The Andean Belt

"The Andes is the type section of a non-collisional orogen that formed a mountain chain by subduction of oceanic crust under a continental plate... The Andean history will be divided in four major stages. The first is related to the reconstruction of the proto-margin of Gondwana, and consists of amalgamation and collision of different terranes against the Late Proterozoic margin of Gondwana. The second stage in the Late Paleozoic is linked to the formation of the Gondwanides, the first mountain chains developed along the Pacific margin by an Andean-type subduction, and the Alleghanides, related to the closing of the Iapetus ocean and the formation of the Pangea Supercontinent. The third stage is related to a generalized extension during Pangea break-up, that predates the opening of the South Atlantic and related oceans, an it is punctuated by collision of island arcs in the Northern Andes. The last stage is responsible of the present orogen, and includes a great variety of tectonic processes from collision of island arcs, seismic and aseismic ridges, as well as normal subduction of oceanic crust under the South American Plate, that defines the proper Andean-type."

(Ramos and Aleman, this volume)

Northern Andes

Antenor Aleman and Victor A. Ramos

From the time of the pioneer work of Gansser (1973) it was clear that there existed a sharp geological boundary at the latitude of the Gulf of Guayaquil (figs.1 and 2). North of this boundary, accreted oceanic terranes and pervasive strike-slip deformation played an important role in the evolution of the Andes. In contrast, the Central Andes was built on an ensialic crust with dominant orthogonal shortening. Furthermore, 200 km S of the crustal boundary of the Gulf of Guayaquil, the Huancabamba deflection marked an abrupt change in structural trend from NW to NNE (Gerth, 1955; Ham and Herrera, 1963). These two lines of evidence were used initially to establish a clear distinction between the Northern Andes and the Central Andes (Gansser, 1973; Aubouin, 1973).

Recently, this boundary has been corroborated by geophysical and geochemical studies. Paleomagnetic studies confirmed the allochthonous nature of the Piñon/Dagua Terrane (Roperch *et al.*, 1987). Bouguer gravity maps provided independent evidence to establish the oceanic and ensialic boundary (Feininger and Seguin, 1983; Mooney, 1979). In addition to the above, wide-angle seismic reflection profiles in the Western Cordillera of Colombia showed upper crustal velocities similar to those found in oceanic crust (Mooney, 1979). These geophysical data have also been used to conduct two-dimensional seismic and gravity modelling and infer the crustal structure of the Andes (Feininger and Seguin, 1983; Mooney, 1979). Preliminary geochemical studies established the Mesozoic age and suggested ocean-ridge tholeiites with higher-than-expected K, Sr, and Rb content (Barrero, 1979; Goossens and Rose, 1973; Goossens *et al.*, 1977). This was followed by studies that refined and provided alternative explanations for the origin of these mafic rocks (Lebrat *et al.*, 1985; Bourgois *et al.*, 1987; Marriner and Millward, 1984). Recently, several papers support a multiple origin for these allochthonous terranes that vary from intra-plate oceanic settings, to back-arc, island-arc, and normal MORB tholeiites (Spadea and Espinosa, 1996; Cosma *et al.*, 1998). However, it is important to note that it was the Mesozoic high pressure/low temperature metamorphic event associated with emplacement of these rocks that is unique to the Northern Andes.

The present Northern Andes may be subdivided into three main segments based on their geological and structural characteristics: the Venezuelan Andes (or Mérida Andes), the Colombian Andes, and the Ecuadorian Andes (Fig.2). Their present political divisions roughly coincide with major changes in the nature and virgation of the different mountain chains. For instance, the Colombian and Ecuadorian Andes correspond to a typical subduction-related mountain chain developed along the continental margin. On the other hand, the intracontinental Mérida Andes were formed as a result of the interaction between the Paleogene Caribbean thrusting and Neogene tectonic inversion during Andean compression. These structures were greatly affected by a complex system of strike-slip faults and folds.

Along-strike changes in the Colombian and Ecuadorian Andes resulted from variations in the Mesozoic extension as well as the fact that Western Colombia was situated along the path of the Caribbean Plate. Pronounced Mesozoic extension in Colombia accounted for the separation of the Central and Eastern Cordilleras. Late Cretaceous collision of the Piñon/Dagua Terrane caused tectonic inversion and uplift of the Central Cordillera with mild deformation in the Eastern Cordillera of Colombia. Although the Ecuadorian Andes were also affected by extension, Late Cretaceous docking of the Piñon/Dagua Terrane caused shortening and flaking of Mesozoic oceanic crust and subsequent amalgamation of the Eastern Cordillera into a single geomorphical unit. Pacific extrusion of the Caribbean Plate was associated with middle Eocene and late Miocene accretion in the Western Cordillera. Complex plate interaction prior to the formation of the Caribbean Plate left a series of small drifting terranes that were later accreted in the northern part of Western Colombia. This periodic terrane docking, enhanced by oblique convergence, caused strain partitioning and continuously reactivated old suture zones.

Because the political frontiers approximately match the geological boundaries, the Northern Andes will be described from the Venezuelan (Mérida) Andes in the N to the Colombian and Ecuadorian Andes to the S in the following sections.

Venezuelan (Mérida) Andes

Gansser (1973), in his review paper, refers to the Mérida Andes as a distinct mountain chain, separated from the main Eastern Cordillera. This distinction is based on important strike changes, and the presence of Precambrian crystalline rocks. Bucher (1952) proposed for its origin a mega-anticline, limited on both flanks by high-angle reverse faults, while Rod (1956) proposed a mushroom-like transpressional uplift with shear deformation along the Boconó Fault System. Imbricate thrusting occurs toward both Andean fronts (Deratmiroff, 1971). Kellogg and Bonini (1982) postulated that crustal shortening along a thrust fault dipping to the SE caused the uplift of the Mérida Andes. Recently, Colletta *et al.* (1997) interpreted the Mérida Andes as a transpressional intercratonic orogenic belt, developed in response to oblique convergence between two continental lithosphere blocks (Fig. 3).

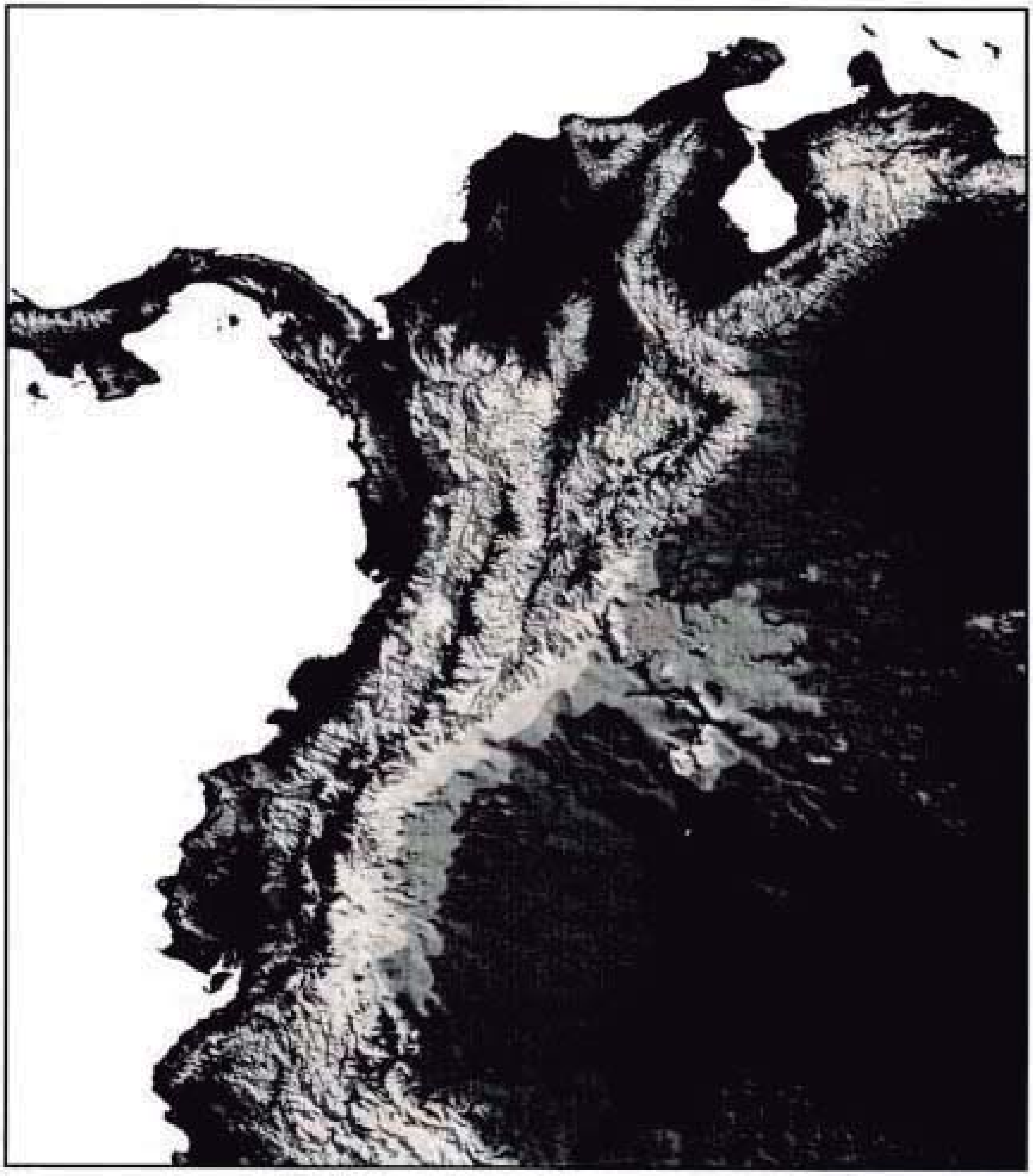
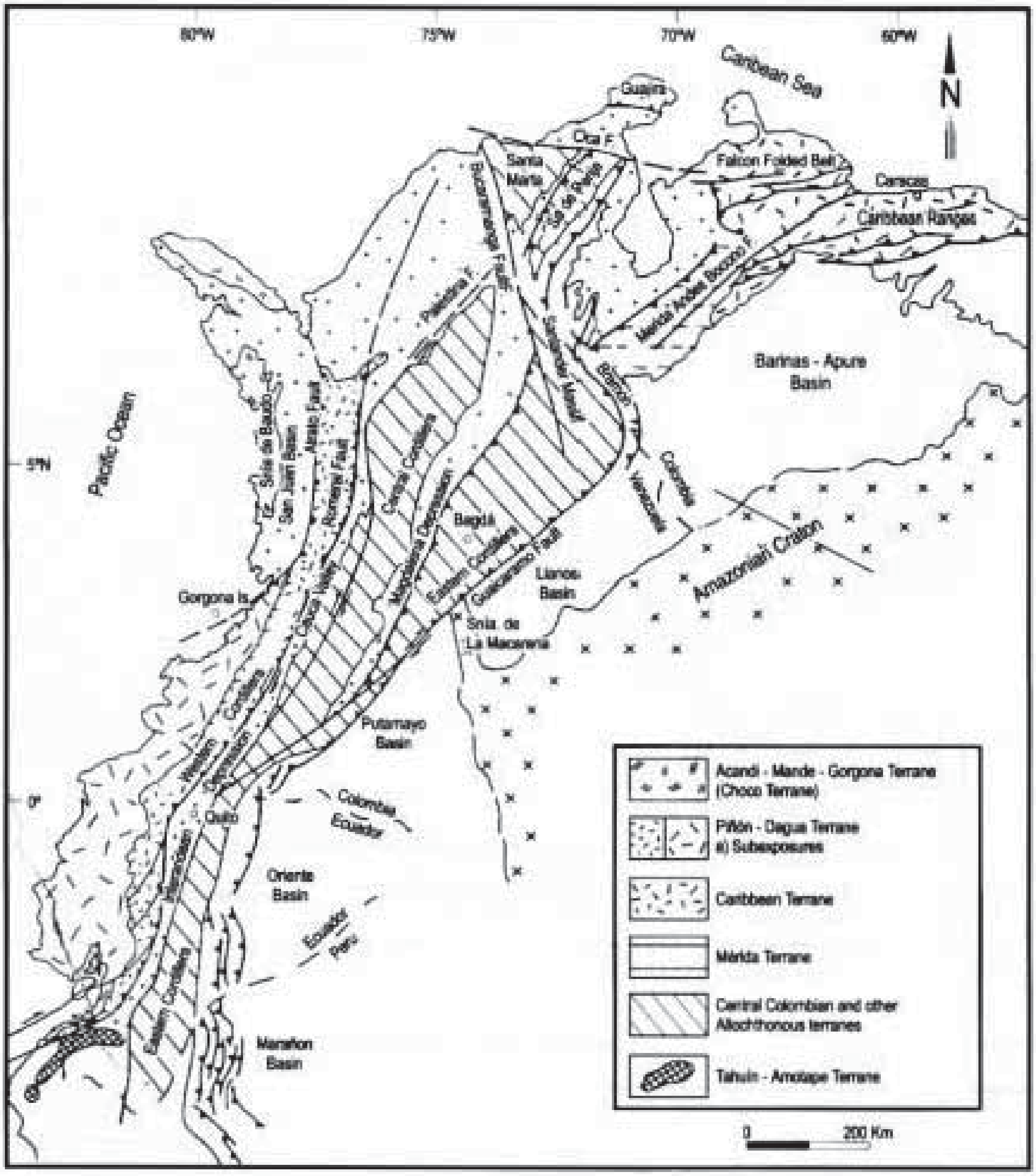


FIGURE 1 - *Digital topographic map of the Northern Andes (from USGS web site). Note the tectonic boundaries among the Santa Marta Block, the Mérida Andes, and Eastern, Central, and Western Cordilleras of Colombia and Ecuador. Compare with the main structural provinces depicted in Fig. 2.*

FIGURE 2 - *Main structural provinces and terranes of the Northern Andes (based on Colleta et al., 1997; Bourgois et al., 1987; Restrepo and Toussaint, 1988; Litherland et al., 1994).*

The last deformation episode, which created the present day Mérida Andes, commenced in the late Oligocene and continues to the present. Recent tectonics are dominated by uplift manifest by the tilting of Quaternary terraces (Meier et al., 1987), and by studies of focal mechanism (Malave and Suárez, 1993). The mountain belt is bounded on the SW by the Santo Domingo Reentrant of the Llanos Basin which abuts against the Santander Massif along the Bramon Fault (Laubscher, 1987). To the NE, the range terminates at the Barquisimeto Depression where it is replaced by the Caribbean or Coastal Ranges. Bellizzia and Rodriguez (1968) emphasized the dominant role of the Bocono Fault in the structural development of the northern Mérida Andes. Although Shagam (1975) described dominant vertical displacement along this fault, 80 km of right-lateral displacement has been reported, mostly occurring in recent times (Kellogg and Bonini, 1982; Stephan, 1982).

White (1985) published the first CDP seismic data



illustrating the structural style. Recently, Audemard (1991), De Toni and Kellogg (1993), Colletta *et al.* (1997), have made significant contributions to deciphering and understanding the structural styles of this foldbelt.

