian faunas in southern North America and northern South America (Barrett and Isaacson, 1988).

Devonian reconstruction. Devonian rocks in the Acatlán Complex are represented by the Tecomate Formation. Although there is no paleontological record within the Tecomate Formation, its age is sufficiently constrained by means of its position and relationships, since it is overlying the 440-425 Ma intrusive rocks and underlying the well paleontologically dated and unmetamorphosed Mississipian sediments. In addition, this unit overlaps both the Petlalcingo and Piaxtla groups. Hence, the best estimated age for this unit could be Early to Middle Devonian, which is supported by the intrusion of the Hornos-Noria granite of 380 Ma. According to the available geochemical data, bimodal volcanic rocks display a within plate affinity, therefore, this unit represents a postcollisional sequence generated in an extensional setting (Figure 4. 7). This event is well documented along the length of the Appalachians, which after the Silurian collision the orogen underwent relaxation and extensional collapse, thus producing bimodal volcanism and shallow marine sedimentation. In the case of the Tecomate Formation, the bottom of the volcanic rocks displays an inherited arc signature as evidence of the pre-collisional setting.

Deformation and metamorphism of these Devonian rocks occurred in the Middle to Late Devonian time during the Acadian orogeny. In the northern Appalachians, the Acadian orogeny has been related to the accretion of the Meguma terrane or an abrupt reorganization of the Paleozoic plates switching the previous sinistral transpression to a mainly dextral transpression (Williams and Hatcher, 1983; Hibbard, 1994; van Staal and de Roo, 1995). This reorganization is also considered to have been produced by the southward movement of Gondwana along the eastern margin of Laurentia according to the model proposed by Dalziel et al. (1994) (Figure 4. 7).

Based on this model, the northwestern margin of South America would be in contact with Laurentia through dextral strike-slip motion, causing in some cases intense thrusting, metamorphism and shearing. But, Devonian rocks in Colombia are unmetamorphosed and relationships between Devonian and Carboniferous sediments appear to be conformable and transitional with no evidence of orogenic activity (Barrett, 1988; Forero-Suárez, 1990). This is contrary to the Acatlán Complex were the Tecomate Formation is intensely deformed and metamorphosed to low grade. Here, the main unconformity is in the Early Mississippian, thus bracketing the upper limit of the Acadian orogeny. In any case, it is important to note that North America and northern South America display similar Early Devonian fauna, which imply the relative proximity between both continents (Barrett and Isaacson, 1988).

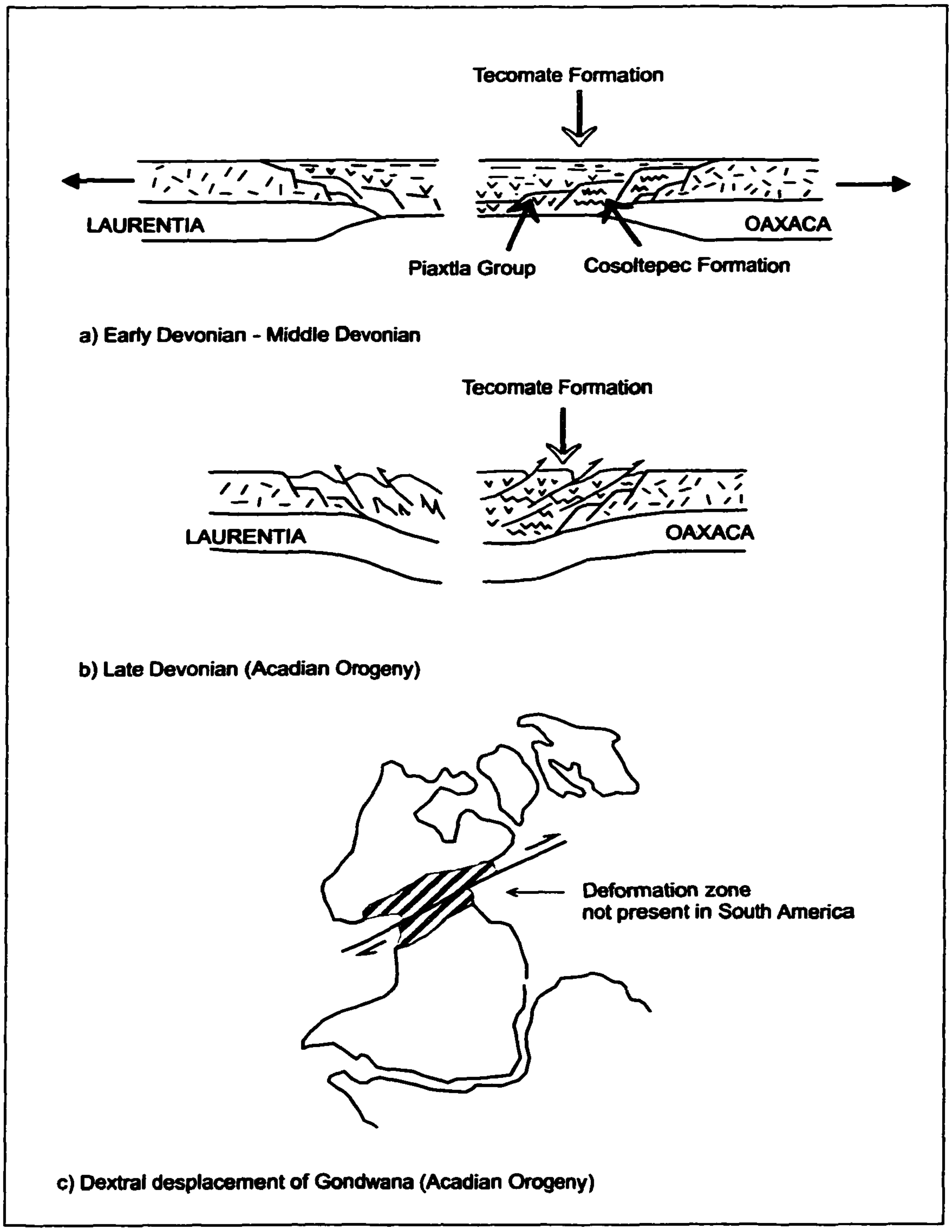


Figure 4.7 a) Deposition of the Tecomate Formation after the Silurian orogeny; b) Deformation of the Tecomate Formation during the Acadian orogeny; c) Paleogeography after Dalziel (1994).

Carboniferous to Permian reconstruction. Carboniferous to Permian rocks are represented in the Mixteco terrane by shallow marine sediments overlying the Acatlán Complex. These fossiliferous sequences consist mainly of shale, sandstone, conglomerate and limestone. Sequences with similar lithology and fauna overlie the Grenville basement therefore, these sediments constitute the first uniform sequence overlapping the Oaxaquia block. Rocks of these periods are widely distributed in southern and northeastern Mexico with faunas similar to those of North America suggesting that all these regions were in proximity and the ocean was practically closed. Gondwana was still moving southward to finally collide against Laurentia by Middle Permian during the Alleghanian orogeny to form the Pangea supercontinent. It is possible that Gondwana along with the Oaxaquia block were moving to their final emplacement in the southernmost margin of Laurentia during the Alleghanian orogeny.

DISCUSSION AND CONCLUSIONS

Discussion. A previous model for the evolution of the Acatlán Complex suggests that it was the consequence of the closure of the Iapetus Ocean because of the collision of two continental plates (Ortega-Gutiérrez, 1975). To the east the overriding plate would be the Oaxaca terrane (Oaxaquia), whereas the subducting western lower plate would be buried by the thick sequence of the Petlalcingo Group, which in turn would represent the trench and forearc deposit of a intervening oceanic arc system (Ortega-Gutiérrez, 1993, 1999). According to this model, the supposed western plate would be part of a Laurentia-type margin supplying sediments to the trench system. This model fits well the structural style of all units included in the Acatlán Complex displaying a general vergence to the west (present cordinates). However, evidence of the buried western continental plate is not supported by any geologic or geophysic data, and the enormous outcrops of the Petlalcingo Group (Cosoltepec-Chazumba-Magdalena units) do not display any chaotic melange with exotic blocks to characterise this unit as a trench-fill deposit.

The model presented in this dissertation also considers the existence of an intra-oceanic arc-trench system, but with a subduction zone dipping to the west. The suggested arc system would be coeval to the deposition of the Petlalcingo Group, which in turn would represent the eastern passive margin of the Iapetus Ocean. This model implies that sedimentation derives from the Amazonian craton represented in this case by the Oaxaca terrane. Sediments included in the trench system would also derive from the same source. The strongest argument against the model suggested here is the structural style of deformation. During the collision of two blocks with a west-dipping subduction zone, the original deformation would be characterised by structures verging to the east, which is the opposite to the structures of the Acatlán Complex. However, it is possible that the present westward vergence of these structures are the result of subsequent orogenies. It is noteworthy that the present westward vergence of the Acatlán Complex is similar to the Laramide structures, where some of them include the thrusting of this complex over the Mesozoic units (e.g., Papalutla Fault). This style of deformation is also present in the eastern region of the Morelos Platform which is characterized by westward recumbent folds.

It is evident that more evidences have to be found to prove or disprove the suggested models and further studies should include:

a) Age of sedimentation. This analysis includes the busqueda of fossils within either, the most pelitic sediments or within the very scarce limestones of the Cosoltepec Formation.

b) Provenance analysis to identify the source for sediments of the Petlalcingo Group. This study should include U/Pb detrital zircon ages. If, sediments derives from Gondwana (South America) the expected ages would include evidence of Transamazonian, Grenville and Panafrican orogenies. This is the case for the Gander terrane, which plays a similar role within the evolution of the Northern Appalachians (van Staal et al., 1996).

c) Structural analysis to reconstruct the complex history of deformation of these metamorphic units. It is possible that the first deformation of the Acatlán Complex, carried out during the Silurian orogeny, was overprinted by subsequent events. This analysis should be accompained by geochronological data and P-T-t paths in order to date accurately every deformation.

Conclusions. Mapping most of 80% of the outcropping Acatlán Complex has increased the knowledge of the distribution and relationships of the units of the complex.

The most important contributions of this study derived from the fieldwork and mapping include: (1) volcanic rocks of the Tecomate Formation are identified for the first time (2) new areas of eclogites were mapped (3) different types of granitoids were identified. Besides, (4) magmatic rocks were geochemically characterized. Fortunately, the new mapping is being used by other colleagues to conduct petrologic and geochronologic studies, which are now in progress opening the possibility of constraining the evolution of the Paleozoic in Mexico. Unfortunately, the present work did not cover 100 % of the Acatlán Complex because of logistic difficulties, however the study represents a significant advance related to previous studies. Further mapping should be focused on covering the remaining area.

