Tectonic Evolution of the Andes

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The Andes is the type section of a non-collisional orogen that formed a mountain chain by the subduction of oceanic crust under a continental plate, as proposed by Dewey and Bird (1970). However, the analysis of the different sectors of the Andes shows a great variety of processes and a vast complexity, including mountain-building processes, which do not fit into the original definition. Therefore, the Andes will be divided following the proposal of Gansser (1973) to examine these tectonic processes (Fig. 1). The proposed division, although descriptive, is appropriate to analyze the distinctive history of the different Andean segments.

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 chain developed along the Pacific margin by an Andean-type subduction; and the Alleghanides, related to the closure 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, and is punctuated by the collision of island arcs in the Northern Andes. The last stage is responsible for 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.

The proto-margin of Gondwana

It is well established that the Gondwana Supercontinent was formed by accretion and amalgamation of different continental blocks and terranes as the result of the Panafrican-Brasiliano Orogeny (Brito Neves and Cordani, 1991; Hoffman, 1991; see recent review of Brito Neves *et al.*, 1999). Most of the western Gondwana margin was formed at that time, although accretion of different small cratonic blocks to this proto-margin, mainly during the Early Paleozoic (Ramos, 1988a; Restrepo-Pace, 1992; Toussaint, 1993), contributed to the final configuration of the present Pacific margin. Some of these cratonic blocks were exotic to the continent, whereas others were parautochthonous or perigondwanan. Perigondwanan terranes are those that were part of western Gondwana during Proterozoic times, but were split away and subsequently accreted to the margin (Ramos and Basei, 1997a; Keppie and Ramos, 1999).

The different terranes identified in the basement of the Andes are shown in Figure 2. In order to reconstruct the proto-margin of Gondwana it is necessary to subtract all the allochthonous terranes accreted during Phanerozoic times. Firstly, the oceanic terranes accreted during the Cenozoic in the Northern and Caribbean Andes such as the Choco Terrane of the Serranía de Baudó and its northern extension in Panamá (Dengo and Covey, 1993), and the Bonnaire Block of Venezuela (Kellogg and Bonini, 1982). This block formed by a series of nappes and oceanic rocks, detached and thrust from the Caribbean Plate, contains slivers of metamorphic rocks that have been defined as the Caribbean Terrane by Bellizzia and Pimentel (1994). This terrane was incorporated to South America during Late Cretaceous-early Paleogene times.

Secondly, it is required to subtract the Mesozoic Piñón-Daua terranes of the Ecuadorian and Colombian Andes (Alemán and Ramos, this volume), and their extension in the Amotape-Tahuín Terrane (Feininger, 1987; Reynaud *et al.*, 1999). However, due to the important sinistral strike-slip displacements produced by the oblique subduction during Meso-Cenozoic times, a palinspastic restoration is needed, similar to the proposed reconstruction of the Alleghenides by Pindell (1985). In this reconstruction the Santa Marta and Santander massives, as well as part of the Central and Eastern Cordilleras should be located farther S (present coordinates) of their present position.

This restoration permits the reconstruction of a large piece of basement named the Chibcha and related terranes (Etayo-Serna and Barrero, 1983; Alemán and Ramos, this volume) or Central Andean terranes by Restrepo-Pace (1992). One common feature of the basement of these terranes is the Grenville signature, based on the Pb-isotope composition (Ruiz *et al.*, 1999), as well as the geochronology and Nd-isotope data of the Colombian terranes (Restrepo-Pace *et al.*, 1997). This large terrane is presently the basement of Eastern Cordillera, and it was interpreted as an allochthonous terrane derived from North America by Forero Suárez (1990), that during the early closure of the Iapetus Ocean was transferred to the proto-margin of Gondwana in pre-Emsian times (pre-late Early Devonian). The magmatic arc at that time was developed on the Paleozoic terrane. The Borde Llanero-Guaicáramo Fault that uplifted the Eastern Cordillera foothills on the foreland basin deposits approximately coincides with the suture between the Paleozoic allochthons and the proto-margin of Gondwana.