Tectono-stratigraphic evolution

Precambrian

Several units in the Mérida Andes have been ascribed to the Precambrian. Unfortunately, the isotope dates display large error brackets. The U/Pb data plotted on the concordia diagram (Burkley, 1976) were re-interpreted by Marechal (1983) who argued against ages older than the Bellavista Formation (650 - 580 Ma). These problematic units include zircon ages for the El Carmen Granite (891 ± 100 Ma), the Quiu Granite (1.061 ± 0.1 Ga), the El Cambur Granite (1.118 ± 0.25 Ga) and the Estanques Granite (1.601 ± 0.35 Ga) (Burkley, 1976). Robust U/Pb Precambrian ages have been reported for the Santander Massif to the S (Restrepo-Pace *et al.*, 1997). However, the lack of granulite precludes the presence of the Grenville Orogeny that is documented in Colombia (Alvarez and Cordani, 1980; Kroonenberg, 1982; Priem *et al.*, 1989; Restrepo-Pace *et al.*, 1997). Thus, these older ages, with large error brackets, only provide a clue to the age of the protolith.

Equally important is the large age error for the Colorado Massif, situated in the Caparo Block. Rocks from the Bellavista Formation described by Burkley (1976), gave 650 - 580 Ma (Bellizzia and Pimentel, 1994). These rocks consist of chlorite-muscovite quartz schist and metachert with subordinate amphibolite and olivine basalt. They underwent a metamorphic event in which there occurred widespread granitic plutonism and penetrative deformation associated with the El Topo Gneiss (660 Ma), the Valera Granite (593 ± 16 Ma) and the Rio Caparo Granite (615 ± 30 Ma). This

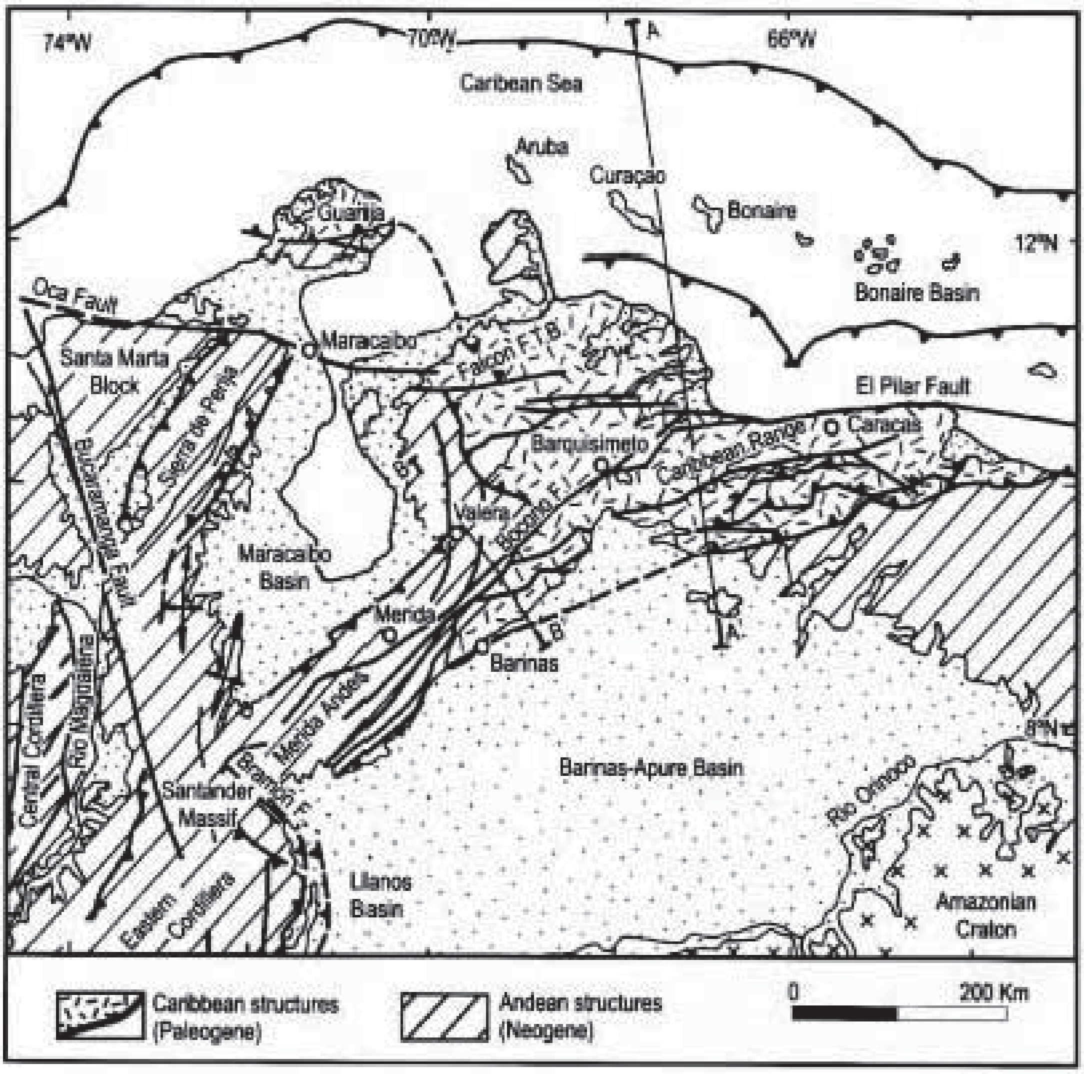
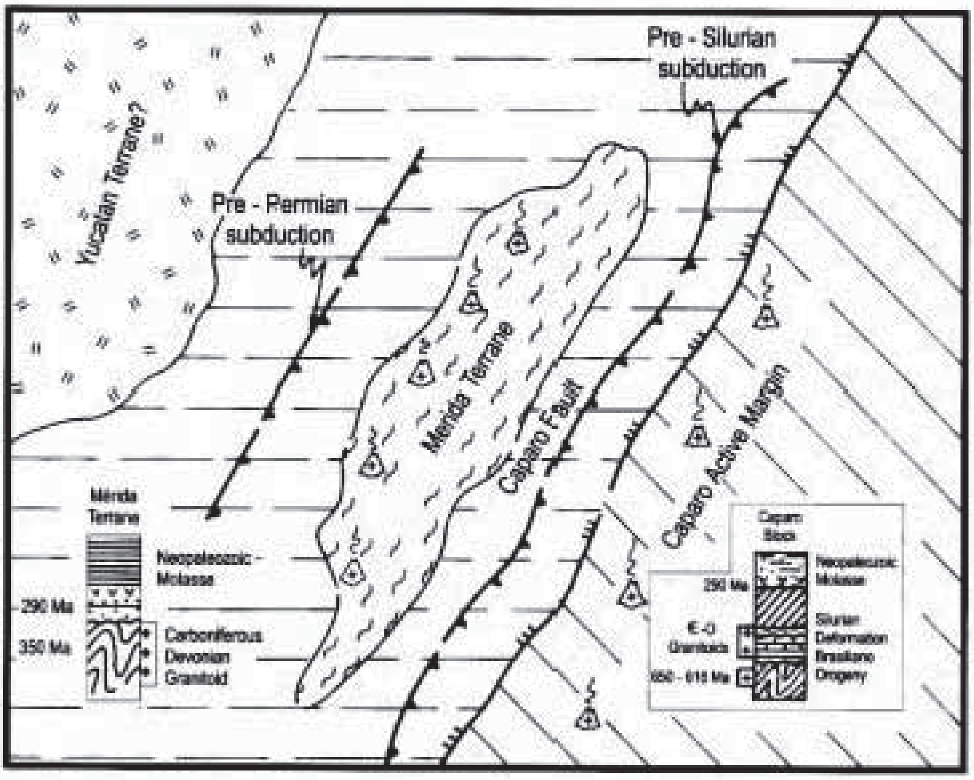


FIGURE 3 - Tectonic setting of the Mérida Andes and the Caribbean Range of Venezuela with main geologic units (based on Kellogg and Bonnini, 1982; Bosch and Rodriguez, 1992; Colleta et al., 1997).

FIGURE 4 - Paleozoic tectonic setting of the Northern Andes of Venezuela (based on Bellizzia and Pimentel, 1994; Pindel, 1985).



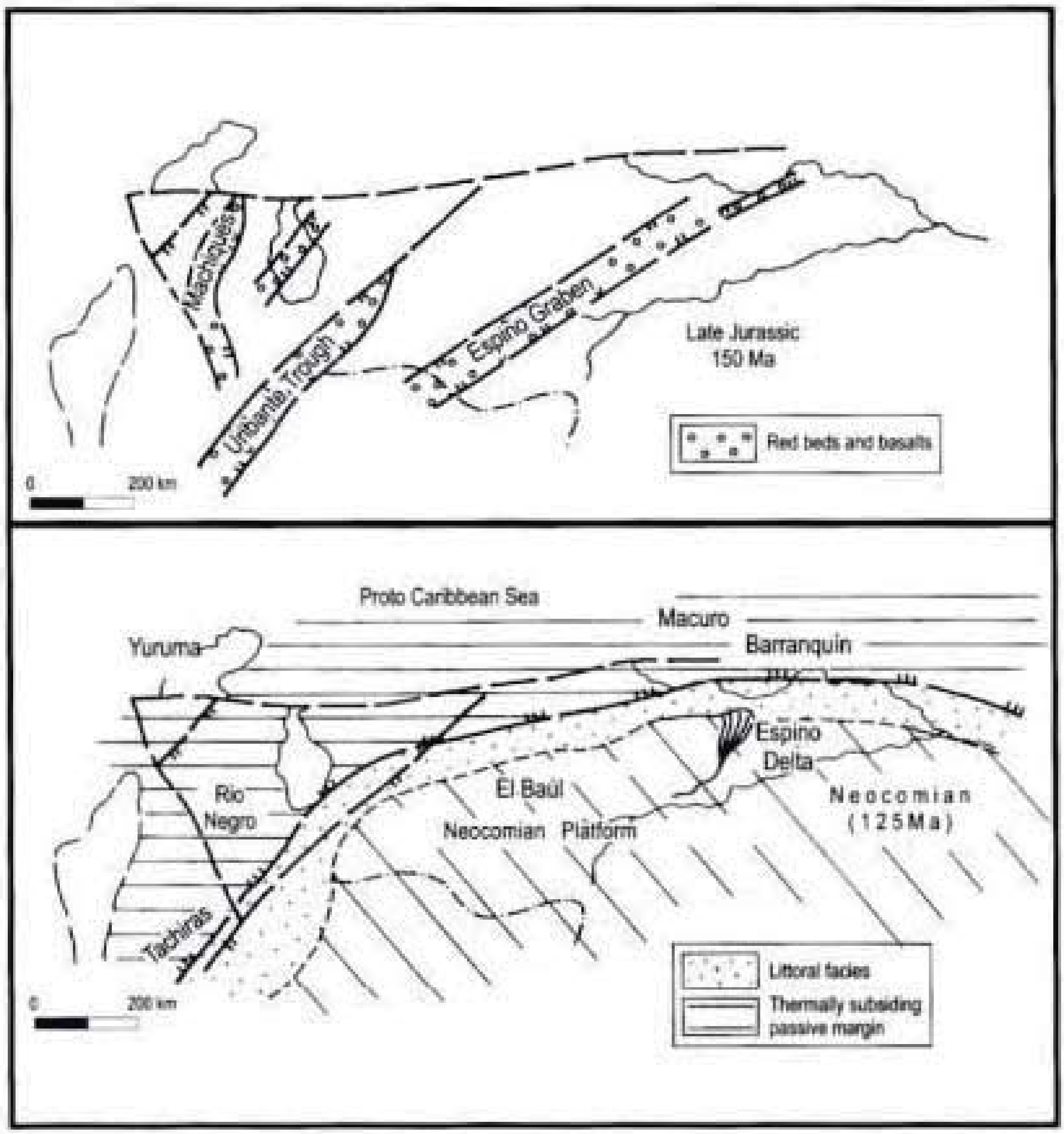


FIGURE 5 - Mesozoic paleogeography of Venezuela (based on Pindell and Barret, 1990; Bartok, 1993).

metamorphic event correlates in time with the Brasiliano Orogeny (Herz *et al.*, 1989).

Paleozoic

Bellizzia and Pimentel (1994) described two distinct geological provinces, separated by the Caparo Fault, and named them the autochthonous Caparo Block and the allochthonous Mérida Terrane (Fig. 4). Although paleomagnetic data to support this model are lacking, there are marked differences in pre-Sabaneta rocks across this fault. Thus, the Precambrian metamorphic rocks of the Caparo Block unconformably overlie 350 m of black shale, siltstone and sandstone of the Caparo Formation containing graptolites, trilobites and brachiopods of Late Ordovician age (Benedetto and Ramirez, 1982). Thinly laminated shale layers interbedded with siltstone, micaceous sandstone and siliceous limestone of the El Horno Formation in turn unconformably overlie the Caparo Formation. A rich fauna of graptolites, corals, trilobites and brachiopods (Boucout, 1972; Sánchez, 1984) indicates a Silurian age (Llandoverian to Ludlovian). These sedimentary rocks were intruded by syn-tectonic granites (Caparoensis Orogeny) from the Early Ordovician to Early Silurian (495 - 425 Ma), and by post-tectonic granites during the Devonian (Burkley, 1976; Marechal, 1983).

The basement of the allochthonous Mérida Terrane (Bellizzia and Pimentel, 1994) is made up of the Sierra Nevada Formation (Lower Paleozoic) consisting of three metamorphic facies, indicating a sedimentary protolith (Kovisars, 1971; Shagam, 1975; Bellizzia and Pimentel 1994; Marechal, 1983). These facies vary from quartz-feldspathic mica schist, sillimanite-bearing assemblages with rare staurolite, to micaceous quartz-feldspathic gneiss with rare sillimanite and garnet to hornblende-epidote amphibolite. Marechal (1983) has estimated temperatures for these rocks between 620 to 720 °C and pressures form 4.3 to 7 kbar. According to Marechal (1983), this formation changes stratigraphically upward to microfolded, bluish-grey quartz-feldspathic garnet schist, phyllite, amphibolite and metaconglomerate of the Tostosa Formation. This unit is followed by slate, thinly laminate phyllite interbedded with metasandstone, locally with conglomerate, crystalline limestone, metachert, felsic metavolcanic units and pyroclastic rocks of the Mucuchachi Formation (Bellizzia and Pimentel, 1994). The fauna and flora suggest a Carboniferous age, no older than Mississippian (Sánchez, 1984). These two units underwent deformation and metamorphism at the same time as the Sierra Nevada metamorphic facies, prior to Westphalian as determined by the concordant foliation planes (Marechal, 1983). This interpretation is based on the presence of unmetamorphosed Upper Paleozoic rocks in the Caparo Block as well as in the Mérida Terrane.

Subsequent docking of the Mérida Terrane was accompanied by molasse deposition of the Sabaneta Formation (Westphalian) consisting of interbedded red sandstone, shale and conglomerate (Thompson and Miller, 1949; Arnold, 1966; Sánchez, 1984). This formation is overlain by a sequence of interbedded sandstone, shale and fusulinid limestone of the Palmarito Formation (Stephanian to Guadalupian) (Pierce *et al.*, 1961; Sánchez, 1984). The Westphalian collision of the Mérida Terrane may have created salients and reentrants with different rheologies. Such heterogeneity controlled not only the relative subsidence rates but also the location and development of structural highs such as the orthogonal Mérida Arch (Lugo, 1994).