Geochemically the Acatlán Complex was characterized, defining the tectonic setting for almost all the units. The exceptions were the Xayacatlán Formation, from the Piaxtla Group, which had been previously studied by others, and the Chazumba and Magdalena Formations from the Petlalcingo Group. From this latter group, the Cosoltepec formation was selected because it represents almost 90% of its lithology. The following conclusions can be drawn from the study:

(1) Cosoltepec Formation (Petlalcingo Group). Even though a small number of samples were analyzed, the oceanic affinity of this group was clearly established and along with geologic and isotopic data suggest a distal passive margin setting. Volcanic rocks of this formation constitute the oceanic floor where sediments with Precambrian provenance were deposited. This unit is correlated with those passive margins mainly made up of siliciclastic sediments bordering the Gondwanan margins. A more specific correlation is made with the Gander terrane of the northern Appalachians, although strong differences exist with respect to the basement. The alternative correlation is made with the western Central Andean Province in Colombia, however the type of basement in this

region is not known. Between the two possibilities, it is more possible that the Petlalcingo Group was part of the miogeoclinal zone of the northern proto-Andean margin of Gondwana placed south westward of the peri- Avalonian terranes, which is consistent with the available Early to Late Ordovician paleo-reconstructions.

(2) Esperanza Granitoids. Even though these granitoids only intrude the oceanic Xayacatlán Formation and trace elements (Nb-Y) plot them indistinctly within the WPG and VAG+Syn-ColG settings, geochemically and isotopically they are characterized as mainly S-type peraluminous intrusions derived from crustal melting. These features along with the age of intrusion (440-425 Ma) and high-pressure metamorphism which is shared with the Xayacatlán Formation, strongly suggest syncollisional magmatism generated during the Caledonian orogeny.

Magmatism was also accompanied by metamorphism and thrusting of the Piaxtla Group (Xayacatlán + Esperanza granitoids) over the Petlalcingo group. The P-T path determined for the Piaxtla Group, showing the progression to the eclogite facies during the subduction and thrusting, and retrogression to amphibolite facies during the exhumation, suggests an Alpine-type collision (Ortega-Gutiérrez, 1975, 1991; Ortega-Gutiérrez and Reyes-Salas, 1997). However, paleo-reconstructions are somewhat controversial. The Silurian orogeny could either result from the collision of the Avalonian terranes against Laurentia and the final telescoping of the resulting orogen, or the collision between Laurentia and Gondwana, which may have sandwiched the peri-Gondwanan terranes. The latter interpretation is consistent with the similarity of Silurian faunas in Venezuela, Mexico and northern Appalachians as well as the Late Silurian-Early Devonian unconformity in northern South America.

(3) Tecomate Formation. The intensive geochemical analyses conducted in volcanic rocks of this unit strongly suggest a within plate setting with magmas derived

from the mantle. This interpretation is consistent with the position of this unit overlapping both the Petlalcingo and Piaxtla groups after the orogenic event. Therefore, the Tecomate Formation represents the first postcollisional unit generated by the extensional collapse of the Silurian orogen. Subsequent southward translation of Gondwana by dextral strike-slip motion produced intense deformation and low-grade metamorphism into the Tecomate Formation during this Late Devonian event, which represents the Acadian orogeny. Small intrusions represented by the Hornos-La Noria granites accompany this event (380 Ma), which geochemically has been characterized as post tectonic or WPG. Similar units characterized by continental to shallow marine sediments interbedded with bimodal volcanic rocks are known along the Appalachian orogen but not in northern South America.

(4) Upper Paleozoic sediments. The metamorphic history of the Acatlán Complex is limited by these Lower Carboniferous to Lower Permian shallow marine sequences which represent the more extensive overlapping units all over the Paleozoic territory of Mexico. Lithology and fossils of this period of time are similar, both in terranes with Gondwanan or Laurentian provenance suggesting that by this time the assembly of Pangea was completed. In addition, very leucocratic granites (287 Ma) accompany the last period of deformation in this Paleozoic cycle.

The later break up of Pangea dispersed small blocks with this Paleozoic history throughout the eastern and southern Mexico as well as the Central American and Caribbean region during Mesozoic time.

Further studies in the Acatlán Complex should focus on dating the units with acceptable accuracy. Sediments of the Cosoltepec Formation should be studied for fossils and provenance, the former centred on graptolites in shale or conodonts in the very scarce limestone, the latter by means of dating detrital zircons. Additionally, geochronology is also needed in the basaltic rocks. The Piaxtla Group needs more dating for crystallisation age of metabasites and granites in addition to the P-T-t path. The Tecomate Formation also needs to be dated. Further studies should also be focused in determine the structural history of this metamorphic complex; this objective should also be considered of first order. In summary, timing and deformation are two of the main problems to be solved in the future.

REFERENCES

Adams, M. G., Stewart, K. G., Trupe, C. H., and Willard, R. A., 1995, Tectonic significance of high-pressure metamorphic rocks and dextral strike-slipe faulting along the Taconic suture, *in* Hibbard, J. P., van Staal, C. R and Cawood, P. A., eds., Current Perspectives in the Appalachian-Calaedonian Orogen: Geological Association of Canada Special Paper 41, p. 21-42.

Anderson, T, H., and Schmidt, V, A., 1983, The evolution of Middle America and the Gulf of Mexico-Caribbean region Sea during the Mesozoic time: Geological Society of America Bulletin, v. 94, p. 941-966.

Astini, R. A., 1998, Stratigraphical evidence supporting the rifting, drifting and collision of the Laurentian Precordillera terrane of western Argentina, *in* Pankhurst, R. J., and Rapeka, C. W., eds., The Proto-Andean Marging of Gondwana. Geological Society, London Special Publications v. 142, p.11-33.

Ballard, M., Van der Voo, R., and Urrutia-Fucugauchi, J., 1989, Paleomagnetic results from Grenvillian-aged rocks from Oaxaca, Mexico: Evidence for a displaced terrane: Precambrian Research, v. 42, p. 343 - 352.

Barker, F., 1979, Trondhjemite: Definition, environment and hypotheses of origin, *in* Barker, F., ed., Trondhjemites, dacites and related rocks: Elsevier, Amsterdam, p.1-12.

Bahlburg, H., and Breitkreus., 1991, Paleozoic evolution of active margin basins in the southern Central Andes (northwestern Argentina and northern Chile): Journal of South American Earth Sciences, vol. 4, no.3, p. 171-188.

Barret, S. F., 1988, The Devonian system in Colombia, *in* McMillan, N. J., Embry, A. F., Glass, D. J., eds., Devonian of the World, v. 1, p. 705-717.

Barrett, S. F., and Isaacson, P. E., 1988, Devonian Paleogeography of South America, *in* McMillan, N. J., Embry, A. F., Glass, D. J., eds., Devonian of the World, v.1, p. 655-667.

Benedetto, J. L., 1998, Early Paleozoic braquiopods and associated shelly faunas from western Gondwana: their bearing on the geodynamic history of the Pre-Andean margin, *in* Pankhurst, R. J., and Rapeka, C. W., eds., The Proto-Andean Marging of Gondwana: Geological Society of London Special Publications, v. 14, p. 57-83.

Benedetto, J. L., and Ramírez-Puig, E., 1982, La secuencia sedimentaria Precámbrico-Paleozoico Inferior pericratónica del extremo norte de Sud-América y sus relaciones con las cuencas del norte de Africa: Quinto Congreso Latinoamericano de Geología, Buenos Aires, Argentina, Actas II, p. 411-425.

Botero, G. A., 1940, Geología sobre el ordoviviano de Antioquia: Minería (Medellín), v. 16 (97-98), p. 41-82.

Boucot, A., Jhonson, J. G., and Shagam, R., 1972, Braquiopodos del Silúrico de los Andes de Mérida: Memoria, Cuarto Congreso Geológico Venezolano, Tomo II. Boletín de Geología. Publicación Especial 5, p. 585-727.

Boucot, A. J., Blodgett, R. B., and Stewart, J. H., 1997, European Province Late Silurian brachiopods from the Ciudad Victoria area, Tamaulipas, northeastern Mexico, *in* Klapper, G., Murphy, M. A., and Talent, J. A., eds., Paleozoic Sequence Stratigraphy, Biostratigraphy, and Biogeography: Studies in Honor of J. Grenville ("Jess") Johnson: Boulder, Colorado: Geological Society of America Special Paper 321, p. 273-293.

Bourque, P. A., Brisebois, D., and Malo, M., 1995, Gaspé belt; *in* H. Williams., ed., Chapter 4 of Geology of the Appalachian-Caledonian Orogen in Canada and Greenland, Geology of Canada: Geological Survey of Canada, no. 6, p. 316-351.

Brunner, P., 1987, Microfacies y microfósiles de las rocas carbonatadas del Paleozoico de San Salvador Patlanoaya, Puebla, México. Revista Sociedad Mexicana de Paleontología, v. 1. p.98-112.

Bullard, E.C., Everett, J. E., and Smith, A. G., 1965, The fit of continents around the Atlantic, *in* Blackett, P.M.S., Bullard, E. C., and Runcorn, S. K., eds., A symposium on continental drift: Royal Society of London, Philosophical Transactions, v. 258, p. 41-51.

Campa, M.F., 1985, The Mexican Thrust Belt, *in* Howell, G. D., ed. Tectonoestratigraphic terranes of the Circum-Pacific Region. Circum-Pacific Council for Energy and Mineral Resources: Earth Science Series, v. 1, p. 299-314.

Campa, M. F., and Ramírez, J., 1979, La evolución geológica y la metalogenesis del noroccidente de Guerrero: Universidad Autónoma de Guerrero, Serie Técnico Científica no. 1, 100p.

Campa, M. F., and Coney, P. J., 1983, Tectono-stratigraphic terranes and mineral resource distribution in Mexico: Canadian Journal of Earth Sciences, v. 20, p. 1040-1051.

Carlson, R. W., Lugmair, W. G., and Macdougall, D. J., 1981, Columbia River flood basalt genesis and the nature of the subcontinental mantle: Geochimical and Cosmochimical Acta, v. 45, p. 2483-2499.

Cawood, P. A., van Gool, J. A. M. and Dunning, G. R., Collisional tectonics along the Laurentian margin of the Newfoundland Appalachians, *in* Hibbard, J. P., van Staal, C. R. and Cawood, P. A., eds., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada Special Paper 41, p. 283-301.