Paleozoic rocks in the Ecuadorian Andes consist of semipelitic schist and gneiss, highly deformed and intruded by S-type Triassic granitoid plutons (Aspden and Litherland,

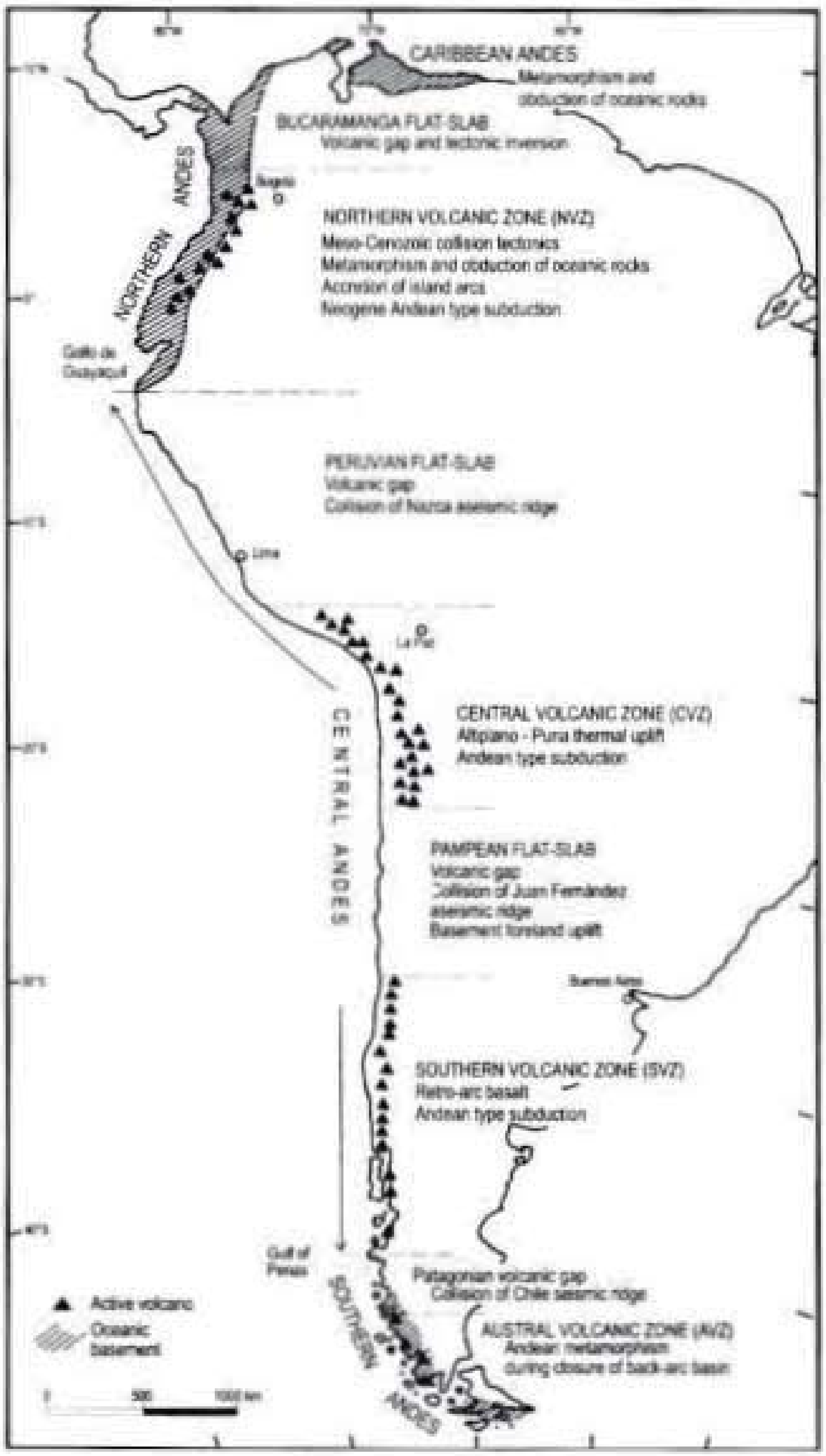
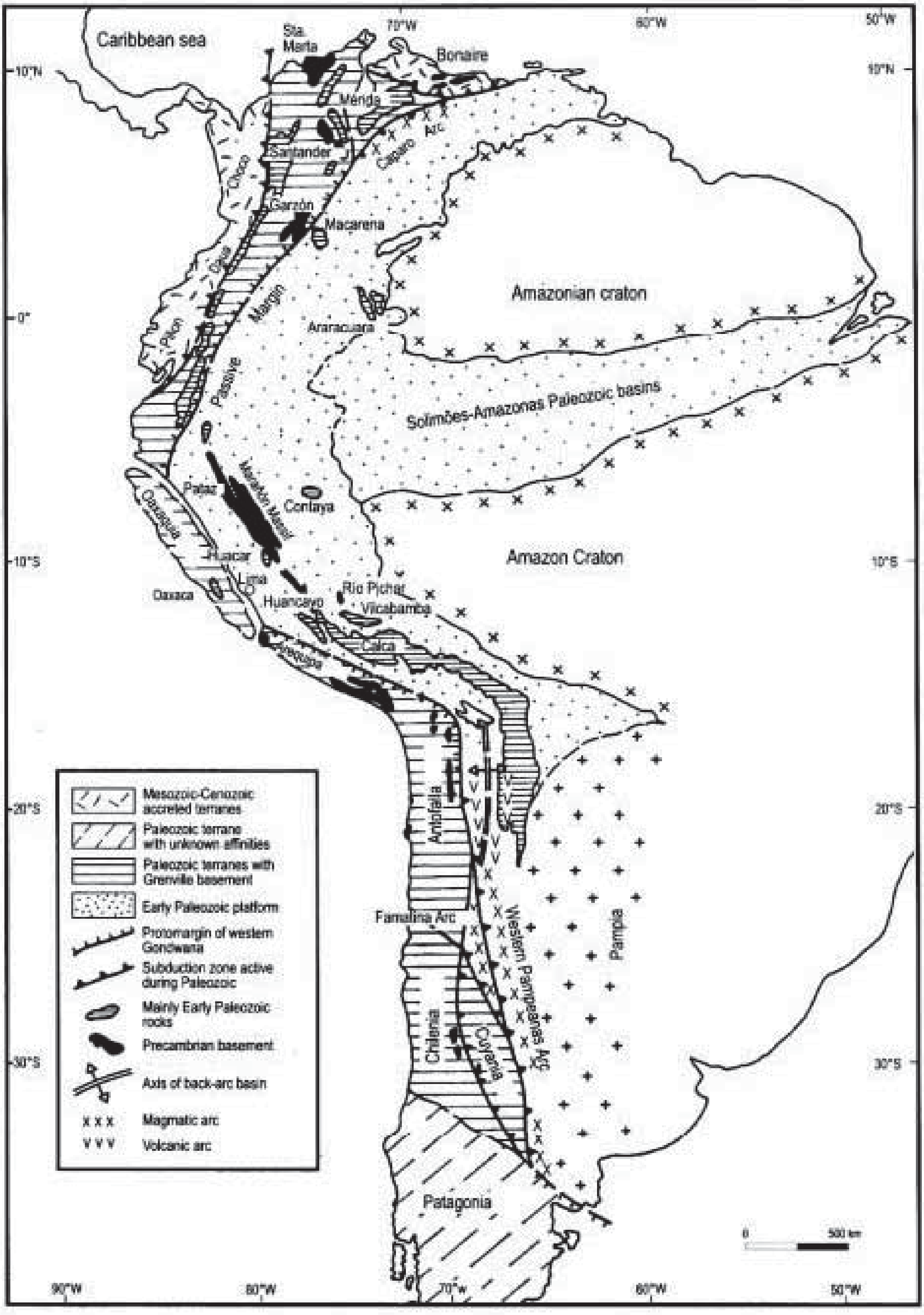