Mesozoic

A period of tectonic quiescence heralded the regional extension associated with deposition of varicolored beds of sandstone, shale, conglomerate and tuff of the La Quinta Formation. The Mérida Arch started to become active and continued active through the Early Cretaceous (Lugo, 1994). This was an important transfer zone during the structural development of the Mérida Andes. Jurassic grabens, trending NE-SW, (Fig. 5) are subparallel to the Uribante Trough (Renz, 1959). Felsic and basic volcanism, coeval with opening of the western Tethys Ocean and the Gulf of Mexico accompanied extension.

Passive margin sedimentation did not started until the Cretaceous (Barremian). It began with the deposition of coarse-grained, thick, cross-bedded arkosic to quartz-rich sandstone units of the Río Negro Formation (Zambrano *et al.*, 1971). The Cogollo Group (Aptian to Cenomanian) overlies this formation, which consists of interbedded limestone, shale and sandstone. Periodic flooding of the continental margin drowned the carbonate platform (Bartok *et al.*, 1981) with subsidence in the major troughs, active until the close of the Cenomanian. This accounts for the thick Lower Cretaceous section. The regional Turonian transgression was followed by periodic transgressions from the Coniacian to Early Santonian. Each transgression was marked by the deposition of black shales, rich in organic matter; and beds of limestone and chert of La Luna/Navay Formation (Tribovillard *et al.*, 1991; Parnaud *et al.*, 1995). An abrupt change in depositional style and rates occurred during Late Campanian-early Maastrichtian. Sediments consisted of thick shale units and thin bedded sandstone of the Colon/Mito Juan Formations which grade laterally to the Burguita Formation (Parnaud *et al.*, 1995), that consists mainly of sandstone.

Cenozoic

During the Caribbean Orogeny (Late Cretaceous to early Eocene), there occurred an orthogonal emplacement of allochthonous terranes and contemporaneous docking of the Central Cordillera of Colombia (Peruvian Phase), associated with the development of a dogleg-shaped foredeep. The trend of the foredeep changed from N-S to E-W on account of the orthogonal nature of these allochthonous terranes. Paleocene to early Eocene reactivation of the Mérida Arch suggests that it may have guided the collision-related thrusting of the Caribbean nappes (Lugo, 1994). Discrete unconformities within the thick molasse sequence of the Oroque Group (Paleocene) occur along the Mérida Andes, suggesting periodic orogenesis. Fluvio-deltaic deposits of sandstone, shale and coal seams of the Mirador/Gobernador and Carbonera/Paguey formations attest to a new pulse of molasse deposition during the Eocene. Marine shale and siltstone beds of the Leon Formation (Oligocene) overlie these deposits (Parnaud *et al.*, 1995).

Eastward movement of the Caribbean Plate from the middle Eocene (Pindell and Barret, 1990), provides a mechanism for the thrusting of the Carora-El Tocuyo nappes in the northern Mérida Andes (Stephan, 1982). These nappes consist of Paleocene to early Eocene flysch deposits and metamorphosed Cretaceous basinal rocks. Oligocene N-S overthrusting has been demonstrated from evidence found in the Guarumen-1S exploration well (Figueroa de Sánchez and Hernández, 1990). Such Caribbean nappes were often reactivated or refolded by younger Neogene Andean structures (Fig. 6a). Thus, an important boundary can be shown to occur across the Humocaro Fault where to the W, autochthonous and allochthonous structures have been overprinted by N-S to N45°E structures, coaxial with the Boconó Fault (Stephan, 1982). Stephan (1982) has estimated about 80 km of right-lateral displacement from the nappe offsets. Truncation of the Caribbean allochthon and foreland sequences S of the Andes suggests regional uplift at the Eocene/ Oligocene boundary.

The Miocene/Pliocene molasse in the southern foredeep, associated with deformation and uplift of the Mérida Andes (Fig. 6b), consists of fluvial conglomerate, sandstone and shale of the Parángula, Río Yuca and Guanapa formations. This southern flank shows both N and S verging basement-involved structures (Audemard, 1991; Colletta *et al.*, 1997). The equivalent molasse of the northern flexural basin has also a distinct fluvial facies of the Chama/Palmar, Isnoto and Betijote formations (Colletta *et al.*, 1997). However, the structural style consists of N-verging thrusts, mainly detached from the pre-Cretaceous substratum, that form an antiformal stack (Colletta *et al.*, 1997). Secondary detachment zones have also been reported in the Upper Cretaceous and Tertiary strata (Audemard, 1991). Multiple unconformities and fining-upward cycles in the molasse sequence record periodic deformation and uplift of this Andean segment. Indeed, a late Miocene intra-molasse unconformity heralds the beginning of the frontal monocline formation and the period of greatest uplift (De Toni and Kellogg, 1993).

According to Colletta *et al.* (1997), Plio-Quaternary

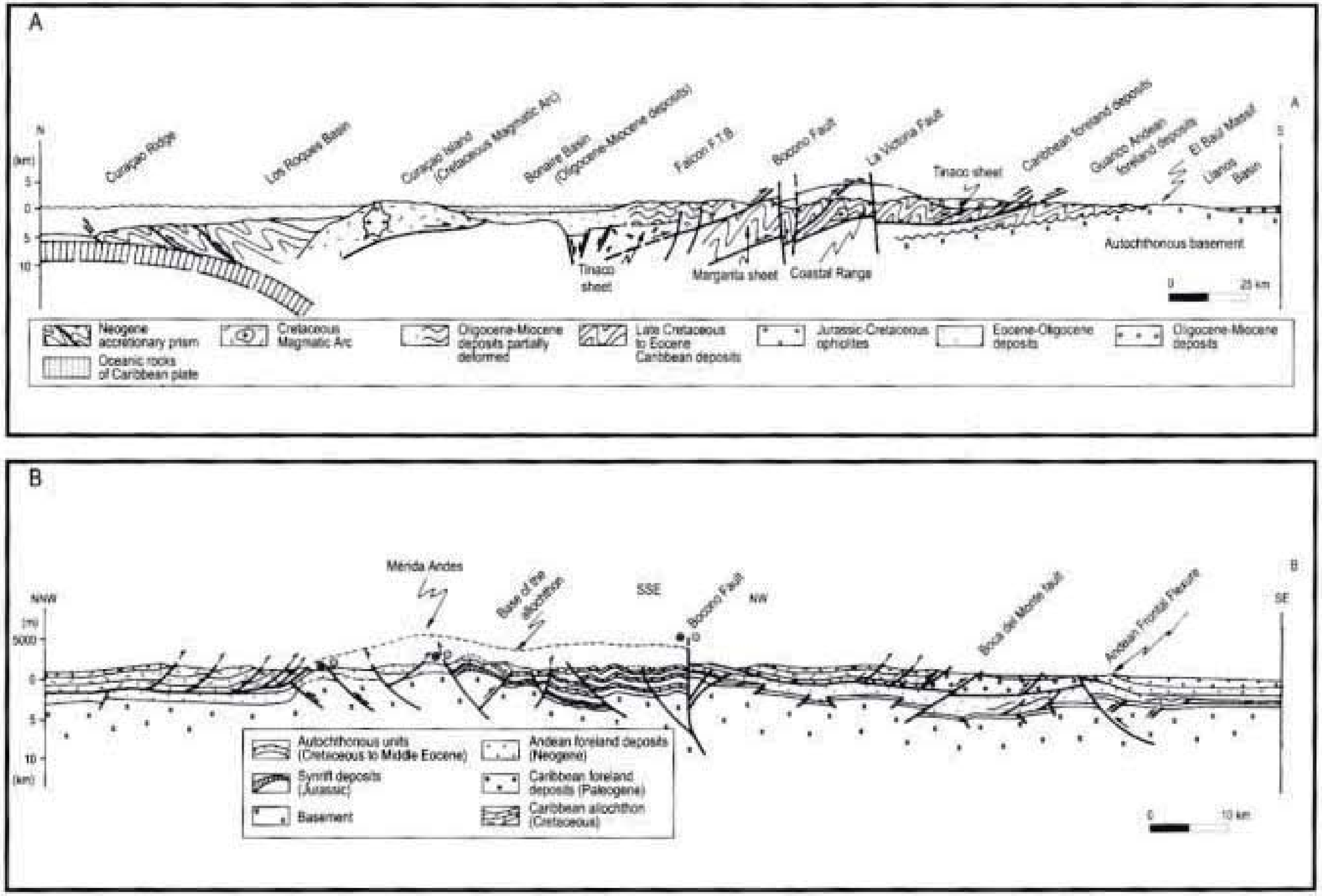


FIGURE 6 - Regional cross-sections of the Venezuelan Andes. a) Caribbean Ranges (after Stephan, 1982). b) Mérida Andes (after Hooper et al., 1999). Location in figure 3.

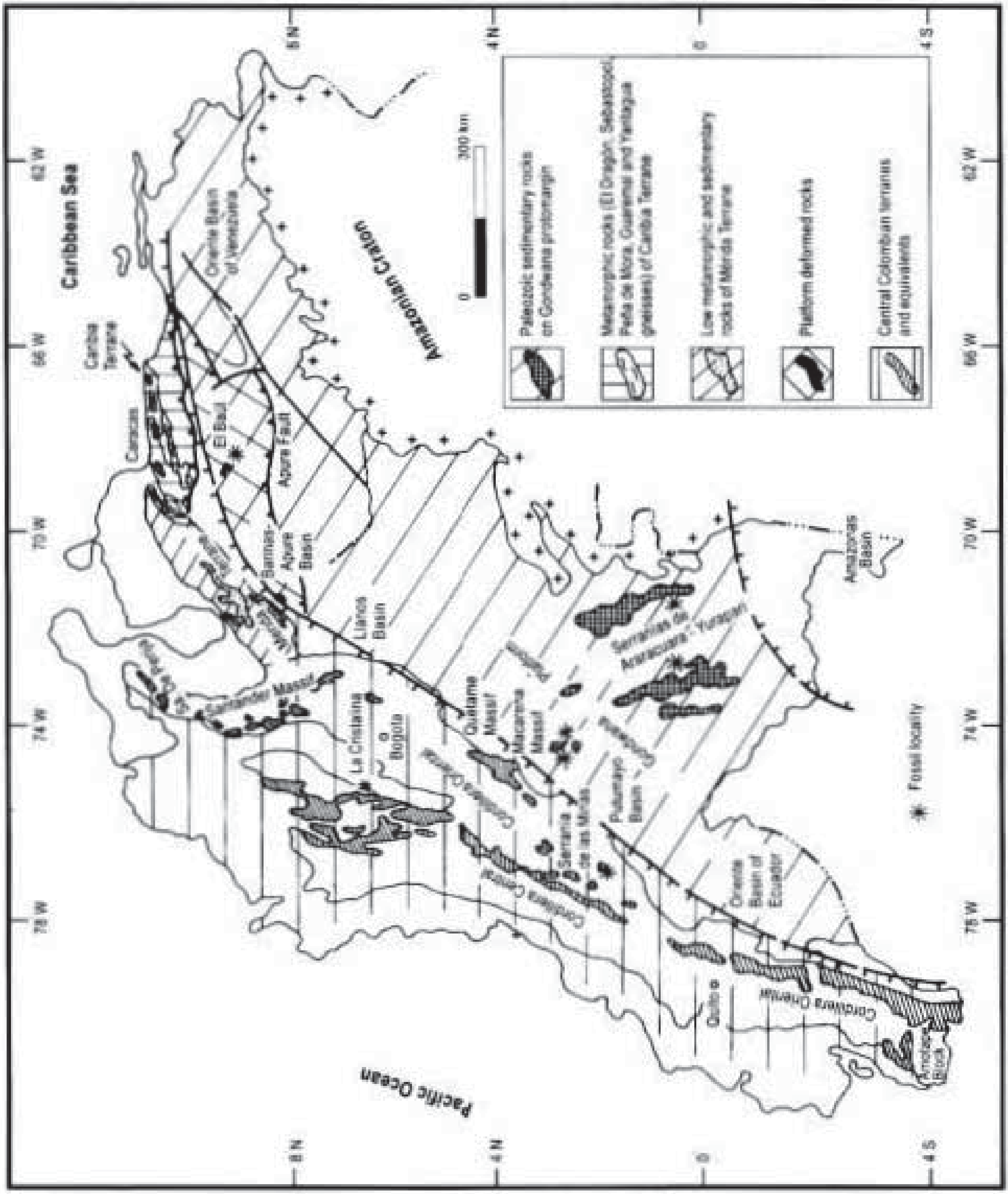


FIGURE 7 - Paleozoic basement of the Northern Andes and accreted terranes (based on Bellizzia and Pimentel, 1994).

strata pinch out against the Andean tectonic wedge. During the Pleistocene, a sequence of fluvio-glacial sandstone and conglomerate beds was deposited along the major valleys, and some of these were periodically uplifted to form distinct terraces. Other sequences have been used to infer Recent displacement along the Boconó Fault (Schubert, 1984).

Crustal growth and tectonic development of the Venezuelan (Mérida) Andes

Basement isotope ages in the Mérida Andes, S of the Caparo Fault, suggest the presence of older protoliths and perhaps a northern extension of the Guiana Shield. However, crustal growth processes did not commenced until the Neoproterozoic, as can be determined by the presence of chlorite-muscovite-quartz schist and metachert with subordinates amphibolite and olivine basalt of the Bellavista Formation (Burkley, 1976; Marechal, 1983; Bellizzia and Pimentel, 1994). The Bellavista Formation is associated with the Valera Granite (593 ± 16 Ma), Río Caparo Granite (615 ± 30 Ma) and El Topo Gneiss (660 Ma), and may correlate in time with the Brasiliano Orogeny (Herz *et al.*, 1989).

Discordantly overlying the basement rocks are Ordovician and Silurian strata with Gondwana-South American faunas (Benedetto and Ramírez, 1982; Benedetto *et al.*, 1992). Intense Lower Paleozoic magmatic activity (Burkley, 1976; Marechal, 1983) may be related to the Andean-type of margin during this time, and may represent the northernmost influence of the Ocloyic Orogen. Devonian age granites imply the proximity of Laurasia to Gondwana (Fig. 7), as supported by the fauna and flora described in the Perijá Mountains (Berry *et al.*, 1997). However, Mérida Terrane docking did not take place until the Westphalian as suggested by the presence of the Sabaneta and Palmarito formations. This collision event was an important episode of crustal growth accompanied by variable Barrowian metamorphism and penetrative deformation of the Mucuchachi and Tostosa formations. Tectonic transport, indicated by stretching lineation of originally horizontal folded structures, suggests a shear couple deformation during terrane emplacement (Kovisars, 1971; Marechal, 1983; Bellizzia and Pimentel, 1994). Culmination of this tectonic event involved development of long cylindrical folds in post-Westphalian rocks with NE-SW upright-oriented axis, or showing slight overturning toward the NW. This deformation was associated with widespread calc-alkaline plutonism from Early Permian to Triassic (290 - 225 Ma).