Cobbing, E. J., Ozard, M. J., and Snelling, N. J., 1977, Reconnaissance geochronology of the crystalline basement rocks of the coastal cordillera of southern Peru: Geological Society of America Bulletin, v. 88, p. 241-246.

Colman-Sadd, S. P., 1982, Two-stage continental collision and plate driving forces: Tectonopysics, v. 90, p. 263-282.

Colman-Sadd, S. P., and Swinden, H. S., 1984, A tectonic window in central Newfoundland? Geological evidence that the Appalachian Dunnage zone is allochthonous: Canadian Journal of Earth Science, v. 21, p. 1349-1367.

Colman-Sadd, S. P., Dunning, R. G. and Dec, T., 1992, Dunnage-Gander relationships and Ordovician orogeny in central Newfoundland: A sediment provenence and U/Pb age study: American Journal of Science, v. 292, p. 317-355.

Cox, K. G., Bell, J. D., and Pankhurst, R. J., 1979, The interpretation of igneous rocks: George, Allen and Unwin., eds., London.

Chappell, B. W., and White, A. J.R., 1992, I and S-type granites in the Lachlan Fold Belt: Transactions of the Royal Society of Edinburgh, Earth Sciences, v. 83, p. 1-26.

Chorlton, L. B., Williams, H., and Wilton, D. H. C., 1995, Cape Ray Belt, *in* H. Williams., ed., Chapter 4 of Geology of the Appalachian-Caledonian Orogen in Canada and Greenland, Geology of Canada: Geological Survey of Canada, no. 6, p. 397-403.

Dalmayrac, B., Laubacher, G., Marocco, C., Martínez, C., Tomasi, P., 1980, La chaine hercynienne d'amerique du sud structure et evolution d'un orogene intracratonique: Geoloische Rundschau, Bd. 69, Seite 1-21.

Dalla-Salda, L. H., Cingolani, C, A., and Varela, R., 1992, Early Paleozoic orogenic belt of the Andes in southwestern South America: result of Laurentia-Gondwana collision ?: Geology, v. 20, p. 616-620.

Dalla-Salda, L. H., López de Luchi, M. G., Cingolani, C. A., Varela, R., 1998, Laurentia-Gondwana collision: the origin of the Famatinian-Appalachian Orogenic Belt, *in* Pankhurst, R. J., and Rapela, C. W., eds.,The Proto-Andean margin of Gondwana: Geological Society of London Special Publication 142.

Dallmeyer, R. D., and Gee, D. G., 1986, $^{40}Ar/^{39}Ar$ mineral dates from retrogressed eclogites within the Baltoscandian miogeocline: implications for a polyphase Caledonian orogenic evolution: Geological Society of America Bulletin, v. 97, p. 26-34.

Dalziel, I.W. D., Dalla-Salda, L. H., and Gahagan, L. M., 1994, Paleozoic Laurentia-Gondwana interaction and the origin of the Appalachian-Andean mountain system: Geological Society of America Bulletin, v. 106, p. 243-252.

Dalziel, I. W. D., 1997, Neoproterozoic-Paleozoic geography and tectonics: revieew, hypothesis, environmental speculation: Geological Society of America Bulletin, v. 109, p. 16-42.

D' Lemos, R. S. and Holdsworth, R. E., 1995, Samarium-Neodymium Isotopic Characteristics of the Northeastern Gander Zone, Newfoundland Appalachians, *in* Hibbard, J. P., van Staal, C. R. and Cawood, P. A., eds., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada Special Paper 41, p. 239-252.

David, J., Gariepy, C., and Machado, N., 1991, U/Pb geochronology of detrital zircons from the Tetagouche Group, New Brunswick: Geological Association of Canada Program with Abstracts, v. 16, p. A28.

Debon, F., and Le Fort, P., 1983, A chemical-mineralogical classification of common plutonic rocks and associations: Transactions of the Royal Society of Edinburgh, Earth Sciences, v. 73, p. 135-149.

De Czerna, Z., 1971, Precambrian sedimentation, tectonics and magmatism in Mexico: Geologische Rundshchau, v. 60, p. 1488-1513.

De Czerna, Z., Ortega-Gutiérrez, F., Palacios-Nieto, M., 1980, Reconocimiento geológico de la parte central de la Cuenca del alto Río Balsas, estados de Guerrero y Puebla: Libro guía de la excursión geológica a la cuenca del alto Río Balsas: Sociedad Geológica Mexicana, p. 1-33.

De Paolo, D. J., 1981, A Nd and Sr isotopic study of Mesozoic calc-alkaline granitic batholiths of the Sierra Nevada and Peninsular Ranges, California: Journal of Geophysical Research, v. 86, p. 10470-10488.

Dungan, M. A., Lindstrom, M. M., McMilan, J. N., Moorbath, S., Hoefs, J., and Haskin, L. A., 1986, Open system magmatic evolution of the Taos plateau volcanic field, northern New Mexico I: The petrology and geochemistry of the Servilleta basalt: Journal of Geophysical Research, v. 91, p. 5999-6028.

Dunning, G. R., O' Brien, S. J., Colman-Sadd, S. P., Blackwood, R. F., Dickson, W. L., O' Neil, P.P., and Krogh, T.E., 1990, Silurian orogeny in the Newfoundland Appalachians; Journal of Geology, v. 98, p. 895-913.

Elías-Herrera, M., and Ortega-Gutiérrez, F., 1998, The Caltepec fault zone, southern Mexico: Devonian dextral shear interaction between the Precambrian Oaxacan and Paleozoic Acatlán basement complexes: Laurentia-Gondwana connections before Pangea, *in* Keppie, J. D., ed., IGCP # 376 Field Trip in the Mixteco and Zapoteco Terranes: Instituto de Geología. UNAM, Mexico, Program and abstracts, p. 15.

Evensen, N. M., Hamilton, P. J., and O' Nions, R. K., 1978., Rare earth abundances in chondritic meteorites: Geochimica et Cosmochimica Acta, v. 42, p. 1199-1212.

Fang, W., Van der Voo, R., Molina-Garza, R., Moran-Zenteno, D., and Urrutia-Fucugauchi, J., 1989, Paleomagnetism of the Acatlan terrane, southern Mexico:

evidence for terrane rotation: Earth and Planetary Science Letters, v. 94, p. 131-142.

Feininger, T., 1982, The metamorphic "basement" of Ecuador: Geological Society of America Bulletin, v. 93, p. 87-92.

Feininger, T., 1987, Allochthonous terranes in the Andes of Ecuador and northwestern Peru: Canadian Journal Earth of Science, v. 24, p. 266-278.

Flores de Dios, L, A., and Buitrón, E. B., 1982, Revisión y aportes a la estratigrafía de la montaña de Guerrero: Universidad Autónoma de Guerrero. Serie Técnico Científica, no.4, p.1-28.

Flores de Dios, L.A., Vachard, D., Buitrón, B. E., Silva, A., 2000, Tha Late Permian sedimentary cover (Olinala, Ihualtepec, and Yododeñe formations) of the Zapoteco and Mixteco Terranes, Mexico: Sedimentological and Paleogeographic interpretations: SEPM-IAS Research Conference on Permo-Carboniferous Carbonate Platforms and Reefs, Program and Absracts, p.65.

Fodor, R. V., 1987, Low- and high-TiO_2 flood basalts of southern Brazil: origin from picritic parentage and a common mantle source: Earth and Planetary Science Letters, v. 84, p. 423-430.

Forero-Suárez, A., 1990. The basement of the Eastern Cordillera, Colombia: An allochtonous terrane in nortwestern South America: Journal of South American Earth Science, v. 3, no (2/3), p. 141-151.

Fries, C., and Rincon-Orta, C., 1965, Nuevas aportaciones geocronológicas y técnicas empleadas en el laboratorio: Universidad Nacional Autónoma de México, Boletín del Instituto Geológico, v. 73, p. 57-133.

González-Arreola, C., Villaseñor -Martínez, B., Corona-Esquivel, R., 1994, Permian fauna of the Los Arcos Formation, Municipality of Olinalá, State of Guerrero, Mexico: Revista Mexicana de Ciencias Geológicas, v. 11, no 2, p. 214-221.

Gose, W.A., and Sanchez-Barreda, L. A., 1981, Paleomagnetic results from southern Mexico: Geofisica Internacional, v. 20, p. 163-175.

Grajales-Nishimura, J.M., 1988, Geology, geochronology, geochemistry, and tectonic implications of the Juchatengo green rock sequence, Oaxaca State. Southern Mexico (M.S. thesis): Tucson, University of Arizona, 145 p.

Grajales-Nishimura, J. M., Centeno-García, E., Keppie, J.D., Dostal, J., 1999, Geochemistry of Paleozoic basalts from the Juchatengo complex of southern Mexico: tectonic implications, Journal of South American Earth Sciences, v. 12, p. 537-544.

Halpern, M., Guerrero-Garcia, J.C., and Ruiz-Castellanos, M., 1974, Rb-Sr dates of igneous and metamorphic rocks from southeastern and central Mexico, a progress report: Union Geofisica Mexicana, Reunion anual, Resumenes, p.30-31.

Harrington, J. H. and Kay, M., 1951, Cambrian and Ordovician faunas of eastern Colombia: Journal of Paleontology, v. 25, p. 655-668.

Hawkesworth, C. J., and Vollmer, R., 1979, Crustal contamination versus enriched mantle: $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ evidence from Italian volcanics: Contribution to Mineralogy and Petrology, v. 69, p. 151-165.

Herrmann, R. U., Nelson, K. B., Ratschbacher, L., 1994, The origin of a terrane: U/Pb zircon geochronology and tectonic evolution of the Xolapa complex (southern Mexico): Tectonics, v. 13, p. 455-474.

Hervé, F., 1988, Late Paleozoic Subduction and Accretion in Southern Chile: Episodes, v. 11, p. 183-188.

Hibbard, J., 1994, Kinematics of Acadian Deformation in the Northern and Newfoundland Appalachians: The Journal of Geology, v. 102, p. 215-228.

Hibbard, J. P. and Samson, S. D., 1995, Orogenesis exotic to the Iapetan cycle in the southern Appalachians, *in* Hibbard, J. P., van Staal, C. R. an Cawood, P. A., eds., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada Special Paper 41, p. 191-205.

Hoffman, P. F., 1991, Did the breakout of Laurentia turn Gondwana inside out?: Science, v. 252, p. 1409-1412.

Horton, W.J.,Jr., Drake, A.A.,Jr., Rankin, W.D., 1989 Tectonostratigraphic terranes and their Paleozoic boundaries in the central and southern Appalachians: Geological Society of America, Special Paper 230, p. 213-245.