FIGURE 1: Main segments of the Andes modified after Gansser (1973), with indication of major tectonic processes involved during their formation.

FIGURE 2: Reconstruction of the proto-margin of western Gondwana with indication of the main Precambrian and Early Paleozoic blocks underlying the Andean cover (modified after Bellizzia and Pimentel, 1994; Restrepo-Pace, 1992; Restrepo-Pace et al., 1997; Keppie and Ortega Gutiérrez, 1999; Ramos, 1988a).


Caribbean sea
Sta. Marta
Bonaire
Mérida
Santander
Caparo Arc
Chocó
Garzón
Macarena
Dagua
Piñon
Passive Margin
Amazonian craton
Araracuara
Solimões-Amazonas Paleozoic basins
Pataz
Marañón Massif
Contaya
Amotape
Huacar
Lima
Oaxaca
Huancayo
Rio Pichar
Vilcabamba
Calca
Arequipa
Amazon Craton
Antofalla
Famatina Arc
Western Pampeanas Arc
Pampia
Chilenia
Cuyania
Patagonia
Mesozoic-Cenozoic accreted terranes
Paleozoic terrane with unknown affinities
Paleozoic terranes with Grenville basement
Early Paleozoic platform
Protomargin of western Gondwana
Subduction zone active during Paleozoic
Mainly Early Paleozoic rocks
Precambrian basement
Axis of back-arc basin
X X X Magmatic arc
V V V Volcanic arc
0
500 km
90°W
80°W
70°W
60°W
50°W
10°N
0°
10°S
20°S
30°S

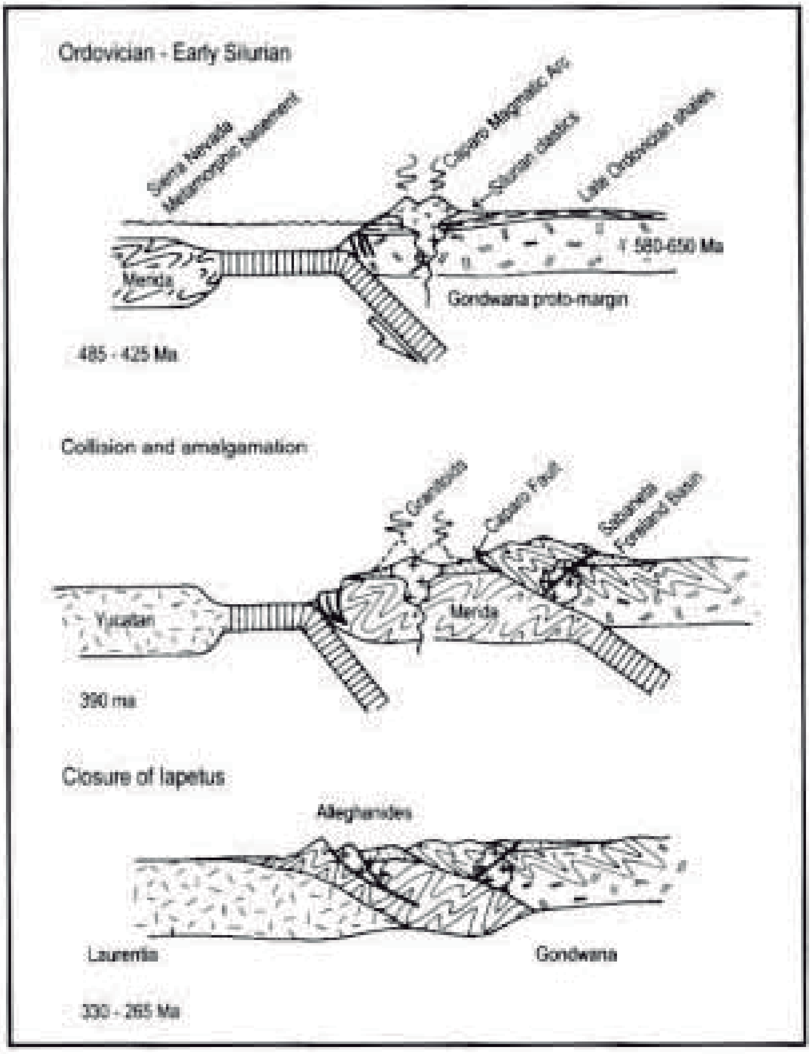


FIGURE 3: Schematic evolution of the Mérida Terrane, and the pre-Late Permian accretion of the Yucatán Block during the closure of the Iapetus Sea (modified after Bellizzia and Pimentel, 1994; Pindel, 1985).

1992). They could be relicts of the allochthonous terranes described farther N in Colombia, deformed during the successive collisions that affected the Eastern Cordillera or Cordillera Real.

Farther S in Peru, the basement of the Arequipa Massif is truncated, either by tectonic erosion due to subduction (Von Huene and Scholl, 1991), or by detachment of a large piece of basement, as was proposed by Dalziel (1994). The location of the Oaxaquia Terrane in Figure 2, is one of the alternatives proposed by Keppie and Ortega Gutiérrez (1999) to explain the gap along the margin between the Arequipa Massif and the Chibcha and related terranes of Colombia.