NW-SE Jurassic extension was coeval with opening of the western Tethys Ocean and Gulf of Mexico, and was marked by an early phase of passive margin development (Pindell and Barret, 1990; Bartok, 1993). Main depositional grabens were oriented subparallel to the present day Mérida Andes which follow the trend of the Uribante Trough (Renz, 1959). These grabens were filled with varicolored siliciclastic sediments, interbedded with felsic and basic volcanic units. However, accommodation zones such as the Mérida Arch were orthogonal to the mountain belt (Lugo, 1994). A fully developed passive margin was achieved during Early Cretaceous while high subsidence rates continued along extensional grabens until the Cenomanian. Late Cretaceous sedimentation took place in the setting of a passive margin. Nonetheless, Paleogene molasse deposition was mainly related to the interplay between thrusting of the Caribbean nappes and deformation in the Colombian Andes. The orthogonal Mérida Arch has been periodically active through early Eocene suggesting reactivation of pre-existing faults (Lugo, 1994)

Deformational processes responsible for the formation of the present-day Mérida Andes did not begin until the late Oligocene and were manifest by a double vergence foldbelt and foredeeps. According to Colletta *et al.* (1997), this paired foldbelt represents relatively minor intraplate readjustment between the Eastern Colombian Cordillera and the South Caribbean transform creating conjugate foreland basins and foothills. During the Neogene, the Maracaibo Block detached from the South American mainland to accommodate most of the deformation at the triple junction between the Pacific, Caribbean oceanic domains, and South America. Colletta *et al.* (1997) explained the surface thrust fronts parallel to the plate boundary, strike-slip motion in the allochthon along the Boconó Fault and asymmetric wedging at depth in terms of strain partitioning during Neogene oblique convergence. The large negative gravity anomaly in the northern Andean trough (- 150 mGal) suggests that this foredeep and the Andes are not in isostatic equilibrium (Kellogg and Bonini, 1982). This view supports the possibility of a SE-dipping subduction of parts of the Maracaibo continental lithosphere (Colletta *et al.*, 1997). However, the southern trough lacks a major gravity anomaly and its syn-orogenic sequence does not fill a flexural foredeep, but rather constitutes a gentle NW-dipping monocline recording Recent Andean uplift (Colletta *et al.*, 1997).

Deratmiroff (1971) mapped low-angle thrusts along the central segment of the northern Mérida front and between Torondoy and Valera. Castrillo and Hervouet (1996) have described complex imbricate thrusting and thin-skinned thrusting associated with strike-slip faults. Recently, Colletta *et al.* (1997) described the puzzling NE Brujas High that involves inversion of Jurassic grabens along high angle faults. Dominant frontal triangle zones are characterized by complex duplexes decoupled in the basement as well as in the Colón Formation (Late Cretaceous) and in Paleogene and Neogene shale beds (Deratmiroff, 1971; Audemard, 1991; De Toni and Kellogg, 1993; Colletta *et al.*, 1997). Miocene syn-flexural strata and backthrusting were involved in the shortening and wedging out toward Lake Maracaibo. However, Plio-Quaternary strata pinch out against the Andean tectonic wedge (Colletta *et al.*, 1997).

Although the bulk of shortening took place during the Miocene, there is still significant present day deformation in the Mérida Andes as recorded in tilted terraces, earthquake focal mechanism and apatite fission track data (Schubert, 1984; Kohn *et al.*, 1984). Thus, uplifting, tilting and lateral displacement of these terraces (Meier *et al.*, 1987; Schubert and Vivas, 1993) and focal mechanism studies (Malavé and Suárez, 1993) have been used to document current tectonic activity. Systematic determinations of the sense and magnitude of the displacement of Quaternary sediments and topographic features along the Boconó Fault have demonstrated vertical and right-lateral displacements exceeding 250 m. To the S, the Bocono Fault abuts against the Punzón de Pamplona reverse faults which ties in with the Eastern Cordillera Frontal Fault System (Beltrán, 1994). Crustal-scale balanced cross-sections have helped to constrain about 60 km of Neogene shortening (Colletta *et al.*, 1997).

The Colombian Andes

Several benchmark papers have been published on the Colombian Andes, including work by Nelson (1957), who made the first traverse across the Central and Western Cordillera from Ibague to Cali; and by Burgl (1961) who summarized his work on the stratigraphy and geological history of Colombia. Radeli (1967) published one of the most comprehensive reviews on the geology of the Colombian Andes based upon his extensive fieldwork. Other important works on the structural and stratigraphic evolution of the mountains include Irving (1971), Julivert (1973), Toussaint (1978), Etayo-Serna and Barrero (1983), and Restrepo and Toussaint (1988).

The evolution of the Colombian Andes encompasses a complex amalgamation of multiple allochthonous terranes in time and space. The mountain belt consists of three distinct and separate chains: the Western, Central, and Eastern Cordillera (Fig. 8), bounded to the E by the Borde Llanero Fault System (Forero-Suarez, 1990) which marks the boundary with the Proterozoic basement rocks of the Guayana Shield. The Western Cordillera consists mainly of Cretaceous tholeiitic basalt and deep-water sedimentary facies resting on oceanic crust (Barrero, 1979). Ophiolitic rocks have been reported from the Baudó Range in the Western Cordillera up to the western flank of the Central Cordillera (Toussaint, 1978). The Romeral crustal-scale fault zone contains lawsonite-glaucophane schist and eclogite (MacCourt and Feininger, 1984). The Romeral Fault separates the Western and Central Cordillera and has been interpreted as a suture or subduction zone (Toussaint and Restrepo, 1982). The Central and Eastern cordilleras are underlain by continental crust. Thus, the basement of the Central Cordillera consists of metapelite and metavolcanic units known as the Central Andean Terrane, with sedimentary cover ranging in age from Paleozoic to Tertiary. Whereas the Central Cordillera has undergone pervasive plutonic and magmatic activity with evidence for migration and widening of the volcanic arc, the Eastern Cordillera lacks extensive volcanism except for isolated, subordinate Cretaceous basic volcanism.

Tectono-stratigraphic evolution

Precambrian

Grenville age rocks have been reported in the Garzón, Santander and Sierra Nevada de Santa Marta massifs (Fig. 9). Perhaps these rocks also occur in the Serrania de Perijá (Tschanz *et al.*, 1974; Alvarez and Cordani, 1980; Kroonenberg, 1982; Priem *et al.*, 1989; Restrepo-Pace *et al.*, 1997). Noteworthy, also is the presence of granulite xenoliths in lava of the Nevado del Ruiz Volcano (Jaramillo, 1978). Restrepo and Toussaint (1978) have reported garnetiferous amphibolite from Caldas, on the eastern flank of the Central Cordillera, yielding a single hornblende K/Ar age of 1.67 ± 0.5 Ga. Trumpy (1943) reported high-grade metamorphic rocks, without geochronological support, in the Serrania de Macarena.

Radelli (1962) made some of the first petrologic descriptions of the rocks of the Garzón Massif and later Kroonenberg (1982) distinguished two petrogenetic units. The dominant Garzón Group consists of banded charnockitic and garnetiferous granulite, mafic granulite, and amphibolite. However, the less extensive Guatopon and Mancagua hornblende-biotite augengneiss are concordantly foliated with the hosting Garzón Group and are interpreted as metamorphosed syn-tectonic granites (Kroonenberg, 1982). Intercalation of calc-silicates and pelitic gneiss with this granulite suggests a sedimentary protolith (Kroonenberg, 1982; Priem *et al.*, 1989). Alvarez and Cordani (1980) have obtained a four points Rb/Sr isochron of 1.18 Ga in these granulite supracrustal rocks. Six of eight samples from this massif define a line corresponding to an age of 1.172 ± 0.09 Ga. However, there are also abundant 1.56 - 1.45 Ga old augen gneiss related to the underlying Parguazan tectonomagmatic event observed in the Guayana Shield (Priem *et al.*, 1982, 1989). The age of the protolith is confirmed by calculated Nd model ages showing a consistent average of *c.* 1.55 Ga (Restrepo-Pace *et al.*, 1997).

Slivers of Orinoquian age (*c.* 1.1 Ga) basement rocks outcrop in the northern part of the Santander Massif (Ward *et al.*, 1973). They include two distinctive petrotectonic units: a low to medium-grade metamorphic metapelitic-metapsammitic unit, and a structurally concordant foliated granite orthogneiss unit with intrafoliated amphibolite dykes displaying tholeiitic chemical signatures. Mineral paragenesis suggest low medium pressure (3 - 9 kbar) and high temperature (300 - 700 °C) greenschist to upper amphibolite facies rocks with local kyanite, implying medium pressures in restricted areas (Restrepo-Pace, 1992; Restrepo-Pace *et al.*, 1997). In like manner to the Garzón Massif, the protolith of these gneiss are considered to be sedimentary rocks.

The isolated, triangular massif of the Sierra de Santa Marta contains Precambrian rocks which consist of quartz-perthite granulite; and intermediate, mafic, ultramafic, calcareous and garnetiferous granulite with mineral paragenesis similar to the Garzón Massif (Tschanz *et al.*, 1974). Strong Rb/Sr isochrons of 1.37 - 1.27 Ga have been recalculated from gneiss (MacDonald and Hurley, 1969; Kroonenberg, 1982). Furthermore, whole rock Rb/Sr data of 1.273 Ga and 736 Ma have been reported from granulite as well as a hornblende K/Ar age of 949 ± 36 Ma (Tschanz *et al.*, 1974). In the Guajira Peninsula, 200 km NE of Santa Marta, a 1.25 Ga U/Pb zircon age has been determined on the Jojocinto Leucogranite-Gneiss (Case and MacDonald, 1973). Calculated Nd model ages show a consistent average of *c.* 1.7 Ga, somewhat older than the Garzón samples (Restrepo-Pace *et al.*, 1997).

Although Kroonenberg (1982), precluded a long metamorphic crustal history for the Garzón and Santa Marta massifs, based on low initial strontium isotope ratios, he recognized an important metamorphic event around 1.2 Ga

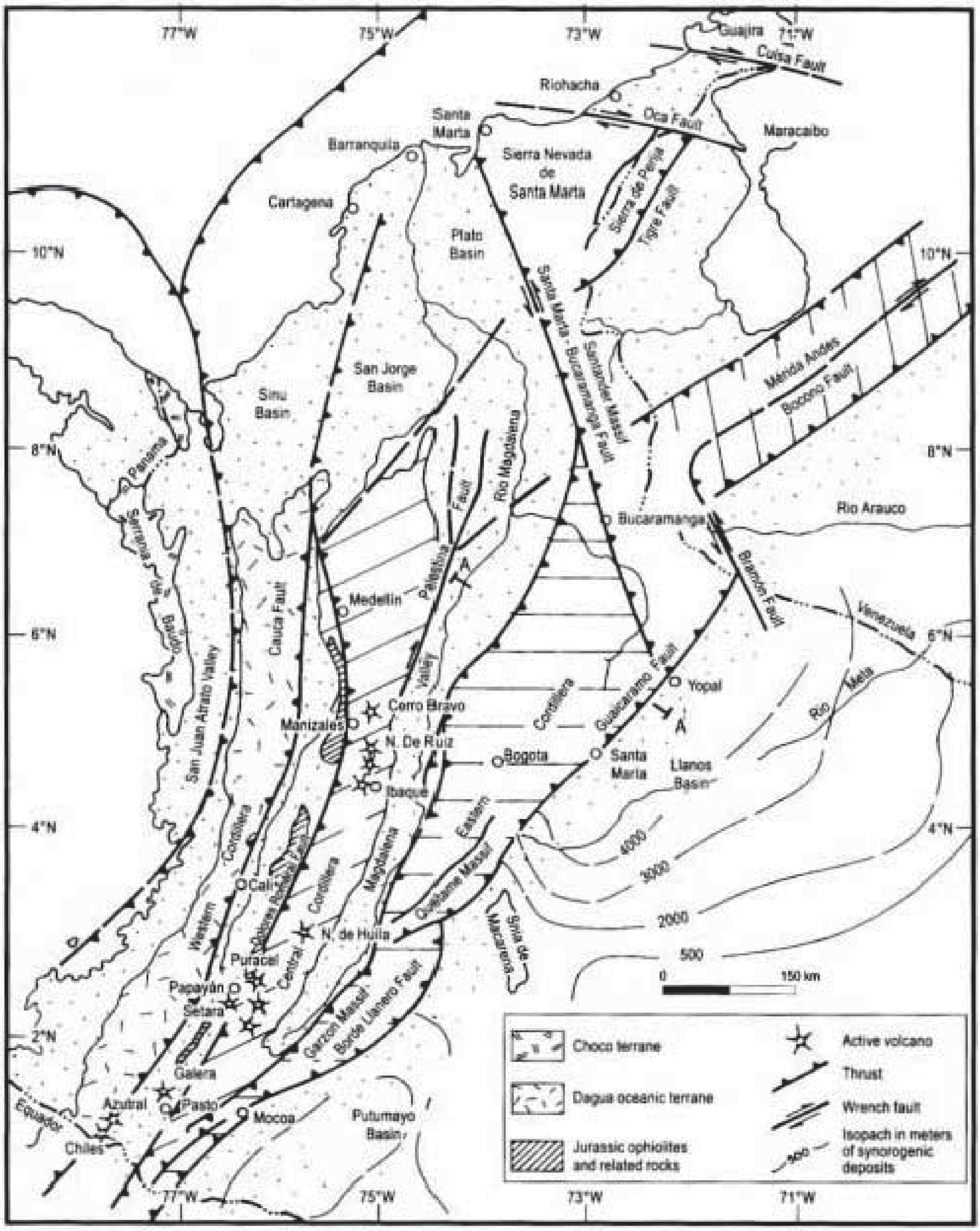


Fig. 8 - Major geologic provinces of the Colombian Andes (based on Bourgois et al., 1987; Restrepo and Toussaint, 1988; Forero-Suárez, 1990).