Jacobsen, S. B., and Wasserburg, G. J., 1980, Sm - Nd isotopic evolution of chondrite: Earth and Planetary Science Letters, v. 50, p.139 - 155.

Keppie, J, D., and Ortega-Gutiérrez, F., 1995, Provenance of Mexican Terranes: Isotopic Constraints: International Geology Review, v. 37, p. 813-824.

Keppie, J, D., and Ramos, A, V., 1999, Odyssey of terranes in the Iapetus and Rheic oceans during the Paleozoic: Geological Society of America, Special Paper 336, p. 267-276.

Kerr, A., Jenner, G. A., and Fryer, B. J., 1995, Sm-Nd isotopic geochemestry of Precambrian to Paleozoic granitoid suites and deep-crustal structure of the southeast margin of the Newfoundland Appalachians: Canadian Journal of Earth Sciences, v. 32, p. 224-245.

Kerr, A., 1997, Space-time composition relationship among Appalachian-cycle plutonic suites in Newfoundlan, *in* Sinha, A. K., Whalen, J. B., and Hogan, J. P., eds., The nature of Magmatism in the Appalachian Orogen, Boulder, Colorado: Geological Society of America Memoir 191, p. 193-220.

Maniar, D. P., Piccoli, M. P., 1989, Tectonic discrimination of granitoids: Geological Society of America Bulletin, v. 101, p. 635-643.

Mc Cabe, Ch., Van der Voo, R., and Urrutia-Fucugauchi, J., 1988, Late Paleozoic or early Mesozoic magnetizations in remagnetized Paleozoic rocks, State of Oaxaca, Mexico: Earth and Planetary Science Letters, v. 91, p. 205-213.

McLeod, M. J. and McCutcheon, S. R., 1995, Mascarene Belt, *in* H. Williams, ed., Chapter 4 of Geology of the Appalachian-Caledonian Orogen in Canada and Greenland, Geology of Canada: Geological Survey of Canada, no. 6, p. 354-360.

Meza-Figueroa, D., 1998, Geochemistry of intermediate temperature eclogites from the Acatlan Complex, Suthern Mexico: Ph. D thesis, University of Arizona, 201 p.

Meza-Figueroa, D., Ruiz, J., Ramírez, J., Ortega-Gutiérrez, F., 1996, Low-temperature high-pressure eclogite and eclogitic rocks from the Acatlean Complex, southern Mexico: evidence of a Paleozoic subduction zone: Geological Society of America., Abstracts with Programs, 28-7, A358.

Mojica, J., Villarroel, C., and Macía, C., 1989, Nuevos afloramientos fosiliferos del Ordovicico medio (Fm. El Hígado) al oeste de Tarqui, Valle Superior del Magdalena (Huila, Colombia): Geología Colombiana, v. 16, p. 51-111.

Murphy, J. B., Nance, R. D., Keppie, J. D., Dostal, J. and Cousens, B. L., 1995, Odyssey of West Avalonian: Isotopic Constraints for Late Proterozoic III-Early Silurian Paleogeography, *in* Hibbard, J. P., van Staal, C. R. and Cawood, P. A., eds., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada Special Paper 41, p. 227-238.

Murphy, J. B., Keppie, J. D., Dostal, J. and Nance, R. D., 1999, Neoproterozoic-early Paleozoic evolution of Avalonia, *in* Ramos, V. A., and Keppie, J. D., eds., Laurentia-Gondwana Connections before Pangea: Boulder, Colorado: Geological Society of America Special Paper 336, p. 253-266.

Neuman, R. B., 1984, Geology and paleobiology of islands in the Iapetus Ocean: Review and implications: Geological Society of America Bulletin, v. 94, p. 1118-1201

Niocaills, C. M., van der Plujin, B. A., Van der Voo, R., 1997, Ordovician paleogeography and the evolution of the Iapetus ocean: Geology, v. 25, p. 159-162.

Ortega-Gutiérrez, F., 1974, Nota preliminar sobre las eclogitas de Acatlán, Puebla: Boletín Sociedad Geológica MexicanaBoletín, v. 35, p. 1-6

Ortega-Gutierrez, F., 1975, The pre-Mesozoic geology of the Acatlan area, south Mexico: Ph. D. thesis, University of Leeds, England 166 p.

Ortega-Gutierrez, F., 1978, Estratigrafía del Complejo Acatlan: Revista del Instituto de Geología UNAM, Mexico, v. 2, p. 112-131.

Ortega-Gutierrez, F., 1981a, La evolucion tectonica premisisipica del sur de Mexico: Revista del Instituto de Geología UNAM, v. 5, p. 140-157.

Ortega-Gutiérrez, F., 1981b, Metamorphic belts of southern Mexico and their tectonic significance: Geofísica Internacional, v. 20, p. 177-202.

Ortega-Gutierrez, F., 1991, Mineralogy and petrology of the high-pressure metamorphic event in the Acatlan Complex, Southern Mexico: Pachuca, Hidalgo, Convención Evolución Geológica de México: Instituto de Geología UNAM, Memoria, p. 141-143.

Ortega-Gutierrez, F., 1993, Tectonostratigraphic interpretations of the Paleozoic Acatlan Complex of southern Mexico and the problem of its regional correlations: First

Circum-Pacific and Circum-Atlantic terrane Conference, Guanajuato Mexico. Memoir, p.54-60

Ortega-Gutiérrez, F., Ruiz, J., Centeno-García, E., 1995, Oaxaquia- a Proterozoic microcontinent accreted to North America during the Late Paleozoic: Geology, v. 23, p. 1127-1130.

Ortega-Gutiérrez, F., and Reyes-Salas, A., 1997, Petrology of two high-pressure garnet megacrysts in the eclogitic rocks of the Acatlán Complex: a study case with limited X-Ray energy dispersive techniques: Actas INAGEQ, v. 3, p. 289-304.

Ortega-Gutiérrez, F., Elías-Herrera, M., and Sánchez-Zavala, J.L., 1997, Tectonic Evolution of the Acatlán Complex: New Insights, *in* II Convención sobre la evolución geológica de México y recursos asociados. Pachuca, Hidalgo. Simposia y Coloquio: Universidad Autónoma de Hidalgo e Instituto de Geología UNAM., p. 19-21.

Ortega-Gutiérrez, F., Elías-Herrera, M., Reyes-Salas, M., Macías-Romo, C., López, R., 1999, Late Ordovician-Early Silurian continental collisional orogeny in southern Mexico and its bearing on Gondwana-Laurentia connections: Geology, v. 27; no.8; p. 719-722.

Pantoja-Alor, A, J., 1970, Rocas sedimentarias Paleozoicas de la región centro septentrional de Oaxaca: Sociedad Geológica Mexicana. Libreto Guía de la Excursión México-Oaxaca, p. 67-84.

Patchett, P. J., and Ruiz, J., 1987, Nd isotopic ages of crust formation and metamorphism in the Precambrian of eastern and southern Mexico: Contribution to Mineralogy and Petrology, v. 96, p. 523-528.

Pearce, J.A., 1982, Trace element characteristics of lavas from destructive plate boundaries, *in* Thorpe, R,S., ed., Andesites: J. Wiley and Sons, Chichester, p. 525-547.

Pearce J.A., 1983, Role of the sub-continental lithosphere in magma genesis at active continental margins, *in* Hawkesworth, C. J., and Norry, M. J., eds., Continental basats and mantle xenoliths: Shiva Publishing limited, p. 230-249.

Pearce, J.A., 1996, A User Guide to Basalt Discrimination Diagrams, *in* Wyman, D.A., ed., Trace Element Geochemistry of Volcanic rocks: Applications for Massive Sulphide Exploration: Geological Association of Canada. Short Course Notes, v. 12, p. 79-113.

Pearce, J. A., and Cann, J. R., 1973, Tectonic setting of basic volcanic rocks determined using trace element analyses: Earth and Planetary Science Letters, v. 19, p. 290-300.

Pearce, J. A., and Norry, M.J., 1979, Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks: Contributions to Mineralogy and Petrology, v. 69, p. 33-47.

Pearce, J. A., Harris, N. B. W., and Tindle, A. G., 1984, Trace element discrimination diagrams for the tectonic interpretation of granitic rocks: Journal of Petrology, v. 25, p. 956-983.

Pearce, J.A., Bender, J.F., De Long, S.E., Kidd, W.S.F., Low, P.J., Guner, Y., Saroglu, F., Yilmaz, Y., Moorbath, S., and Mitchell, J. G., 1990, Genesis of collision volcanism in eastern Anatolia, Turkey: Journal of Volcanological and Geothermal Research, v. 44, p. 189-229.

Prestvik, T., and Goles, G. G., 1985, Comments on the petrogenesis and the tectonic setting of Columbia River Basalts: Earth and Planetary Science Letters, v. 72, p. 65-73.

Pindel, J, L., and Dewey, J F., 1982, Permo-Triassic reconstruction of western Pangea and the evolution of the Golf of Mexico/Caribbean region: Tectonics, v. 1 p. 179-211.

Powell, J.T., Keppie, J.D., Ortega-Gutiérrez, F., Nance, R.D., 1998, Structural history of the Magdalena Migmatite, Acatlán Complex, Southern Mexico: Laurentia-Gondwana connections before Pangea, *in* Keppie, J. D., ed., IGCP # 376 Field Trip in the Mixteco and Zapoteco Terranes: Instituto de Geología UNAM, Mexico. Program and abstracts, p. 28.

Ramos, V.A., Jordan, T.E., Allmendinger, R.W., Mpodozis, C., Kay, S., Cortés, J.M. and Palma, M.A, 1986, Paleozoic terranes of the Central Argentine-Chilean Andes: Tectonics, v. 5, no 6, p. 855-880.

Ramos, V.A., 1988, Late Proterozoic-Early Paleozoic of South America- a collisional history: Episodes, v. 11, no. 3, p. 168-174.

Ramos, V. A., Dallmeyer, R. D. and Vujovich, G. I., 1998, Time constrains on the Early Paleozoic docking of the Precordillera central Argentina, *in* Pankhurst, R. J., and Rapeka, C. W., eds., The Proto-Andean Marging of Gondwana: Geological Society of London Special Publications v. 142, p.11-33.

Rast, N., 1989, The evolution of the Appalachian chain, The Geology of North America. Vol. A, The Geology of North America - An overview: Geological Society of America, Chapter 12, p. 323-348.

Ratschbacher, L., Riller, U., Meschede, M., Herrmann, R.U., Frisch, W., 1991, Second look at suspect erranes in Southern Mexico: Geology, v. 19, p. 1233-1236.