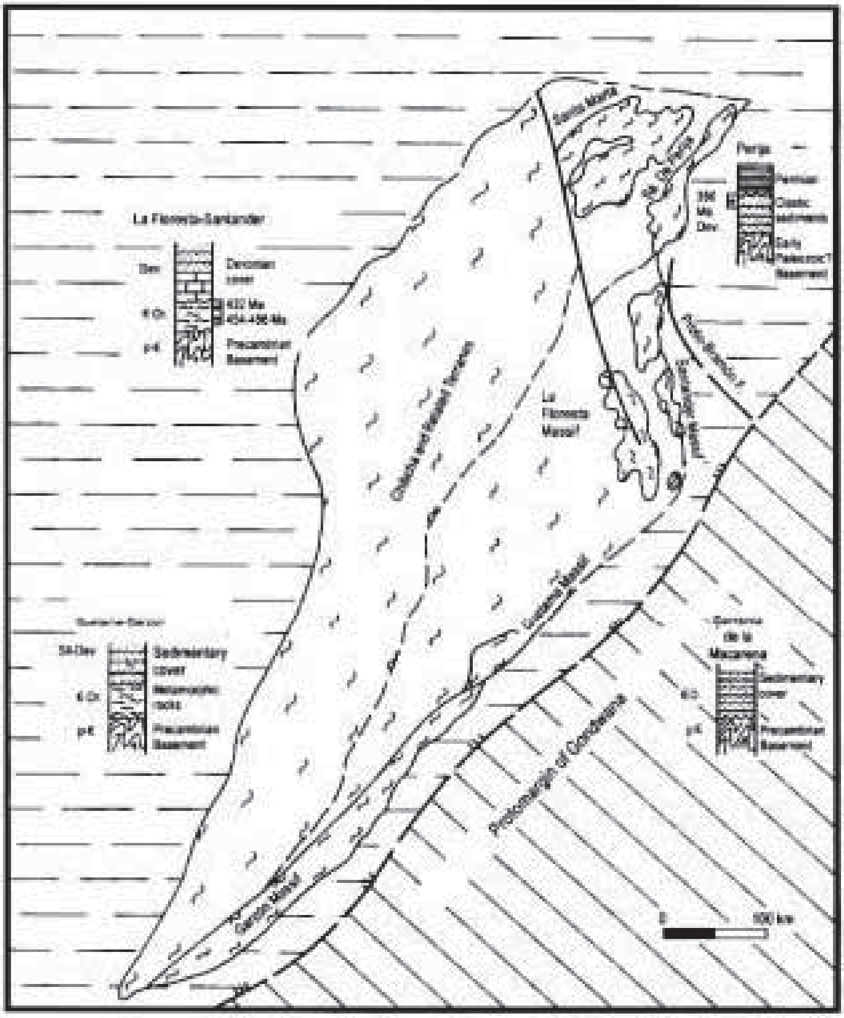
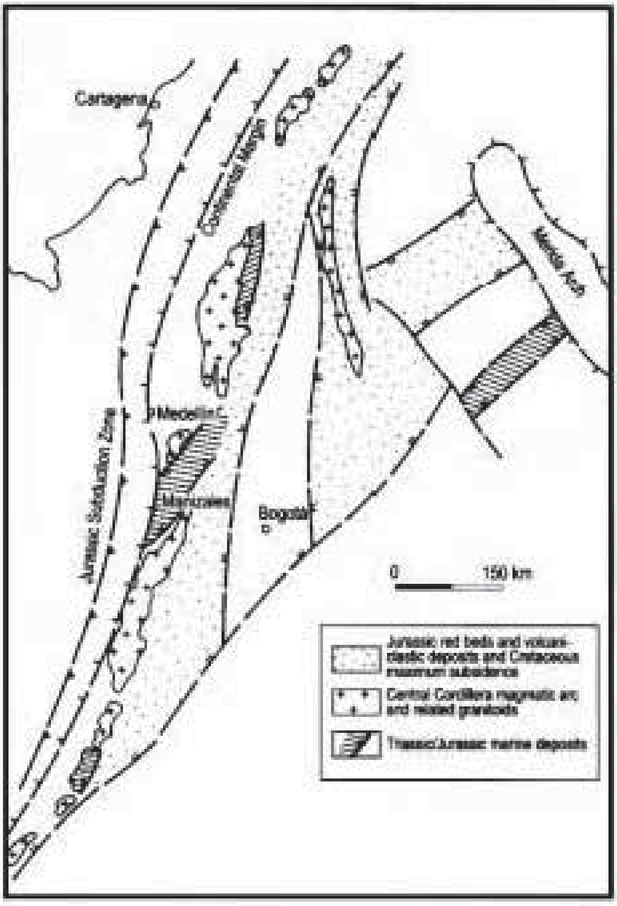


FIGURE 9 - Paleozoic tectonic setting of the Colombian Andes with main Precambrian outcrops. The Chibcha (or Central Colombian) and related terranes have docked to the proto-margin of Gondwana during the Late Ordovician (based on Etayo-Serna and Barrero, 1983; Toussaint, 1993). The Santa Marta, Santander, and Perijá blocks have been palinspatically restored to their Paleozoic position.

FIGURE 10 - Mesozoic tectonic settings of the Colombian Andes (based on Mojica et al., 1996). Maximum Cretaceous subsidence was located along Jurassic grabens, and continued until Cenomanian.



along the Guaiana Shield. He interpreted this granulite belt as having formed in the western border of the Paleo to Mesoproterozoic nucleus of the Guiana Shield by an orogenic event around 1.2 - 1.4 Ga (Orinoquian). Thus, the lithology of the belt matches an ensialic calc-alkaline volcanic suite with minor pelitic, semipelitic and calcareous intercalations, probably deposited in a shallow marine environment, and cut by small ultramafic and anorthositic intrusives (Kroonenberg, 1982).

In contrast, Priem *et al.* (1989) recognized a *c.* 1.6 Ga old augen gneiss (Guatopon and Mancagua) in the Garzón Massif, similar to that observed in the adjacent parts of the Guiana Shield (Parguasan tectonomagmatic episode). They also described a supracrustal sequence metamorphosed in the granulite facies around 1.2 Ga (Orinoquian Orogeny), and a set of pegmatite dykes (*c.* 850 Ma). The low initial strontium isotope has been interpreted, alternatively, as an indication that the metamorphism took place shortly after deposition. Thus, most of the sporadic Precambrian slivers distinguished thus far along the Colombian Andes have shared the same tectonic history as the Garzón Massif. Because the Orinoquian Orogeny (1.2 Ga) trended subparallel to the present Andes (Litherland *et al.*, 1985), it was originally interpreted to represent a continental collision between the western margin of the Guiana Shield and the eastern margin of the Canadian Shield (Priem *et al.*, 1989). However, an alternative interpretation suggests a closer relationship with the Oaxacan Complex of southern Mexico (Keppie and Ortega-Gutierrez, 1995; Restrepo-Pace *et al.*, 1997).

Paleozoic

In the Serrania de la Macarena, Trumpy (1943) reported Middle Cambrian to Lower Ordovician trilobites, brachiopods and graptolites in the Güéjar Group (Fig. 9); also described by Harrington and Kay (1951). This group consists of dark mudstone, phyllite to slate interbedded with quartz-rich sandstone and greywacke. Trumpy (1943) speculated that the Quetame Group might be the metamorphic equivalent of the Güéjar Group. Radelli (1967) later confirmed this. To the N, quartzite, micaschist and gneissic schist of the Perijá Formation; and to the W, micaceous gneiss, mica schist, phyllite and marble of the La Ceja Group in the Central Cordillera, have been also ascribed to the Cambro-Ordovician and may correlate with the Güéjar Group. Villaroel *et al.* (1997) described on the western flank of the Eastern Cordillera a sequence of dark grey micaceous shale beds, interbedded with arkósic sandstone and polymictic conglomerate beds of the Venados Formation (Middle Ordovician).

To the E of the Borde Llanero Fault, Ulloa *et al.* (1982) described trilobites, brachiopods and graptolites in grey to black micaceous shale beds that grade upward to interbedded siltstone and sandstone units with subordinates limestone members of the Lower Negritos Formation (Lower Ordovician). Along the Serrania de Las Minas, in the Central Cordillera, Mojica *et al.* (1988) reported graptolites and trilobites in fine grained arkosic sandstone beds, that grade upward to dark grey to black shale beds of the Higado Formation (Middle Ordovician), informally described as La Cristalina Formation. These units may be equivalent to the graphitic schist of the Cajamarca Group (Nelson, 1957; McCourt *et al.*, 1984). Structurally, they are seen as well-developed isoclinal folds with penetrative schistosity, including crenulation cleavage indicative of a polyphase deformational history.

This deformation could be related to a major orogenic event at the close of the Ordovician, correlated with the Ocloyic Orogeny of northern Argentina (Ramos, 1988). The deformation accounts for a pronounced angular unconformity between Cambro/Ordovician rocks and the Devonian basal conglomerates. In the Santander Massif, another major unconformity also separates Middle Devonian siliciclastic rocks from potentially laterally equivalent metamorphic Ordovician rocks (Boinet *et al.*, 1986). Indeed, new isotope ages from the Santander Plutonic Group suggest Late Ordovician (457 Ma) to Early Devonian (413 Ma) for some of the granites, and a Late Ordovician age (456 ± 22.8 Ma) for one gabbro (Boinet *et al.*, 1985). Late Silurian (Ludlovian) palynomorphs have recently been described by Grösser and Prössl (1991) in interbedded shales of the Areniscas de Guttierrez of the Quetame Massif. These were originally ascribed to the Middle Devonian (Renzoni, 1968).

Devonian rocks are restricted to the Eastern Cordillera and its northward extension in the Sierra de Perijá. However, in the Central Cordillera, near the town of Rovira, a sequence of black shale beds with irregular interbedded sandstone units of the Amoya Formation contains Middle Devonian palynomorphs (Prössl and Grösser, 1995). To the E, in the Santander Massif, the Floresta Formation consists of a basal conglomerate that grades upward to interbedded quartz-rich sandstone and shale units with abundant Early to Middle Devonian fauna (Boinet *et al.*, 1986; Barrett, 1988). According to Boinet *et al.* (1986), the Floresta Formation is also intruded by the Onzaga Monzonite (394 ± 23 Ma). Northwards, in the Serrania de Perijá, the thick Rio Cachiri Group (Devonian) consists of more than 1300 m of interbedded shale and sandstone with a few biostromal limestone units containing a Givetian to Frasnian fauna and flora (Berry *et al.*, 1997). It is noteworthy that Late Devonian isotopic ages have been reported for the Las Lajas Granite (370 ± 20 Ma).

Carboniferous rocks are also well represented in the Eastern Cordillera and its northern extension. Late Carboniferous fauna has been described in interbedded sandstone, conglomerate and graphitic shale units of the Gachala Formation in the Quetame Massif and in the Garzón Massif. To the N, in the Santander Massif the stratigraphic thickness increases to a maximum of 2900 m of red colored sandstone and shale with subordinated limestone and conglomerate beds known as the Labateca Formation. Farther to the N, in the Serrania de Perijá, a sequence of thin-bedded black shales interbedded with thick massive fossiliferous limestone members and varicolored siltstone and sandstone beds of the Tinacoa Formation (Carboniferous) is reported.

Permian rocks are absent in the Garzón and Quetame massifs. However, in the Garzón Massif, Mojica *et al.* (1988) reported limestone members of possible Wolfcampian age. In the Santander Massif, fine-grained sandstone and shale is interbedded with thick-bedded fusulinid-bearing limestone units of the Diamante Formation. These sediments are overlain by massive limestone conglomerate beds of the Tiburón Formation of Lower Permian age (Ward *et al.*, 1973). To the N, in the Sierra de Perijá, siltstone and sandstone beds grade upward to dark grey fossiliferous limestone units of the Palmarito Formation (Late Wolfcampian to Early Leonardian; Thompson and Miller, 1949). The northernmost Permian outcrops are in the Sierra Nevada Massif, and consist of crystalline limestone and marble units with wollastonite interbedded with quartz-rich sandstone and conglomerate that Gansser (1955) has assigned to the uppermost Paleozoic. These Permian rocks have been intruded by Late Permian to Early Jurassic calc-alkaline granites varying from I-type to S-type. The S-type granites are related to shear zones before Late Triassic separation of North and South America (Pindell and Barrett, 1990). Indeed, amphibolite and sheared ultrabasic pods in the Rosario Complex along with the metagabbro and amphibolite of the Bolo Azul Complex, may represent the accretion of Upper Paleozoic rocks onto the South American Craton along the Palestina Fault (Feininger, 1970; McCourt and Feininger, 1984).

Mesozoic

A small, foliated, calc-alkaline plutonic precursor of Late Triassic age situated in the Magdalena Valley was related to ensialic back-arc extension similar to that proposed for the Choyoi Group in Central Argentina (Ramos and Kay, 1990). Indeed, Jurassic isotope ages (163 ± 10 Ma and 131 ± 9 Ma) for the Cauca Ophiolitic Complex (Restrepo and Toussaint, 1973) and the blueschist facies associated with the Romeral Fault System provide evidence for a subduction zone. They also explain the presence of abundant Jurassic batholiths in the Eastern Cordillera (Fig. 10). The NE-SW oriented Triassic/Jurassic grabens (Cediel, 1981; Mojica *et al.*, 1996) are filled with pre-Norian arkosic sandstone beds, polymictic conglomerate and breccia (Luisa Formation); dark grey limestone units, locally fossiliferous (Payandé Formation); Rhaetic to Middle Jurassic pyroclastic, ash and lapilli tuff deposits, and rhyolitic to dacitic agglomerates (Saldaña Formation). This sequence is cut by several diorite and gabbro bodies coeval with the San Augustin and Gallego Granites (172 - 159 Ma). In the Santander Massif, to the E, varicolored mudstone beds, siltstone, and sandstone units interbedded with welded tuff and conglomerate of the Jordan Formation (Late Triassic) filled the graben. In turn, varicolored conglomerate beds, sandstone and shale beds of the Girón Formation overlie this unit. It is locally intruded by the Aguablanca (196 ± 7 Ma) and Mogotes (193 ± 6 Ma) batholiths (Ward *et al.*, 1973).

In the Serrania de Perijá, NW-SE oriented grabens are filled by a sequence of varicolored conglomerate units, interbedded with arkosic sandstone, tuff and mudstone of the La Quinta Formation. This formation is also present in the Guajira Peninsula where polymictic conglomerate beds change from predominantly volcanic to mainly granitic and metamorphic with basaltic andesite and abundant ash layers near the base (Maze, 1984). The northern and southern Central Cordillera contains Late Triassic grabens filled with red beds (El Sudán Formation) and overlain by Jurassic dark grey shale, interbedded with volcaniclastic sediments (Geyer, 1980). However, the central part of the Central Cordillera contains abundant Early Triassic to Jurassic stocks and plutons that are also present in the Eastern Cordillera from the Garzón Massif to the Sierra Nevada (Maya, 1992). Mantle derived, calc-alkaline Jurassic plutonism is also present in the western margin of the Central Cordillera such as at the Mocoa (180 - 170 Ma), Segovia (160 Ma) and Ibague (150 - 140 Ma) batholiths.

A new pulse of ensialic extension took place during Early Cretaceous in two separate grabens. To the E, and to the S of the Machiques Trough, a new graben developed between the paleo Guaicáramo and Chiscas faults and was filled by Berriassian to Aptian conglomerate, sandstone and shale. To the W, the graben along the Mundo Nuevo Syncline continued to subside and was filled by Late Valanginian transgressive sandstone units of the Tambor/Yavi formations (Fabre, 1987; Etayo-Serna, 1979). This period of crustal extension was perhaps related to an increase in the subduction angle and concomitant trenchward volcanic arc migration from Jurassic to Late Cretaceous (McCourt and Feininger, 1984, Toussaint and Restrepo, 1982). A regional sagging phase took place during the Aptian, and was accompanied by widespread subsidence and deposition in the Eastern Cordillera. Regional flooding during the Albian resulted in the deposition of dark grey shale beds and limestone units. In the Cenomanian to Turonian there occurred the deposition of limestone and shale in a high-upwelling regime during a series of transgressive pulses (Villamil and Arango, 1998). Maastrichtian regressive sandstone beds were associated with the rise of the Central Cordillera.

To the W, a marginal basin to island-arc tholeiitic assemblage developed between the Central Cordillera and the Baudó Range (Bourgois *et al.*, 1987; Spadea and Espinosa, 1996). Aptian to Senonian deposition in this basin included siliceous and carbonaceous phyllite and slate of the Dagua Group; basaltic lava flows, hyaloclastic breccia, dolerite, gabbro and minor pyroclastics of the Diabase Group; and a very thick sequence of subaqueous pyroclastics and volcaniclastic sandstone beds, interbedded with mudstone and chert of the Espinal Formation. Interpretation of this suite of rocks vary from oceanic island arc (Barrero, 1979) to ocean floor (Pitchler *et al.*, 1974) and oceanic flood basalt (Millward *et al.*, 1984). Late Cretaceous tectonic escape and formation of the Caribbean Plate (Stephan *et al.*, 1990) caused obduction of this marginal basin sequence as indicated by complex folding and thrusting. Alpine-type nappes were thrusted from NW to SE with development of recumbent folds, isoclinal folding and crenulation cleavage (Bourgois *et al.*, 1987). Indeed, according to Burgois *et al.* (1987), the structure of the Western Cordillera consists of a stack of nappes deformed to an antiform.

First obduction over the Central Cordillera took place during Late Jurassic or Early Cretaceous (Fig. 8), as suggested by the Jambaló Glaucophane Schist (Feininger, 1982) and other metaophiolite complexes with westward dipping foliation along the Romeral Fault (Bourgois *et al.*, 1987). A second phase of obduction can be interpreted from the Yurumal Complex that consists of serpentinite, peridotite, gabbro, massive tholeiitic basalt flows, pillow basalt associated with chert, tuff, and volcaniclastic turbidite beds toward the top (Restrepo and Toussaint, 1977). The complex is intruded by the Antioquia Batholith (80 - 60 Ma). During this obduction phase, the Atrato fore-arc basin was formed, and high subsidence rates provided the favourable conditions for marine deposition throughout Pliocene. Oblique Eocene collision of the Acandí-Mandé-Gorgona Terrane enhanced subsidence rates in fore-arc basins and triggered strain partitioning causing reactivation of the Romeral and Cauca-Patia paleo-sutures as strike-slip faults.

The trailing edge of the Western Cordillera, represented by Late Cretaceous komatiitic rocks of the Gorgona Terrane (Echeverría, 1980; Etayo-Serna and Barrero, 1983). These mafic rocks were interpreted to represent an oceanic plateau formed by extensive decompression melting of an uprising deep mantle plume with similar REE content to rocks in Curaçao (Spadea and Espinosa, 1996; Kerr *et al.*, 1996). Farther to the W, the Late Cretaceous arc sequence of the Baudó Range docked during the late Miocene.

Cenozoic

The Calima Orogeny of Late Cretaceous to Paleocene times (Barrero, 1979) was heralded by uplift of the Central Cordillera. This resulted in the deposition of a molasse wedge, interrupted by local basement uplifts such as that of the Santander Massif in the Eastern Cordillera. Deformation was coeval with the formation of nappes, the extrusion of Cretaceous volcanic rocks in the Western Cordillera, and the emplacement of the Antioquia Batholith in the Central Cordillera. Whereas Tertiary sedimentation in the Eastern Cordillera was dominated by deposition of fluvio-deltaic clastic deposits, sedimenation in the Western Cordillera was marine, and included Miocene deep-water facies in the Atrato Basin. The Central Cordillera, on other hand, was uplifted and affected by pervasive plutonism and subaerial volcanism throughout the Tertiary.

Oblique convergence between the Farallón and South American plates during the Eocene resulted in strain partitioning into E-W compression in the Magdalena Valley and pull-apart basin formation along the Cauca/Patia-Romeral Fault System. Upper Eocene to lower Miocene fluvial sandstone and shale beds were deposited in these basins. N-S strike-slip faulting related to the Romeral/Palestina Fault System developed during the rise of the Western Cordillera. Docking of the Acandí-Mandé-Gorgona Terrane caused a trench jump, starting a new arc-trench system and renewed transpressional deformation and uplift in the Eastern Cordillera as occurs in the Magdalena Valley (Butler and Schamel, 1988). This is interpreted from mesoscopic structures cut by 38 - 35 Ma hydrothermal veins (Cheilletz *et al.*, 1994; Branquet *et al.*, 1996) as well as the creation of independent foredeeps in the Magdalena Valley and the Sabana de Bogotá basins and an increase of Eocene subsidence in the fore-arc Atrato Basin (Incaic Orogeny). Moreover, from late Eocene to Miocene, there is a well-documented eastward migration of the volcanic arc (Toussaint and Restrepo, 1982).

During the Oligocene there occurred extensional collapse of the Northern Central Cordillera with concomitant formation of the Plato and San Jorge basins. A low rate of convergence (Pardo-Casas and Molnar, 1987) and a decrease in subduction angle of the Nazca Plate may explain the lack of magmatic activity, and relative tectonic quiescence during the Oligocene. A major marine transgression in the Eastern Cordillera foredeep is manifest in widespread deposition of shale beds.

During the Miocene, a time of maximum paroxysm and mountain building in Colombia, there occurred renewed folding and thrusting in the Eastern Cordillera (Fig. 11a), with the deposition of a thick molasse sequence in the present-day foredeep (Dengo and Covey, 1993) and by an increase in the subsidence in the Middle and Upper Magdalena basins. Strain partitioning during oblique convergence resulted in significant compressional deformation and reactivation of pre-existing dip-slip faults as strike-slip faults in the Magdalena Valley. Eastern Cordillera shortening was coeval with emplacement of the Baudó Range, development of orogenic parallel strike-slip faults and sinistral strike-slip displacement along the Santa Marta Bucaramanga Fault (Campbell, 1968). Miocene plutonic and volcanic activity varying from alkaline to calc-alkaline in composition was widespread in the Western Cordillera, in the upper reaches of the Magdalena River Valley, and along the axis of the Central Cordillera (Alvarez, 1983; McCourt *et al.*, 1984). Thus, the fluvial conglomerate and sandstone of the Combia Formation (Restrepo *et al.*, 1981), deposited in an extensional basin trending ENE-E, contain andesitic lava flows and pyroclastic beds with island-arc tholeiitic affinities at the top of the section.

Plio-Pleistocene calc-alkaline volcanism continued in the Central Cordillera and along the Cauca/Patia and Romeral faults. Segmentation of the arc-trench system explains the lack of volcanism in the Bucaramanga region, N of the Atrato-San Juan lineament, where earthquake focal mechanisms suggest a shallow subduction angle (Pennington, 1981). Upper mantle Pliocene to Pleistocene alkali basaltic and ultramafic magmatism has been observed in the uppermost regions of the Magdalena Valley and may be attributed to subduction-related deep crustal fractures (Kroonenberg *et al.*, 1982). This volcanism is coeval with rhyolite ignimbrites found in the Central Cordillera.

Crustal growth and tectonic development of the Colombian Andes

Based on Nd model ages, early crustal growth processes in the Colombian Andes began during the Parguasan (1.45 Ga) tectonomagmatic event (Restrepo-Pace *et al.*, 1997). Similar ages have been reported in the Guiana Shield (Priem *et al.*, 1982, 1989) suggesting that these basement rocks extend under the Colombian Andes. A second tectonic metamorphic event occurred at approximately 1.1 Ga involving formation of a granulite belt around the nucleus of the Guyana Schield in the Paleo to Mesoproterozoic times during the collisional Grenville Orogeny (Kroonenberg, 1982; Priem *et al.*, 1982, 1989; Restrepo-Pace *et al.*, 1997).

Cambro-Ordovician siliciclastic rocks were deposited on a passive margin, devoid of volcanic activity. A Late Ordovician Cordilleran-type margin is interpreted from the

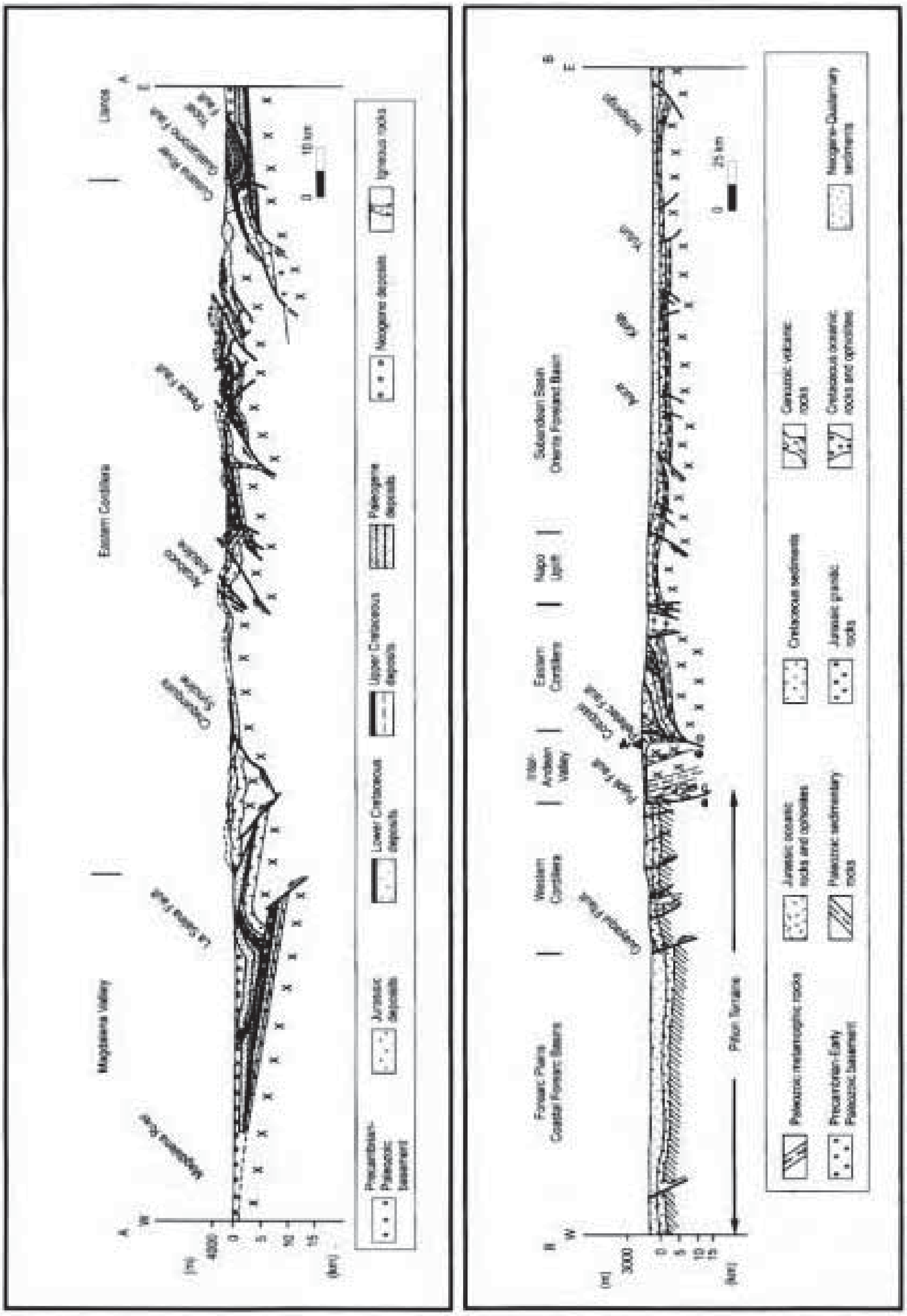


FIGURE 11 - Regional cross-sections: a) Eastern Cordillera of Colombia (based on Colletta et al., 1990). b) Ecuadorian Andes. Location in figs. 8 and 12.

presence of gneiss, schist, migmatite and granite (Santander Plutonic Group) of the Central and Eastern Cordillera (Trumpy, 1943; Harrington and Kay, 1951; Boinet *et al.*, 1985) and may correlate with the Ocloyic Orogenic event of northern Argentina (Ramos, 1988). The beginning of Gondwana assemblage and the closing of the ocean between Gondwana and Laurasia is suggested by Lower Paleozoic plutons in the Mérida Andes, the Serrania de Perijá, and the Santander, Floresta and Quetame massifs in the Eastern Cordillera (Boinet *et al.*, 1985). This event was followed by deposition of siliciclastic sediments and subordinate carbonate rocks containing a Devonian flora and fauna with affinity to the Eastern Americas Realm, suggesting the absence of barriers between Euramerica and Gondwana during the Middle Devonian (Berry *et al.*, 1997). Carboniferous deposition took place in a foredeep filled with thick siliciclastic sediments that are capped by Permian shallow marine carbonates. A second collisional event in northern South America is suggested by widespread S-type granite magmatism during Late Permian to Early Triassic. This collision took place along the Palestina Fault where McCourt and Feininger (1984) suggested the presence of a paleo-suture.

Right-lateral Triassic shear of the Northern Andes away from North America accompanied by crustal-scale transpression has been speculated for the northern Andes (Jaillard *et al.*, 1990; Aspden *et al.*, 1992). However, the small, foliated, calc-alkaline plutonic precursors of regional Late Triassic extension in the Magdalena Valley could be related to ensialic back-arc extension similar to those proposed for the Choyoi Group in Central Argentina (Ramos and Kay, 1990). In fact, this event was followed by ensialic Jurassic back-arc extension associated with high heat flow and volcanism, which thinned the crust and formed a series of N-NW oriented grabens. These new Mesozoic basins were the depositional loci of volcanic and sedimentary rocks, that later were inverted during the Andean Orogeny. Calc-alkaline volcanism migrated seaward during trench rollback as interpreted by the position of the Jurassic Ibague and Segovia batholiths, relative to the position to the Late Cretaceous Antioquia Batholith.

The most important crustal growth process was the Late Cretaceous accretion of the Central and Western Cordilleras during oblique convergence of the Farallon Plate (Pardo-Casas and Molnar, 1987). This event involved intensive magmatism, eastward arc migration, regional metamorphism, and reactivation of paleo-sutures such as the Romeral and Palestina faults (Feininger, 1970; McCourt and Feininger, 1984; Restrepo and Toussaint, 1988). This tectonic event also provided the mechanism for separation of the upper part of the Magdalena Valley from the Putumayo Basin.

The middle Eocene westward shift of the loci of volcanic activity was followed by eastward arc migration (Toussaint, 1978) associated with a significant increase in convergence rates (Pardo-Casas and Molnar, 1987). An arc-trench system, developed after emplacement of the Late Cretaceous Gorgona Komatiites, encompassed pervasive calc-alkaline to tholeiitic magmatism and marine deposition in the Atrato fore-arc basin. Middle Eocene deformation was recorded by compressional deformation and molasse deposition. Equally important was the dual Middle Magdalena and Sabana de Bogotá foredeeps where synchronous, thick molasse was deposited. Compressional deformation and uplift of the Sierra de Perijá was initiated as long ago as the early Eocene (Kellogg, 1984). At that time, the Cauca Patia and Romeral sutures were reactivated as strike-slip faults with the formation of pull-apart basins, filled with terrestrial deposits along this inter Andean depression.

Although convergence rates decreased significantly during Oligocene (Pardo-Casas and Molnar, 1987), transpressive deformation continued in the Eastern Cordillera with formation of E-verging thrusts. Cessation of igneous activity in the Colombian Andes may be explained by decreased subduction angle. Eastward thrust propagation and foredeep migration were hampered by Precambrian and Paleozoic rocks in the Eastern Cordillera, which behaved as a backstop along the paleo-Guaicáramo Fault. The Sierra de Perijá also underwent compressional deformation and uplift (Kellogg, 1984), and the Caño Limón strike-slip fault, in the present-day Llanos Basin, was active. Crustal thickening and uplift of the Central Cordillera was related to igneous underplating during the process of collision/subduction. This uplift accompanied subsidence in pull-apart basins along the Cauca/Patia and Romeral faults as well as in the Magdalena Valley intermontane basins.