Restrepo, J.J. and Toussaint, J. F., 1988, Terranes and continental accretion in the Colombian Andes: Episodes, v. 11, no. 3, p.189-193.

Restrepo-Pace, P. A., 1992, Petrotectonic Characterization of the Central Andean terrane, Colombia: Journal of South American Earth Science, v.5. no 1, p. 97-116.

Restrepo-Pace, P. A., 1995, Late Precambrian to Early Mesozoic tectonic evolution of the Colombian Andes, base on new geochronological, geochemical and isotopic data: Ph.D thesis, University of Arizona, 195p.

Restrepo-Pace, P. A., Ruiz, J., Gehrels, G., and Cosca, M., 1997, Geochronology and Nd isotopic data of Grenville-age rocks in the Colombian Andes: new constraints for Late Proterozoic-Early Paleozoic paleocontinental reconstructions of the Americas: Earth and Planetary Science Letters, v. 150, p. 427-441.

Riller, U., Ratschbacher, L., Frisch, W., 1992, The tierra Colorada deformation zone: Left-lateral transtention along the northern boundary of the Xolapa complex, Southern Mexico: Journal of South American Earth Science, v. 5, p. 237-249.

Robinson, J.R., Tracy, R. J., Santallier, D. S., Andreasson, P. G., Gil-Ibarguchi, J. I., 1988, Scandian-Acadian-Caledonian *sensus strictu* metamorphism in the age range 430-360 Ma, *in* Harris, A.L., and Fettes, D. J., eds., The Caledonian-Appalachian Orogen: Geological Society Special Publication 38, p. 453-467.

Robinson, K. L., Gastil, R. G., Campa, M.F., and Ramirez, J., 1989, Geochronology of basement and metasedimentary rocks in southern Mexico and their relations to metasedimentary rocks in peninsular California: Geological Society of America. Abstracts with programs, v. 21, p. 135.

Robinson, L. K., 1990, U-Pb Geochronology of basament terranes and the tectonic evolution of southwestern mainland Mexico: Ms thesis, San Diego State University, 188p.

Robinson, R., and Pantoja-Alor, J., 1968, Tremadocian trilobites from Nochixtlan region, Oaxaca, Mexico: Journal of Paleontol, v. 42, p. 767-800.

Rodríguez-Torres, R., 1969, Geology of the southeast quarter of the Acatlán quadrangle, Puebla State, Mexico: Ms thesis, Cornell University, 178p.

Rodríguez-Torres, R., 1970, Geología Metamorfica del área de Acatlán, Estado de Puebla: Excursión México-Oaxaca. Sociedad Geológica Mexicana. p. 51-54.

Ruiz-Castellanos, M., 1970, Reconocimiento Geológico en el área de Mariscala-Amatitlán, Estado de Oaxaca: Excursión México-Oaxaca: Sociedad Geológica Mexicana, p. 55-66.

Ruiz-Castellanos, M., 1979, Rubidium-Strontium geochronology of the Oaxaca and Acatlán metamorphic areas of southern Mexico: Ph.D thesis, University of Texas at Dallas, 178p.

Ruiz, J., Pattchhet, P, J., and Ortega-Gutiérrez, F., 1988, Proterozoic and Phanerozoic basement terranes of Mexico from Nd isotopic studies: Geological Society of America Bulletin, v. 100, p. 274-281.

Ruiz, J., Tosdal, M. R., Restrepo-Pace, P., Murillo-Muñetón, G., 1999, Pb isotope evidence for Colombia-Southern México connections in the Proterozoic, *in* Laurentia-Gondwana Connections before Pangea: Geological Society of America Special Paper 336, p. 183-197.

Sánchez-Zavala, J. L., and Ortega-Gutiérrez, F., 1998, The Tecomate Formation and its role in the Early and Middle Devonian Tectonic Evolution of the Acatlán Complex, Southern Mexico, *in* Keppie, J. D., ed., Laurentia-Gondwana connections before Pangea, IGCP # 376 Field Trip in the Mixteco and Zapoteco Terranes: Instituto de Geología UNAM, Mexico. Program and abstracts, p. 32.

Sedlock, R., Ortega-Gutiérrez, F., Speed, R. C., 1993, Tectonoestratigraphic terranes and tectonic evolution of Mexico: Geological Society of America Special Paper 278, 153 p.

Shervais, J. W., 1982, Ti-V plots and the petrogenesis of modern and ophiolitic lavas: Earth and Planetary Science Letters, v. 59, p. 101-118.

Silva-Pineda, A., 1970, Plantas del Pensilvánico de la región de Tehuacán, Puebla: UNAM, Instituto de Geología, Paleontología Mexicana. v. 29. p. 1-47.

Silver, L.T., and Anderson, T.H., 1974, Possible left-lateral middle Mesozoic disruption of the southwestern North American craton margin: Geological Society of America. Abstracts with programs, v. 6, p. 955.

Stewart, J. H., Blodgett, R., Boucot, A. J., Carter, J. L., and López, R., 1999, Exotic Paleozoic strata of Gondwanan provenance near Ciudad Victoria, Tamaulipas, México, *in* Ramos, V. A., and Keppie, J. D., eds., Laurentia-Gondwana Connections before Pangea: Boulder, Colorado, Geological Society of America Special Paper 336. p. 227-249.

Stockmal, G. S., Colman-Sadd, S. P., Keen, C. E., Marillier, F., O" Brien, S. J., and Quinlan, G., 1990, Deep seismic structure and plate tectonic evolution of the Canadian Appalachians: Tectonics, v. 9, p. 45-62.

Strachan, R. A., Chadwick, B., Friend, C. R. L. and Holdsworth, R. E., 1995, New perspectives on the Caledonian orogeny in Northeast Greenland, *in* Hibbard, J. P., van Staal, C. R. and Cawood, P. A., eds., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada Special Paper 41, p. 303-322 .

Tarney, J., Saunders, A, D., and Weaver, D, S., 1977, Geochemistry of volcanic rocks from the island arcs and marginal basin of the Scotia Arc region, *in* Talwani, M., and Pitman, C. W., eds., Island arcs, deep sea trenches and back-arc basins: American Geophysical Union, p. 367-377.

Tolson, J. G., 1998, Deformación y Neotectónica de la Márgen Continental de Oaxaca: Datos estructurales, petrológicos y geotermobarometericos: Tesis de Doctorado UNAM., Mexico, 98 p.

Torres, R., Ruiz, J., Patchett, J., 1999, Permo-Triassic continental arc in eastern Mexico: Tectonic implications for reconstructions of southern North America, *in* Bartolini, C., Wilson, L.J., Lawton, F.T., eds., Mesozoic Sedimentary and Tectonic history of North-Central Mexico: Geological Society of America Special Paper 340, p.191-196.

Torsvik, T. H., Trench, A., Svensson, I., and Walderhaug, H., 1993, Paleogeographic significance of mid-Silurian paleomagnetic results from southern Britain-Major revision of the apparrent polar wander path for eastern Avalonia: Geophysical Journal International, v. 113, p. 651-688.

Torsvik, T. H., Smethurst, M. A., Meert, J.G., Van der Voo, R., McKerrow, W. S., Brasier, M. D., Sturt, B. A., Walderhaug, H. J., 1996, Continental break-up and collision in the Neoproterozoic and Paleozoic- A tale of Baltica and Laurentia: Earth -Science Reviews, v. 40, p. 229-258.

Trompette, R., 1994, Geology of western Gondwana (2000-500 Ma), *in* Carozzi, V. A., ed., Pan-African-Brasiliano aggregation of South America and Africa: Rotterdam, Brookfield, V. T., U S A.

Tull, J. F., and Telle, W. R., 1988, Tectonic setting of the Silurian (?)-Devonian Talladega Belt Successor Basin, Southern Appalachians, *in* McMillan, N. J., Embry, A. F., Glass, D. J., ed: Devonian of the World, v. 2, p. 209-223.

Urrutia-Fucugauchi, J., 1984, On the tectonic evolution of Mexico: paleomagnetic constraints, *in* Van der Voo, R., ed., Plate Reconstruction from Paleozoic paleomagnetism: American Geophysical Union, Geodynamics Series, v. 12. p. 29-47.

Van der Voo, R., Mauk, F. J., and French, R. B., 1976, Permian-Triassic continental configuration and the origins of Gulf of Mexico: Geology, v. 4, p. 177-180.

Van der Voo, R., 1988, Paleozoic paleogeography of North America, Gondwana, and intervening displaced terranes: Comparison of paleomagnetism with paleoclimatology and biogeographical patterns: Geological Society of America Bulletin, v. 100, p.311-324.

van Staal, C. R., 1987, Tectonic setting of the Tetagouche Group in northern New Brunswick: implications for plate tectonic models of the northern Appalachians: Canadian Journal of Earth Science, v. 24, p. 1329-1351.

van Staal, C. R., 1994, The Brunswick subduction complex in the Canadian Appalachians: Record of the Late Ordovician to Late Silurian collision between Laurentia and the Gander margin of Avalon: Tectonics, v. 13, p. 946-962.

van Staal, C. R. and de Roo, J. A., 1995, Mid-Paleozoic Tectonic Evolution of the Appalachian Central Mobile Belt in Northern New Brunswick, Canada: Collision, Extensional Collapse and Dextral Transpression, *in* Hibbard, J. P., van Staal, C. R. and Cawood, P. A., (eds)., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada, Special Paper 41, p. 367-389.

van Staal, C. R., Ravenhurst, C. E., Winchester, J. A., Roddick, J. C., and Langton, J. P., 1990, Post-Taconic blueschist suture in the northern Appalachians of northern New Brunswick, Canda: Geology, v. 18, p. 1073-1077.

van Staal, C. R., Sullivan, R. W., and Whalen, J, B. 1996, Provenance and tectonic history of the Gander Margin in the Caledonian/Appalachian orogen: Implications for the origin and assembly of Avalonian, *in* Nance, R.D., and Thompson, M.D., eds., Avalonian and related peri-Gondwanan terranes of the circum-North Atlantic: Geological Society of America Special Paper 304 p. 347-367.

Vázquez-Echavarría, A., 1986, Descubrimiento de una nueva localidad de rocas marinas del Paleozoico al suroeste del Estado de Puebla: Informe Inédito, Petróleos Mexicanos.

Villaseñor-Martínez, A.B., Martínez-Cortes, A., Contreras y Montero B., 1987, Bioestratigrafía del Paleozoico Superior de San Salvador Patlanoaya, Puebla México: Sociedad Mexicana de Paleontología Boletín, v. 1. p. 396-417.

Weber, R., and Ceballos, S., 1994, Perfil actual y perspectivas de la paleobotánica en México: Sociedad Botánica de México Boletín, v. 55. p. 141-148.

Whalen, J. B., Currie, K. L., and Chappell, B. W., 1987, A-type granites: geochemical characteristics, discrimination and petrogenesis: Contributions to Mineralogy and Petrology, v. 95, p. 407-419.

Whalen, J.B., Jenner, G. A., Hegner, E., Gariépy, C. and Longstaffe, F. J., 1994, Geochemical and isotopic (Nd, O, and Pb) constraints on granite source in the Humber and Dunnage zones, Gaspésie, Quebec, and New Brunswick: implications for tectonics and crustal structure: Canadian Journal of Earth Science, v. 31 p. 323-340.

Whalen, J. B., Jenner, G. .A., Longstaffe, F. J., and Hegner, E., 1996, Nature and evolution of the eastern margin of Iapetus: Geochemical and isotopic constraints from Siluro-Devonian granitoid plutons in the New Brunswick Appalachians: Canadian Journal of Earth Sciences, v. 33, p. 140-155.

White, W. M., and Hofmann, A. W., 1982, Sr and Nd isotope geochemistry of oceanic basalts and mantle evolution: Nature, v. 296, p. 821-825

White, W. M., and Patchett, J., 1984, Hf-Nd-Sr isotopes and incompatible element abundance in island arcs: implications for magma origins and crustal-mantle evolution: Earth and Planetary Science Letters, v. 67, p. 167-185.

Whitford, D. J., and Jzek, P. A., 1982, Isotopic constraints on the role of subducted sialic material in Indonesian island-arc magmatism: Geological Society of America Bulletin, v. 93, p. 504-513.

Whittington, H.B., and Huges, B., 1972, Ordovician Geography and faunal provinces deduce from trilobite distribution: Philosophical transactions of the Royal Society of London, ser. B, v. 263, p.235-278

Wilson, J. T., 1966, Did the Atlantic close and then re-open?: Nature, v. 211, p. 676-681

Wilson, M., 1989, Igneous Petrogenesis. A global tectonics approach: Chapman and Hall ed., p. 466.

Williams, D., 1979, Appalachian Orogen in Canada: Canadian Journal of Earth Science, v. 16, p. 792-807.

Williams, H., 1995, Temporal and spatial divisions; Chapter 2, *in* H. Williams., ed., Geology of the Appalachian-Caledonian Orogen in Canada and Greenland, Geology of Canada: Geological Survey of Canada, no. 6, p. 21-44.

Williams, H., and Hatcher, R.D. Jr., 1983, Appalachian suspect terranes, *in* Hatcher, R. D. Jr., Williams, H., and Zietz. I., eds., Contributions to the tectonics and geophysics of mountain chain: Geological Society of America, Memoir 158, p. 33-53.

Winchester, J. A., and Floyd, P. A., 1977, Geochemical discrimination of different magma series and their differentiation products using immobile elements: Chemical Geology, v. 20, p. 325-343.

Winchester, J. A. and van Staal, C. R., 1995, Volcanic and Sedimentary terrane correlation between the Dunnage zones of the Canadian Appalachians and the British Caledonides reviewed, *in* Hibbard, J. P., van Staal, C. R. and Cawood, P. A., eds., Current Perspectives in the Appalachian-Caledonian Orogen: Geological Association of Canada Special Paper 41, p. 95-114.

Wood, D. A., 1980, The application of a Th-Hf-Ta diagram to problems of tectonomagmatic classification and to establishing the nature of crustal contamination of basaltic lavas of the British Tertiary volcanic province: Earth and Planetary Science Letters, v. 50, p. 11-30.

Yañez, P., Ruiz, J., Patchett, P.J., Ortega-Gutierrez, F., Gehrels, G.E., 1991, Isotopic studies of the Acatlan Complex, southern Mexico: Implications for Paleozoic North American tectonics: Geological Society of America Bulletin, v.103, p. 817-828.

APPENDIX

Appendix 1)

Table 3.1 Geochemical data from lavas of the Cosoltepec Formation.

Table 3.2 Geochemical data from lavas of the Tecomate Formation.

Table 3.2a Geochemical data from lavas of the Tecomate Formation.

Table 3.3 Geochemical data from granitoids of the Acatlán Complex.

Table 3.4. Nd isotopic data from magmatic units of the Acatlán Complex.

Appendix 2)

Regional map of the Acatlán Complex.

Sample Rock type	CF 37 Basalt	CF 110 Basalt	CF 39 Basalt	CF 91 Basalt	CF 126 Basalt	CF 230 Basalt
SiO2	49.8	45.05	47.55	46.2	46.1	45.7
TiO2	3.8	2.48	1.46	1.38	0.94	1.28
Al2O3	14.9	11.15	14.3	15.5	12	15
Fe2O3	13.2	11.4	11.45	12.6	10.6	12.3
MnO	0.08	0.17	0.18	0.2	0.18	0.18
MgO	2.4	7.44	6.51	6.28	10.7	6.58
CaO	4.05	10.2	11.6	8.82	13.2	10.2
Na2O	6.22	3.465	2.545	3.51	1.46	2.66
K2O	0.04	0.545	0.025	0.09	0.13	0.18
P2O5	0.75	0.405	0.115	0.11	0.06	0.1
L.O.I.	3.05	5.95	3.05	4.25	3.35	3.05
Total	98.4	98.5	98.6	99	98.8	99.2
Rb	1.9	25.5	7.5	11	1.9	23
Sr	131	608	115	151	44	422
Ba	154	1760	68.5	123	77	77
Cs	1	2	0.4	7	0.4	1
Ta	3.1	1.2	0.6	0.4	0.4	0.4
Th	1.6	3.1	0.19	0.19	0.3	0.4
U	0.2	0.7	0.45	0.1	0.09	0.09
Y	38	21	25	26	16	20
Zr	299	128	74	66	41	47
Hf	8	4	2	2	1	1.4
Nb	24	37	5	4	3	4
Ni	61	119	72.5	46	132	104
Co	47	40	44	42.5	42	48
V	208	247	365.5	370	297	330
Cr	69	290	205	38	390	320
Sc	15	31	44.5	38	54	45
La	19.7	28.7	4.9	4.4	3.1	4.2
Ce	52.7	55.9	13.1	12.3	8.3	11.3
Pr	7.7	6.8	2.05	1.95	1.3	1.8
Nd	37.9	29.1	9.85	9.7	6.7	9.2
Sm	12.1	7.2	3.6	3.55	2.3	3.2
Eu	3.58	3.04	1.33	1.165	0.93	1.06
Gd	10.5	6.4	4.45	4.2	2.8	3.9
Tb	1.6	0.9	0.7	0.75	0.5	0.7
Dy	8.9	4.8	4.9	4.75	3.3	4.4
Ho	1.68	0.82	1.08	1	0.67	0.89
Er	4.2	2.3	2.95	2.85	1.7	2.7
Tm	0.6	0.3	0.4	0.45	0.3	0.4
Yb	3.3	1.5	2.8	2.85	1.7	2.3
Lu	0.42	0.22	0.43	0.45	0.25	0.37
#Mg	15.4	39.4	33.26	36.34	50.2	41.09
(La/ Yb)N	4.03	12.92	1.18	0.99	1.23	1.23
Zr/ Y	7.6	6.5	2.9	2.73	2.75	2.35
Y/ Nb	1.58	0.57	5	6.5	5.3	5

Table 3.1. Geochemical data from volcanic rocks of the Cosoltepec Formation

TECOMATE FORMATION (GROUP A)									
Sample Rock	TE 145 Basalt	TE 222 Basalt	TE 224 Basalt	TE 301 Basalt	TE 304 Basalt	TE 07 Basalt	TE 06 Rhyolite	TE 42 Rhyolite	TE 307 Rhyolite
SiO2	50.5	51.1	50.7	44.2	45.5	50.9	68.2	76	71.1
TiO2	1.49	1.19	1.27	0.3	0.5	0.858	1.45	0.991	0.241
Al2O3	14.2	15.5	15.2	15.8	19.9	15.5	12	8.54	15
Fe2O3	10.9	10	11.1	9.4	6.44	6.97	5.3	5.47	1.45
MnO	0.15	0.15	0.17	0.14	0.09	0.13	0.14	0.05	0.02
MgO	6.57	7.66	6.53	12.2	10.1	7.52	1.51	1.29	0.33
CaO	9.05	6.93	8.83	6.81	9.85	9.71	2.72	3.13	0.07
Na2O	2.98	3.41	2.7	1.53	1.57	3.49	3.94	2.03	3.62
K2O	0.94	1.51	0.5	1.8	1.41	1.34	0.91	0.31	5.27
P2O5	0.17	0.14	0.15	0.03	0.05	0.01	0.13	0.12	0.02
L.O.I.	1.8	2.5	1.95	7.25	4.65	2.2	2.3	1.2	0.9
Total	98.9	100.1	99.1	99	100.3	98.8	98.7	99.2	98.7
Rb	28	59	27	47	89	43	34	8	184
Sr	177	118	157	81.5	216	306	62	218	45.8
Ba	337	411	194	1040	267	909	282	160	1090
Cs	1	2	1	6	11	3	3	1	2
Ta	1.1	0.4	0.4	0.9	0.9	1	1.2	0.4	0.6
Th	2.4	1.6	2.2	0.4	0.4	1.19	7.6	8	18
U	0.9	0.4	0.5	0.4	0.4	0.2	2.2	2.2	4.2
Y	37	24	30	5	17	16	28	19	60
Zr	125	80	69	20	40	59	264	288	188
Hf	4	2.4	2.4	0.6	0.9	2	8	8	8
Nb	7	5	2	1.9	2	2	15	11	21
Ni	60	74	35	140	203	53	26	19	2
Co	28	34	31	52	28	20	7	6	0.9
V	317	272	308	168	135	250	118	101	13
Cr	250	200	92	200	840	110	64	42	18
Sc	37	34	37	32.7	27.4	45	12	9	5.8
La	13	9	9	2.8	2.7	4.6	26.9	18.4	63.4
Ce	30.9	22.1	21.9	6.5	7.5	11.6	62.2	40.8	133
Pr	4.2	2.9	3	1	0.9	1.7	7.4	5	16.4
Nd	19.1	14.4	15.5	4.1	3.8	8.1	30.9	20.4	60.1
Sm	6.1	4.5	4.9		1.2	2.5	7.1	5	12.4
Eu	1.65	1.31	1.36	0.5	0.47	0.98	1.55	1.18	2.02
Gd	6.3	4.8	5.5	1.5	1.4	3.1	6.2	4.3	11.2
Tb	1.1	0.8	0.9	0.3	0.25	0.5	0.9	0.7	2.1
Dy	7.2	5.2	6.1	1.8	1.6	3.5	5.6	4.3	12.4
Ho	1.55	1.02	1.33	0.36	0.34	0.75	1.12	0.9	2.59
Er	4.5	3.1	3.8	1	1	2.2	3.2	2.5	7.2
Tm	0.7	0.4	0.6	0.13	0.12	0.3	0.5	0.4	1.1
Yb	4.2	3.1	3.7	0.8		2	3.1	2.7	6.2
Lu	0.6	0.46	0.54	0.13	0.12	0.31	0.45	0.37	0.96
#Mg	37.6	43.4	37	56.2	61.1	51.89	22.17	19.1	18.53
(La/Yb)N	2.09	1.96	1.64	1.72	1.82	1.55	5.86	4.6	6.9
Ti/Y	2.41	2.92	2.53	1.41	2.99	3.21	3.1	3.12	0.24
Nb/Y	0.18	0.2	0.06	0.09	0.18	0.125	0.53	0.57	0.35
Zr/Nb	17.85	16	34.5	30	16	29.5	17.6	26.18	8.95

Table 3.2 Geochemical data from volcanic rocks of Group A of the Tecomate Formation (Bimodal magmatism).