Break-up of the Farallon Plate during the early Miocene was associated with a change in the angle of convergence from oblique to orthogonal, with a concomitant increase in rate of plate convergence (Handschumacher, 1976; Pardo-Casas and Molnar, 1987). This event was accompanied by accretion of the Baudó Range, deformation of the pull-apart basins along the Cauca Valley, the beginning of new magmatic cycle, and diachronous tectonic uplift of the three cordilleras. The Panamá-Baudó Arc that dips to the S and W was accreted along a major strike slip fault-suture that later was intruded by ultramafic rocks and komatiitic basalt (20 Ma) of subcrustal origin (Salinas and Tistl, 1991). Shortening caused the inversion of Mesozoic half grabens characterized by high-angle, basement-involved reverse faults, folding, thrusting and uplifting of the Eastern Cordillera (Colletta *et al.*, 1990; Dengo and Covey, 1993).

Widespread deformation of the Eastern Cordillera during the middle Miocene was displayed by large antiforms and pop-up structures formed by opposite verging thrust faults, short and large wavelength and folds detached in Cretaceous rocks and complex imbricate thrusts (Butler and Schamel, 1988; Colletta *et al.*, 1990; Dengo and Covey, 1993). The foldbelt asymmetry was related to the presence of two deformation fronts and two foreland provinces (Colletta *et al.*, 1990). Indeed, loading of thrust sheets increased subsidence in the Magdalena and the Llanos basins. Northwards, the Sierra de Perijá underwent a third period of compressional deformation and inversion (Kellogg, 1984). About 125 km of left-lateral displacement of the Santa Marta strike-slip fault in the northern part of the Central Cordillera (Campbell, 1968) was perhaps related to the eastward motion of the Caribbean Plate.

Compressional deformation continued in the Colombian Andes throughout Plio-Pleistocene. It was accompanied by calc-alkaline volcanism along the Cauca Patia and Romeral faults; rhyolite volcanism in the Central Cordillera; and alkali basaltic and ultramafic magmatism in the uppermost part of the Magdalena Valley (Kroonenberg, 1982). Whereas compressional processes continued uplifting the Eastern Cordillera, igneous underplating enhanced uplift in the Western and Central Cordilleras. Equally important was the late Miocene-Pleistocene uplift of the Garzón Massif (Van der Wiel, 1991). Strain partitioning caused reverse faults reactivation as strike-slip faults in the Magdalena Valley, separation and individualization of the Neiva Basin, and formation of orogen-parallel strike-slip faults such as the Apiáy Fault in the Llanos Basin. Eastward propagation of compressional deformation of the Eastern Cordillera Foldbelt, across the Precambrian to Paleozoic backstop, took place by out of sequence movement of the Guaicáramo Fault, which splayed into the Yopal and San Juan de Ariporo frontal thrusts. The last two faults are bedding-parallel within the Eocene molasse, and this accounts for significant disharmonic folding and blind thrusting in the frontal thrust.

Ecuadorian Andes

Unlike the Colombian Andes, the evolution of the Ecuadorian Andes can be described in terms of only the Western Cordillera and Eastern Cordillera (Cordillera Real), each with an unique rock assemblage (Fig. 12). The Central and Eastern Cordillera of Colombia merge near the Ecuadorian border. A well-defined inter-Andean graben, formed by strike-slip processes, separates these cordilleras, and is bounded by the Pujili and Peltetec faults (Campbell, 1974; Baldock and Longo, 1982).

The Western Cordillera consists of a strongly deformed Cretaceous allochthonous terrane capped by Tertiary continental arc rocks and a Paleogene marine clastic and carbonate sequence intruded by Tertiary plutons. The eastern boundary is sharp and marked by the Pujili Fault, whereas the western flank consists of accreted Mesozoic back-arc rocks overlain by Tertiary fore-arc deposits. The Cordillera Real consists of highly metamorphosed, thrusted and folded Paleozoic to Mesozoic rocks, partially covered by Tertiary intrusive and volcanic rocks. The Peltetec Fault and the Oriente Basin mark the western and eastern boundaries, respectively. The Inter-Andean graben, between the two Cordilleras, is a tectonic depression approximately 50 km wide, which formed during late Miocene transtension, and filled with non-marine clastic deposits. This depression is the loci of large number of volcanoes located along major bounding faults (Litherland and Aspden, 1992).

Some of the first contributions to the geology of the Ecuadorian Andes include those by Sauer (1971) who summarized his fieldwork. Other pioneers include Colony and Sinclair (1928), Sheppard (1937), Barrington (1938), Tschopp (1953), and Marchant (1961). Faucher and Savoyat (1973) updated Sauer's classical paper in their review of Ecuador paleogeography and orogenic events. Later a posthumous publication by Kennerley (1980) provided a significant advance in the understanding of the geological chronology of Ecuador. This was followed by the explanatory paper by Baldock and Longo (1982) of the geological map of Ecuador. Work by the British Geological Survey has been incorporated in a new geological map of Ecuador (Litherland *et al.*, 1994). This work includes a large number of new isotope ages, petrographic and geochemical analysis, and documentation of several terranes, which provides a more coherent picture of the tectonic evolution of the Ecuadorian Andes.

Tectono-stratigraphic evolution

Precambrian

There are no Precambrian rocks cropping out in the Ecuadorian Andes. However, xenocrystic zircon from analyzed rocks suggest that mafic and granitic magmas were contaminated during crustal ascent, or were partially derived from Guiana Shield crustal recycling (Noble *et al.*, 1997). Thus, inherited zircon ages from the Marcabeli Pluton (540 Ma, 2.22 Ga, and 2.876 Ga) as well as zircon upper intercept ages from Tres Lagunas granitoids (2.9 to 2.85 Ga) in the southern Cordillera Real are interpreted to represent detrital input from metasedimentary precursors. Furthermore, Nd-depleted mantle model ages (T_{DM}) for the Tres Lagunas and Marcabeli granitoids support 1.6 - 1.4 Ga ages for the protoliths (Noble *et al.*, 1997).

Similar Sm/Nd ratios and T_{DM} ages have been calculated by Restrepo-Pace *et al.* (1997) for Orinoquian granulite in the Colombian Eastern Cordillera. Priem *et al.* (1989) have reported similar Rb/Sr whole rock ages for the Parguazan tectonomagmatic episode for the Mitu Migmatitic Complex in the Guiana Shield. Although no granulites have been reported, isotope ages seem to suggest the presence of the Guiana Shield under the Ecuadorian Andes.

Paleozoic

Rocks of probable Lower Paleozoic age were overprinted by pervasive anatexis during the emplacement of Triassic S-type granite and extensive Jurassic cordilleran I-type granitoid (Aspden *et al.*, 1992). The rocks include quartzite, metasiltstone, graphitic schist, black phyllite and rare metagreywacke of the Chiguinda Unit which dominates the Southern Cordillera Real (Litherland *et al.*, 1994). Bedding and primary cleavage of this unit are subparallel and steeply dipping, and follow the Andean trend with evidence for more than one phase of deformation (Litherland *et al.*, 1994). Indeed, Kennerley (1980) correlated this unit with low-grade metamorphic phyllite and slate beds of the Salas Formation (Ordovician-Silurian) in the Olmos Massif in Peru (Mourier, 1988). To the NE, near Baños, there are also some 417 Ma pelitic schist beds of the Agoyan Unit, which have been correlated with the Chiguinda Unit by Litherland *et al.* (1994) who also reported reworked Early Ordovician acritarchs in the Jurassic Maguazo tectonic unit.

Upper Paleozoic rocks are present in the Cutucú High and in the Tahuin/Amotape Terrane. The oldest Paleozoic unit, the Devonian Pumbuiza Formation, consists of highly folded, grey to black slate with fine-grained quartz sandstone that form the core of the Cutucú High. Similar rocks have been reported to the W in the Tahuin/ Amotape Terrane (Martinez, 1970). This unit is unconformably overlain by the Macuma Formation (Lower Pennsylvanian) that consists of interbedded bioclastic limestone and black shale beds. These two units may represent shallow-water

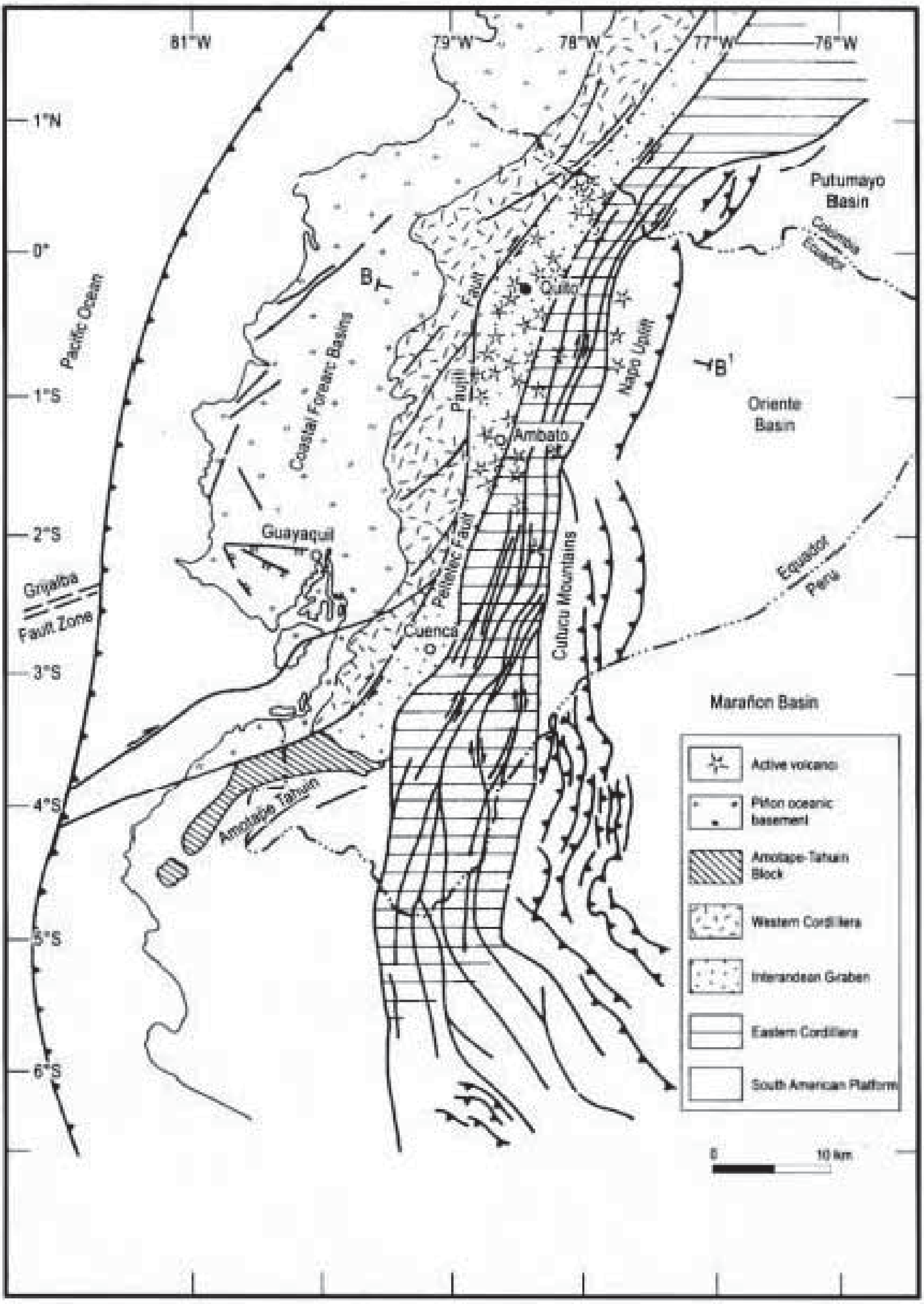


Fig. 12 - Major geological provinces of the Ecuadorian Andes with active volcanoes (based on Litherland et al., 1994).

facies equivalent to the thick Devonian to Pennsylvanian quartz-rich arkose and feldspathic wackes interbedded with shale, interpreted as a flysch sequence in the Tahuin/Amotape Block (Martinez, 1970; Mourier, 1988).

Litherland *et al.* (1994) described within the Loja Terrane, in the southern Cordillera Real, the Isimachi Unit of possible Upper Paleozoic age consisting of low-grade phyllite and marble. They also described the Monte Olivo amphibolite, which include an amphibolite dyke with relict igneous texture of Upper Devonian age (363 ± 9 Ma and 371 ± 10 Ma), and a garnet amphibolite showing Carboniferous ages (306 ± 10 Ma and 342 ± 23 Ma). These amphibolites are associated with the Chiguinda and Agoyan units of the Loja Terrane.

Mesozoic

Litherland *et al.* (1994) described the Triassic Piuntza unit as a continental/marine volcaniclastic sequence, between the Macuma and Santiago formations. This is equivalent to the Upper Triassic high-grade migmatite of the Sabanilla Complex, which is part of the Tres Lagunas S-type granite suite (227.6 ± 3.2 Ma) in the Cordillera Real, and the Moromoro Granite (220 ± 6 Ma) in the El Oro Metamorphic Belt (Litherland *et al.*, 1994). The Tres Lagunas blue quartz granite has been traced laterally into gneissic belts related to steep westward-dipping Andean-trending shear zones with widely developed S-C mylonites. Such shear zones suggest at least several phases of dextral transpression (Aspden *et al.*, 1992; Litherland *et al.*, 1994). To the S the Sabanilla Orthogneiss, consisting of foliated biotite-muscovite garnet granite, contains metasedimentary xenoliths in various stages of digestion with $^{87}Rb/^{86}Sr$ age of 224 ± 37 Ma (Litherland *et al.*, 1994). These S-type granites are interpreted to have formed during the separation of North and South America during the Triassic. However, they may represent a Late Triassic delayed phase collision between Gondwana and Laurasia followed by right-lateral shear. The pegmatitic and saussuritized Piedras Amphibolite (221 ± 16 Ma), originally a low-K basalt, is present within the El Oro Metamorphic Complex, and has been interpreted as a slice of metamorphosed oceanic crust (Litherland *et al.*, 1994).

Jurassic deposition started to the E of the present Eastern Cordillera with deposition of dark grey limestone units, calcareous sandstone and shale beds of the Santiago Formation (Lower Jurassic) (Geyer, 1974). This unit is overlain by more than 2500 m of varicolored shale beds and sandstone beds with thin zones of dolomite and gypsum of the Chapiza Formation. Basaltic to rhyolitic lava flows, interbedded with dacitic to rhyolitic ignimbrites and unwelded lapilli tuff of the Misahualli Formation (Middle Jurassic) cap this extensional cycle on the eastern flank of the Eastern Cordillera (Tschopp, 1953; Romeuf *et al.*, 1995). REE and multi-element scans of these calc-alkaline rocks suggest an arc-trench system (Romeuf *et al.*, 1995) and perhaps deposition in an ensialic back-arc setting. Thus, the Misahualli Formation (172.3 ± 2.1 Ma) is cut by the Abitagua Granite (162 ± 1 Ma). To the S these volcanic rocks are equivalent to the Zamora Batholith (171 ± 6 Ma and 152 to 180 Ma) as documented by Litherland *et al.* (1994). Furthermore, low initial Sr isotope ratios and major element analysis of the Late Jurassic Abitagua and Zamora granitoids suggest cordilleran I-type plutons representing the southern extension of the Colombia Magmatic Belt (Aspden *et al.*, 1992).