TECOMATE FORMATION (GROUP B)											
Sample Rock	TE144 Baslat	TE 169 Basalt	TE 119 Basalt	TE 125 Basalt	TE 146 Basalt	TE 172 Basalt	TE 214 Basalt	TE 28 Basalt	TE 121 Basalt	TE 148 Basalt	TE 114 Basalt
SiO2	48.1	43	46.3	47.1	47.2	43.3	49	47.3	47.4	51.5	47.8
TiO2	1.81	1.91	1.81	2.34	2.98	2.5	2.74	2.14	2.52	2.69	2.56
Al2O3	13	15.1	11	14	13.4	13.3	13.2	11.3	13.8	13.8	13.2
Fe2O3	14.3	15.1	12.7	12.3	14	14	12.7	13.4	12.3	10.1	12.7
MnO	0.24	0.25	0.18	0.17	0.19	0.17	0.18	0.17	0.18	0.15	0.16
MgO	7.03	7.61	12.2	7.37	8.51	4.41	6.13	8.16	6.38	4.97	5.99
CaO	10.3	5.95	9.5	9.95	4.39	9.2	10.1	11.8	9.72	7.2	10.7
Na2O	2.58	3.37	1.45	2.8	3.43	4.79	2.33	1.82	3.34	5.62	2.34
K2O	0.22	0.05	0.19	0.25	0.09	0.11	0.25	0.29	0.25	0.11	0.36
P2O5	0.14	0.18	0.18	0.22	0.31	0.24	0.27	0.2	0.24	0.28	0.26
L.O.I.	2	7.6	3.05	2	3.75	7.85	2.05	1.55	2	1.95	2.35
Total	99.8	100.2	98.7	98.6	98.3	99.9	99.8	98.3	98.2	98.4	98.6
Rb	7	4	1	10	5	17	30	9	1.9	14	8
Sr	261	67.7	211	252	68	128	276	331	305	177	314
Ba	51	19	119	145	118	66	142	198	127	111	120
Cs	2	2	1	0.4	0.4	0.4	0.4	1	1	1	2
Ta	0.6	0.6	1.1	2.1	0.6	0.4	1.9	0.4	0.4	1.5	1
Th	0.8	0.6	0.8	0.6	1.7	1	1.1	1.1	1.2	0.9	1.1
U	0.4	0.4	0.1	0.3	0.4	0.5	0.4	0.4	0.8	0.5	0.7
Y	24	25	17	20	26	22	24	20	24	23	22
Zr	117	158	125	122	181	109	123	116	137	140	213
Hf	2.5	3	3	3	5	3.3	3.7	3	4	4	4
Nb	8	9	11	12	19	16	18	12	15	13	14
Ni	56	58	359	138	61	65	71	166	86	32	71
Co	41	39	55	38	33	40	34	43	34	22	32
V	416	338	310	360	379	335	374	342	378	324	363
Cr	180	110	730	300	120	66	170	510	140	19	230
Sc	46	41.7	28	27	24	27	30	35	28	21	30
La	6.5	8	9.4	11.5	16.2	13	15.2	10.2	14.3	13.7	14.6
Ce	16	17.5	23.4	29	39.8	31.6	36.8	24.9	34.5	33.4	35
Pr	2.3	2.5	3.2	4	5.6	4.2	5	3.7	4.8	4.6	5
Nd	9.9	11.5	15.2	19.2	26.7	20.6	24.1	17.3	22.4	20.8	21.1
Sm	3.4	3.6	4.7	5.7	7.3	6.2	7	4.9	6.4	6.2	5.5
Eu	1.15	1.3	1.5	1.81	2.23	1.57	2.03	1.55	1.92	1.72	1.91
Gd	4	4.3	4.5	5.2	6.7	5.7	6.6	4.8	5.8	5.8	5.5
Tb	0.7	0.8	0.66	0.8	1	0.9	1	0.8	0.9	0.9	1
Dy	4.8	5.2	4	4.7	6.3	5	5.6	4.3	5.2	5	5.6
Ho	1	1.2	0.75	0.92	1.13	0.93	1.05	0.88	1.11	0.96	1.07
Er	3	3.2	2.1	2.4	3	2.5	2.8	2.2	2.6	2.4	2.8
Tm	0.45	0.5	0.3	0.3	0.4	0.3	0.4	0.3	0.4	0.4	0.4
Yb	2.65	3.1	1.8	2	2.5	2	2.3	1.9	2.4	2.3	2.4
Lu	0.45	0.5	0.24	0.28	0.38	0.27	0.33	0.32	0.35	0.38	0.32
#Mg	32.9	34.1	49	37.46	37.8	24	32.5	37.84	34.15	32.97	32
(La/Yb)N	1.31	1.66	3.52	3.88	4.37	4.39	4.46	3.62	4.02	4.02	4.1
Ti/Y	4.56	3.46	6.38	7.01	6.87	6.81	6.84	6.41	6.29	7.01	5.29
Nb/Y	0.19	0.27	0.64	0.6	0.73	0.72	0.75	0.6	0.62	0.56	0.72
Zr/Nb	19.5	17.5	11.36	10.16	9.52	6.81	6.83	9.66	9.13	10.76	10.74

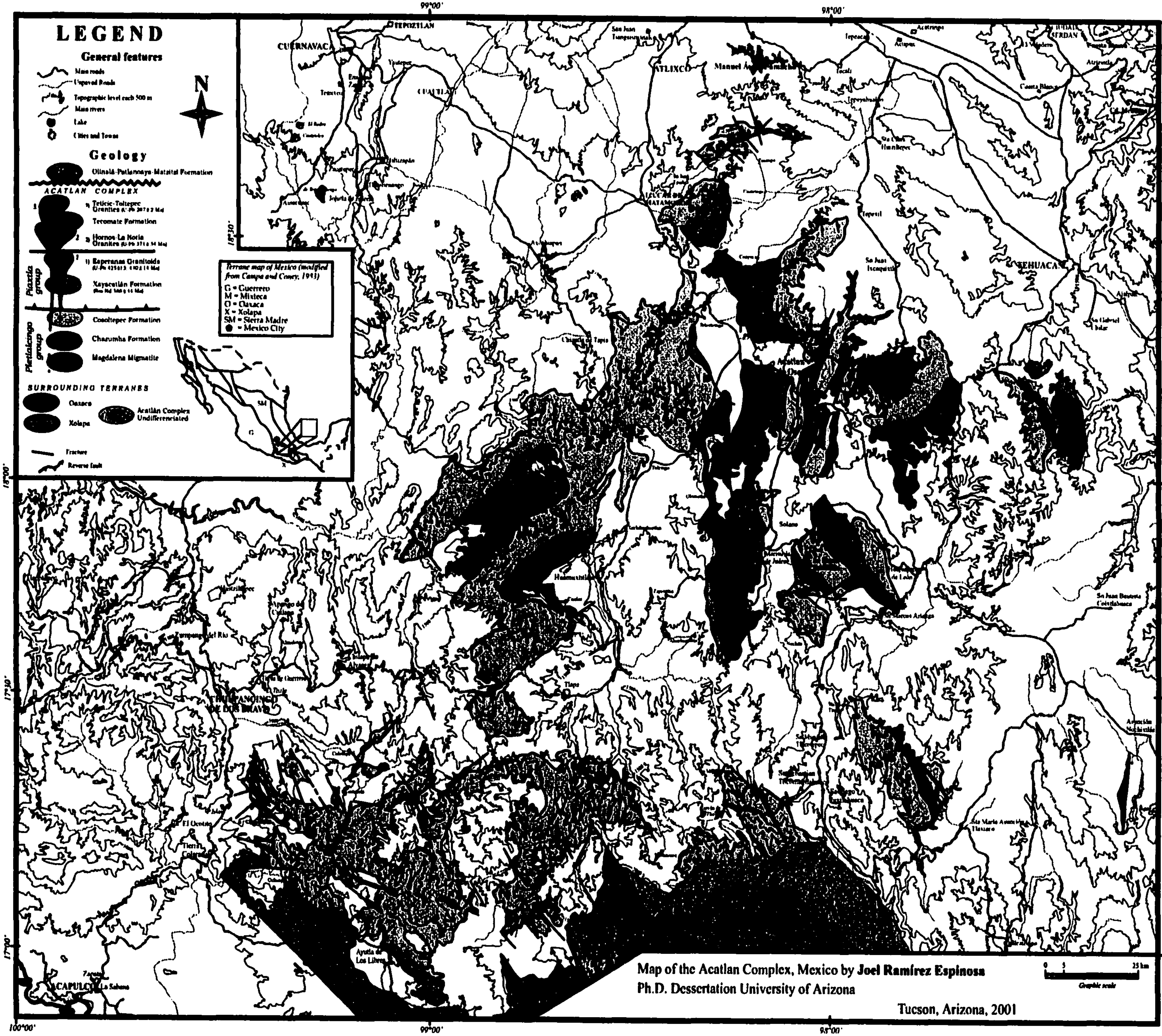
Table 3.2a Geochemical data from volcanic rocks of Group B of the Tecomate Formation (Within Plate Basalts).

	ESPERANZA GRANITOIDS										HORNOS GRANITES		TETICIC GRANITES		
SAMPLE	EG 183	EG 024	EG 142	EG 137	EG 71	EG 138	EG 185	EG 139	EG 251	EG 245	HG 101	HG 102	TG 242	TG 182	TG 150
SiO_2	73.5	71.1	69.5	69.2	69.1	68.3	66.5	56.4	48.8	46.7	68.7	69.2	81.3	79	77.4
Al_2O_3	13.3	13.7	13.6	14.3	14.3	14.1	15.2	14.4	15.1	15.2	14.1	13.9	11.1	12.1	12.5
TiO_2	0.26	0.521	0.54	0.596	0.6	0.57	0.679	2.76	1.2	1.41	0.74	0.738	0.073	0.084	0.08
Fe_2O_3	2.84	3.95	3.78	4.67	4.49	4.05	5.58	14	10.3	11.7	5.37	5.18	0.93	0.99	0.8
MnO	0.04	0.06	0.06	0.07	0.08	0.08	0.08	0.18	0.16	0.18	0.07	0.06	0.01	0.01	0.009
MgO	0.71	1.03	0.72	1.3	1.91	1.13	2.12	3.7	8.03	5.79	1.42	1.4	0.03	0.12	0.17
CaO	2.03	0.67	1.66	0.96	1.34	1.97	2.04	3	9.61	10.9	1.73	1.29	0.04	0.11	0.27
Na_2O	2.41	2.57	2.55	2.8	2.72	2.81	2.39	2.19	2.16	2.41	2.51	3.44	5.4	4.53	3.57
K_2O	3.91	4.82	4.88	4.44	2.78	4.37	3.09	2.05	1.32	0.17	4.35	4.14	0.9	3.06	4.52
P_2O_5	0.06	0.13	0.17	0.15	0.17	0.16	0.16	0.28	0.12	0.16	0.15	0.16	0.01	0.01	0.12
LOI	1.1	1.4	1.6	1.45	2.2	1.45	2.05	0.9	1.8	2	1	1.2	0.4	0.3	0.4
TOTAL	100.2	100	99.2	100	99.8	99.2	100	99.9	98.7	96.7	100.2	99.8	100.2	100.3	100
Rb	136	190	171	184	76	155	108	82	49	3	186	173	38	73	217
Sr	126	53	82.2	77	95.7	96	190	190	202	271	112	95	16	39	30.1
Cs	3	3	2	3	2	2	1	6	1	2	5	4	0.4	0.4	2
Ba	591	616	949	653	586	640	746	384	554	33	607	586	123	453	727
Th	8.8	15	19.1	11	14.4	13.7	9.4	6.8	0.7	4	14	15	7.9	10	7.5
U	3.3	3.6	3.3	2.8	2.1	1.9	1.8	1.2	0.2	4	3.5	5.2	1.8	2.1	2.6
Ta	0.8	1	3	2	0.9	3	0.4	1.7	0.4	0.9	1.1	1.7	1.6	1.2	4
Hf	4.1	6.5	8	9	7	6	7.1	9.5	1.8	3	9.5	8	2.2	1.6	3
Y	25	31	54	31	31	44	30	50	23	29	41	38	22	12	31
Zr	110	204	334	298	283	273	239	315	90	91	308	265	33	41	87
Nb	13	17	20	19	24	18	11	31	4	8	21	19	12	11	23
V	35	48	45	60	67	59	81	343	289	296	75	76	2	4	7
Cr	320	210	16	220	47	21	150	320	260	131	250	230	270	200	10
Co	2	6	5	6	6	4	10	32	40	32	9	9	1	1	0.9
Ni	11	13	9	13	22	9	40	62	115	60	18	21	7	5	2
Sc	8	9	8.4	10	9.1	9.5	12	26	36	33	12	12	3	3	2.9
La	12	34.8	49.8	33.4	36.2	37.8	39.8	45.9	5.8	8.5	43.3	41	9.1	30.6	42
Ce	19	76.6	109	69.9	74.9	78.6	83.9	98.7	14.8	20	96.7	90.4	20.4	56.2	12.4
Pr	3.1	8.8	13.9	8.3	10.2	10.6	9.8	12.2	2.2	3.2	11.3	10.6	2.7	6.3	1.5
Nd	13	35.6	49.4	35	36.7	37.3	41.3	53.2	11.6	14	45.7	43.5	10.5	22.4	5.1
Sm	3.7	8.6	11.2	8.5	8	8.5	9.3	12.9	4	4.2	10.8	10.6	3.4	4.2	1.9
Eu	1.03	1.21	1.6	1.46	1.5	1.5	1.69	2.75	1.19	1.3	1.67	1.49	0.23	0.61	0.2
Gd	3.7	7	10.2	7	6.6	7.5	7.9	11.7	4.3	4.1	9.3	9	3.4	3.2	2.1
Tb	0.6	1.1	1.2	1.1	0.8	0.9	1.2	1.8	0.7	0.6	1.4	1.4	0.6	0.5	0.4
Dy	4.5	6.8	9	6.6	5.4	7.2	6.8	10.8	4.8	4.7	8.8	8.5	4.7	2.6	4.1
Ho	1.1	1.37	1.9	1.34	1.1	1.6	1.37	2.06	1.01	1	1.73	1.65	0.92	0.51	1
Er	3.3	3.8	5.7	4	3.3	4.8	3.7	6	3	3	5.2	4.8	2.9	1.5	3.3
Tm	0.5	0.6		0.6		0.8	0.5	0.8	0.4		0.7	0.7	0.5	0.2	0.6
Yb	3.9	3.9	5	3.8	3.1	4.6	3.3	5.2	2.8	2.7	4.8	4.4	3.2	1.5	3.8
Lu	0.58	0.57	0.7	0.58	0.5	0.6	0.52	0.75	0.41	0.4	0.68	0.65	0.47	0.24	0.5
A/CNK	1.11	1.28	1.1	1.28	1.44	1.08	1.38	1.27	0.67	0.63	1.17	1.11	1.22	1.1	1.11
$(La/Yb)_N$	2.08	6.02	6.73	5.93	7.91	5.55	8.14	5.95	1.39	2.13	6.08	6.36	1.91	13.77	7.47
Ga/Al	1.7	2.21		2.11			3.48	3.02	2		2.41	2.18	1.36	1.25	

Table 3.3. Geochemical data from granitoids of the Acatlán Complex.

SAMPLE	Sm (ppm)	Nd (ppm)	147Sm/ 144Nd	143Nd/ 144Nd	Age (Ma)	ε Nd (0)	ε Nd (t)	T(DM)
TECOMATE FORMATION								
*TE-145	12.82	64.26	0.1207	0.512123	400	-10.04	-6.18	1503
*TE- 06	6.68	30.83	0.1311	0.512173	400	-9.07	-5.74	1600
*TE- 42	5.34	24.97	0.1294	0.512238	400	-7.81	-4.38	1451
*TE-301	0.78	3.39	0.1402	0.512307	400	-6.45	-3.59	1519
*TE- 222	3.89	14.49	0.1623	0.512624	400	-0.27	1.47	1239
*TE- 224	4.26	15.29	0.1687	0.512643	400	0.09	1.51	1345
**TE-119	3.99	15.51	0.1558	0.512816	400	3.47	5.57	622
**TE-214	6.11	24.6	0.1503	0.512801	400	3.18	5.54	607
**TE-172	5.13	20.39	0.1522	0.512798	400	3.12	5.38	631
**TE-146	6.77	27	0.1517	0.512795	400	3.06	5.35	634
**TE-148	5.3	20.86	0.1537	0.512742	400	2.02	4.22	781
ESPERANZA GRANITOIDS								
EG-183	3.19	19.87	0.0972	0.512144	440	-9.64	-4.05	1177
EG-137	7.36	35.58	0.125	0.512221	440	-8.14	-4.11	1410
EG-24	7.39	36.29	0.1232	0.512223	440	-8.09	-3.96	1377
EG-185	2.19	9.32	0.1425	0.512229	440	-7.97	-3.93	1734
EG-139	10.68	50.06	0.129	0.512418	440	-4.31	-0.49	1126
EG-251	3.24	11.04	0.1779	0.512755	440	2.27	3.35	1220
#AC06w	9.39	46.9	0.1211	0.512127	440	-10.01	-5.72	1500
HORNOS GRANITES								
HG-101	9.38	46.23	0.1227	0.512177	370	-8.98	-5.5	1446
#AC-18w	9.14	45.3	0.122	0.512199	370	-8.61	-5.03	1400
TETICIC GRANITES								
TG-242	3.1	11.57	0.1621	0.512317	287	-6.26	-4.99	2144
TG-63	6.05	23.25	0.1574	0.512323	287	-6.15	-4.61	1944
TG-182	5.18	18.98	0.1653	0.512659	287	0.39	1.56	1206
TOTOLTEPEC GRANITE								
#AC10w	0.31	1.71	0.1099	0.512605	287	-0.7	2.5	660
#AC21w	0.42	1.85	0.1369	0.512643	287	0.1	2.3	810

Table 3.4 Isotopic data from magmatic units of the Acatlán Complex *Group A and **Group B of the Tecomate Formation. #Data from Yáñez et al. (1991) recalculated to the shown age.



LEGEND
General features
Main roads
Unpaved Roads
Topographic level each 500 m
Main rivers
Lake
Cities and Towns
N
Geology
Olinalá-Patlanoaya-Matzitzi Formation
ACATLAN COMPLEX
Teticic-Totltepec Granites
Tecomate Formation
Hornos-La Noria Granites
Esperanza Granitoids
Xayacatlán Formation
Piaxtla group
Cosoltepec Formation
Chazumba Formation
Magdalena Migmatite
Petlalcingo group
SURROUNDING TERRANES
Oaxaca
Xolapa
Acatlán Complex Undifferenciated
Fracture
Reverse fault
Terrane map of Mexico (modified from Campa and Coney, 1983)
G = Guerrero
M = Mixteca
O = Oaxaca
X = Xolapa
SM = Sierra Madre
● = Mexico City
CUERNAVACA
TEPOZTLAN
ATLIXCO
TEHUACAN
Tlapa
Ayutla de Los Libres
ACAPULCO
El Ocotito
Tierra Colorada
99°00'
98°00'
100°00'
19°00'
18°30'
18°00'
17°30'
17°00'
Map of the Acatlan Complex, Mexico by Joel Ramírez Espinosa
Ph.D. Dessertation University of Arizona
Tucson, Arizona, 2001
Graphic scale