In the northern part of the Cordillera Real, there is a continuous and narrow belt (15 km wide) of the Jurassic Upano Unit. According to Litherland *et al.* (1994), this belt consists of andesitic greenstone, greenschist and metagreywacke intercalated with pelitic and graphitic schist with steep tectonic foliation. To the W, they also described grey to black, graphitic-muscovite schist of the Cayuja Unit in tectonic contact with metamorphosed black limestone beds and black calcareous phyllite, with some marble and calc-silicate rocks, of the Cerro Hermozo Unit. These units may represent metamorphosed equivalents of the Santiago, Chapiza and Misahualli formations. To the S, along the Cordillera Real, there occurs a sequence of agglomerate beds and green phyllite of volcaniclastic origin of the Jurassic Alao-Paute Unit (Litherland *et al.*, 1994). These authors also describe in fault contact, the Jurassic (Middle Callovian to Middle Oxfordian) Manguazo Unit, a slightly metamorphosed turbidite sequence interbedded with andesitic basalt, black phyllite and chert, interpreted as a marine fore-arc sequence.

A narrow zone of highly deformed, steeply-dipping ophiolitic rocks along the Peltetec Fault, on the western slope of the Cordillera Real, is interpreted as a mélange with island arc signatures and oceanic crust structure (Litherland *et al.*, 1994). The sheared metagabbro, metabasalt, foliated tremolitic serpentinite, spilitized dolerite, and volcaniclastic mélange are interpreted as a Jurassic subduction zone or suture similar to those described along the Romeral Fault in Colombia (McCourt *et al.*, 1984; Restrepo and Toussaint, 1988). Noteworthy, is the Raspas Formation in the El Oro Ophiolite Complex consisting of garnetiferous pelitic schist and layers of glaucophane schist, quartzite layers, eclogite (132 ± 5 Ma) and eclogite amphibolite (Feininger, 1980). This high pressure/low temperature unit was probably formed during accretion and clockwise rotation of the Tahuin Block.

Cretaceous sedimentation in the Cordillera Real started with deposition of the fluvial to shallow marine, quartz rich sandstone beds of the Hollin Formation. This was followed during the Middle Albian to Early Maastrichtian by deposition of dark grey shale and limestone interbedded with sandstone of the Napo Formation. A regional unconformity separates the Napo Formation from conglomerate, sandstone and shale of the Tena Formation (Maastrichtian/Paleocene). To the W, along the coast, the Piñon Terrane consist of basaltic pillow lava flows, hyaloclastite, dolerite, boninite, and cumulate gabbro of the Piñon Formation (Goossens *et al.*, 1977; Lebrat *et al.*, 1987; Van Thournout *et al.*, 1992). In turn the Piñon Formation is overlain by more than 2000 m of deepwater volcaniclastic sandstone, pyroclastic breccia, basaltic andesite lava flows and subordinate limestone, rich in organic material, of the Cayo Formation (Cenomanian to Maastrichtian). This formation is transitionally overlain by rhythmically-bedded silicified tuff and tuffaceous shale of the Guayaquil Formation (Late Maastrichtian to Early Paleocene).

In the Western Cordillera, the Cretaceous rocks consist of mafic lava flows and marine greywacke of the Toachi and Pilaton units, respectively (Macuchi Formation), and mainly shale, greywacke, tuff and andesitic sills and lava flows of the Yunguilla Formation (Campanian) and the Callo Rumi Formation (Maastrichtian) (Cosma *et al.*, 1998). To the S, in the Celica-Lancones Basin, E of the Tahuin-Amotape Block, there are massive andesitic lava flows, welded tuff beds, pillow breccias and agglomerates of the Celica Formation (Albian). The Celica Formation is overlain by more than 2000 m of coarse-grained deep-water volcaniclastic sediments of the Alamor Formation (Cenomian-Coniacian). Unconformably overlying these beds is a sequence of interbedded greywacke, shale and marl of the Naranjo Formation (Santonian/Campanian) and thin bedded greywackes and nodular limestone with conglomerate lenses of the Casanga Formation (Jaillard *et al.*, 1996).

Cenozoic

Deformation in the Cordillera Real is manifest as thick molasse deposition from Maastrichtian to Miocene of the Tena, Tiyuyacu, Orteguaza/Chalcana, Arajuno and Chambira formations in the Subandean Basin. However, the Western Cordillera and coastal basins contain Eocene to Miocene deep to shallow water siliciclastic sediments and carbonate rocks, deposited in a fore-arc setting, overprinted by strike-slip faults (Bourgois *et al.*, 1990).

The Cordillera Real, the Western Cordillera and the Inter-Andean Graben were sites of widespread, pervasive volcanism during the Tertiary. A higher percentage of calc-alkaline rocks is present in the Central and Southern Andes suggesting a thicker crust (Hörmann and Pichler, 1982). Basalt, andesite and dacite of the Eocene Tandapi Formation (Cosma *et al.*, 1998) are the dominant rocks of the Western Cordillera. Miliolid-rich limestone beds of the Unacota Formation (middle Eocene) and its lateral equivalent deep-water volcaniclastic sediments of the Apagua Formation, overlie this formation.

The middle Eocene is unconformably overlain by Oligocene volcaniclastic sediments and lava flows of the Saraguro Formation (Eguez, 1986; Bourgois *et al.*, 1990). In the Cordillera Real, Lavenu *et al.* (1992) described calc-alkaline volcaniclastic and lava flows of the Sacapalca (Paleocene-Eocene), the Saraguro (Oligocene), the Pisayambo (Mio-Pliocene) and the Cotopaxi (Plio-Quaternary) volcanic events. Post-Oligocene Saraguro, Pisayambo and Cotopaxi volcanic events are also present along the Inter-Andean Graben and consist of volcaniclastic sediments and lava flows.

Today, most active volcanoes are along the Peltetec and Pujili bounding faults. Miocene-Pliocene reactivation of these major faults generated pull-apart basins along the Inter-Andean Graben. These basins are filled by volcaniclastic sediments derived from the Cordillera Real and deposited in fluvial and lacustrine environments from the middle Miocene onward. Late Miocene to Pliocene compressional deformation of the Andean Orogeny, has inverted these basins (Barragán *et al.*, 1996; Hungerbuehler *et al.*, 1996).

Crustal growth, tectonic evolution and structural styles of the Ecuadorian Andes

As indicated by inherited older zircons ages from the Marcabeli, Tres Lagunas Granite and El Oro Complex, crustal growth in the Ecuadorian Andes may have begun during the collisional Precambrian Orinoquian (Grenville) Orogeny. Indeed, Nd-depleted mantle model ages (T_{DM}) from the Tres Lagunas and Marcabeli granitoids indicate a 1.6 - 1.4 Ga age for the protoliths (Litherland *et al.*, 1994; Noble *et al.*, 1997). This collisional belt may be a continuation of the granulite belt in Colombia, where similar Sm/Nd ratios and T_{DM} ages have been calculated (Restrepo-Pace *et al.*, 1997).

Like the Precambrian event, the Ocloyic Orogeny (Late Ordovician) is poorly age-constrained since Lower Paleozoic rocks have undergone pervasive metamorphism and anatexis, often with schistosity following the Andean trend (Fig. 7). However, Kennerley (1980) has correlated the Chiquinda Unit of the Cordillera Real with the low grade Ordovician phyllite and slate from the Olmos Massif. Furthermore, reworked Ordovician acritarchs have been reported in the Jurassic Maguazo tectonic unit (Litherland *et al.*, 1994).

A possible Late Devonian to Lower Carboniferous tectonic event is also suggested by amphibolite dykes (363 ± 9 Ma) and garnet amphibolite (342 ± 23 Ma) associated with the Chiquinda and Agoyan units of the Loja Terrane (Litherland *et al.*, 1994). This event may correlate with an incipient Gondwana-Laurasia assemblage (Berry *et al.*, 1997). Mixed siliciclastic and carbonate sedimentary rocks of Devonian and Permo-Carboniferous age are reported in the Cutucú Mountains and Tahuin/ Amotape Terrane. In the Tahuin/Amotape Terrain these rocks have undergone significant Late Paleozoic folding and thrusting related to final Late Paleozoic Pangea assemblage.

Late Triassic anatectic processes are recorded in the Tres Lagunas and Moromoro S-type granites of the Cordillera Real and El Oro Metamorphic Belt. S-C mylonites in these plutons suggest several episodes of dextral transpression (Litherland *et al.*, 1994) when North America rifted away from South America (Aspden *et al.*, 1992). However, the Piedras amphibolites (221 ± 16 Ma) within the El Oro Metamorphic Complex suggest a collisional event (Feininger, 1980) during the closure of an oceanic basin. Alternatively, S-type granites could represent collisional plutons, and S-C mylonites could have been formed thereafter by transpressional processes. This tectonic event has been named as the Moromoro Orogenic Event (Litherland *et al.*, 1994).

An arc-trench was established during the Middle Jurassic as interpreted from the presence of cordilleran I-type Zamora (171 ± 6 Ma and 180 to 152 Ma) and Abitagua (162 ± 1 Ma) granites (Litherland *et al.*, 1994). Such granites were coeval with ensialic back-arc extension and thick red-bed deposition interbedded with bimodal volcanics. The subduction zone may have been situated along the Peltetec Fault where inliers of sheared metagabbro, metabasalt, foliated tremolitic serpentinite, spilitized dolerite and volcaniclastic rocks occur. The assemblage may represent an accretionary prism. Subduction continued in the S in the El Oro Ophiolite Complex, which contains foliated eclogite

(132 ± 5 Ma) described by Feininger (1980). Indeed, this deformation may be related to a 110° clockwise rotation of the Amotape/Tahuin Block, followed by dextral shearing prior to the Neocomian (Mourier *et al.*, 1988).

Perhaps the most important tectonic event in the Ecuadorian Andes evolution was related to the diachronous emplacement of mafic to ultramafic oceanic back-arc and intra-oceanic arc sequences of the Piñon Terrane (Goossens *et al.*, 1977; Feininger and Bristow, 1980; Lebrat *et al.*, 1987; Cosma *et al.*, 1998). This event has been identified as the Peruvian Phase of the Andean Orogeny (Peltetec Orogeny). Paleomagnetic data suggest large-scale (70°) clockwise rotation of the Piñon Terrane (Roperch *et al.*, 1987). Regionally, docking of this allochthonous terrane was accompanied by Late Cretaceous uplift and erosion of the proto-Cordillera Real as can be seen by the presence of the Tena Formation molasse of Late Maastrichtian age, and by inversion of half grabens in the Oriente Basin (Marksteiner and Alemán, 1997). This collision, manifest by major shearing and thrusting of Mesozoic rocks in the Eastern Cordillera, was accompanied by Barrovian-type upper amphibolite to greenschist facies metamorphism and the resetting of isotopic ages (Litherland *et al.*, 1994). The main Late Cretaceous suture was located along the Pujili Fault where an antiformal duplex of allochthonous rocks has been emplaced in the Western Cordillera. Basement involved deformation began and continued with different intensities throughout the Plio-Pleistocene. However, W of the Cosanga Fault, a stack of nappes involving crystalline basement was emplaced. To the S, in the Celica-Lancones Basin, this orogenic event is represented by an abrupt change in provenance from a dominantly volcanic provenance of the Alamor and El Naranjo formations to basement-derived conglomerate of the Cosanga Formation.

Termination of Piñon Terrane accretion was accompanied by westward trench migration, increased convergence rates (Pardo-Casas and Molnar, 1987) and the emplacement of the calc-alkaline Tandapi Arc (Eguez, 1986; Cosma *et al.*, 1998) during the Incaic Phase of the Andean Orogeny. This event was coeval with deposition of miliolid-rich limestone units around the arc and deepwater volcaniclastic beds in the Western Cordillera. The Cordillera Real was uplifted and the Oriente Basin was the site of Eocene Tiyuyacu molasse deposition. Inversion of half grabens, compressional deformation and uplift of the Napo and Cutucú Mountains continued (Marksteiner and Alemán, 1997). Excellent evidence for this event is found to the W of Macas, where 42 Ma aplite dykes cut mesoscopic folds in the Hollin Formation. To the W, along the coastal basins, the deepwater quartz-rich sandstone of the Ancon Group, containing abundant Piñon Terrane olistoliths record this orogenic event.

The Oligocene was a time of relatively tectonic quiescence and low convergence rate described as the Aymara Phase of the Andean Orogeny (Pardo-Casas and Molnar, 1987; Soler and Bonhomme, 1990). The Oligocene Saraguro volcanic arc sequence suggests continuous eastward arc migration (Lavenu *et al.*, 1992). This volcaniclastic sequence was deposited along the Western Cordillera and the western flank of the Cordillera Real.

The Andean Orogeny started as a consequence of the break-up of the Farallon Plate about 25 Ma (Handschumacher, 1976). Oblique compression and relatively high rates of plate convergence (Pardo-Casas and Molnar, 1987) account for early Miocene reactivation of the Peltetec and Pujili faults and the development of pull-apart basins along the Inter-Andean Graben (Litherland and Aspden, 1992). These basins filled with fluvial and lacustrine deposits derived from the East, suggesting that the Western Cordillera did not reach its present elevation until the Pliocene (Hungerbuehler *et al.*, 1996). Subduction of the Grijalva Fracture Zone (Lonsdale, 1978) coincides with the northern limit of low angle subduction and volcanic gap that characterized the Ecuadorian Andes segmentation (Hall and Wood, 1985; Gutscher *et al.*, 1999). Indeed, Miocene volcanic activity was restricted to the southern and central Ecuadorian Andes (Lavenu *et al.*, 1992).

Late Miocene/Pliocene uplift and inversion of transtensional basins along the Inter-Andean Graben was marked by large scale thrusting, inverse faulting and folding as determined by apatite fission track dating (Hungerbuehler *et al.*, 1996). This deformation was accompanied by continuous volcanic activity located along the present volcanic front from Cuenca in the S to central Colombia in the N (Barberi *et al.*, 1988). Additional fission track studies along the Abitagua Batholith suggest transpressional uplift in the last 5 Ma. During this time, the volcanic axis shifted eastwards to the Cordillera Real. Since the early Quaternary the volcanic belt has widened to the Western Cordillera and along the length of the Inter-Andean Graben (Fig. 11b) where a large number of volcanoes are present (Barberi *et al.*, 1988). Pliocene to Quaternary volcanoes along this depression contain andesites with a clear calc-alkaline affinity and lower K_2O content than those of the Cordillera Real (Barberi *et al.*, 1988). This may be related to deep crustal fractures controlling the emplacement of volcanoes along the Inter-Andean Graben. Large volumes of Holocene lahars formed during dome collapse, which melted part of the volcano's icecap and transformed rapidly into a debris flow. These lahar deposits have been documented by Mothes *et al.* (1998), in the Cotopaxi Volcano (4700 years BP) and by Samaniego *et al.* (1998) in the Cayambe Volcano (4000 - 650 years BP).

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