This alternative shows a fringe of Grenville terranes overlain by Early Paleozoic deposits. Some of them such as the Oaxaquia Terrane, have as in Oaxaca, a Late Cambrian-Early Ordovician cover with *Parabolina argentina* (Robison and Pantoja-Alor, 1968), a typical Gondwana trilobite very frequent in northern Argentina (Harrington and Leanza, 1957). The correlation between Northern Argentina, and the Oaxaca fauna is also enhanced by the fauna found by Moya *et al.* (1993) in northern Puna of Argentina over the Arequipa-Antofalla Terrane.

Another point to be noted, is the Paleozoic cover of these terranes. There are Cambro-Ordovician trilobite-bearing sequences as in the Garzón Massif (Ruiz *et al.*, 1999) or graptolite-bearing as in Central Cordillera (Mojica *et al.*, 1988), partially preserved in metamorphic facies and associated with Ordovician magmatic rocks. These terranes are bounded to the E by a Cambrian platform in Venezuela and Colombia (Bordonaro, 1992), and by more extensive and well-described Ordovician platform deposits from northern Argentina to Venezuela. This shallow water platform developed on the Gondwana autochthonous basement faced deep-water deposits as proposed in the paleogeographic reconstruction of Moya (1988) and Aceñolaza (1992).

Based on these characteristics this fringe of basement terranes of Grenville affinities are interpreted as part of the Rodinia Supercontinent (Hoffman, 1991; Dalziel, 1991 and subsequent papers), that were attached to the Amazonian Craton after collision and amalgamation. The active margin was located in the Sunsas Belt of Amazonia, whereas Laurentia was the passive margin as proposed by Sadowski and Bettencourt (1994). The separation between Laurentia and Amazonia left behind some Grenville terranes in South America. This late Proterozoic rifting and the subsequent Early Paleozoic extension were responsible for separation of part of the terranes. For example Oaxaquia, left western Gondwana after the Early Ordovician, while other blocks remained as perigondwanan terranes (Ramos and Basei, 1997b). Part of the Grenville sutures were the foci of new extension in Cambro-Ordovician times developing a passive margin, and the resulting subsidence formed the extensive clastic platform that characterized the Ordovician paleogeography from Venezuela to northern Argentina (Aceñolaza, 1992).

There is an exception along this Early Paleozoic passive margin in the Caparo Arc developed in eastern Venezuela (Bellizzia and Pimentel, 1994). This area shows magmatic rocks emplaced in deformed Ordovician and Silurian fossiliferous deposits that are unconformably overlying autochthonous Precambrian basement (650 - 580 Ma) of western Gondwana (Marechal, 1983). The syn-tectonic granitoid rocks range from Early Ordovician to Early Silurian (495-425 Ma), whereas the post-tectonic granites are Devonian (Fig. 3). This magmatic activity recorded the approximation and collision of the Mérida Terrane by the end of the Ordovician and beginning of the Silurian (Bellizzia and Pimentel, 1994). Later on, once amalgamated to western Gondwana, the Mérida Terrane was affected by deformation and the Sabaneta Formation was deposited in a peripheral foreland basin. Subsequent deformation followed the closure of the Iapetus Ocean and the collision of Laurentia with the northwestern sector of Gondwana to form the Alleghenides in the Late Permian (Pindel, 1985).

Some of the detached basement blocks such as the Chibcha and related terranes, were separated by oceanic crust, and later after subduction (Fig. 4) re-amalgamated to western Gondwana (Restrepo-Pace *et al.*, 1997).

Extension along the Arequipa Massif and its southern prolongation in the Antofalla basement was probably controlled by the previous Grenville-age suture (Suárez-Soruco, 1999). This extension formed an important Ordovician to Devonian basin along southern Peru, Bolivia and northern Argentina (Sempere, 1995). In northern Bolivia there is no evidence of Ordovician oceanic crust, whereas to the S there are oceanic rocks (Allmendinger *et al.*, 1982; Bahlburg and Hervé, 1997).

The history of the southern Arequipa-Antofalla Block is complex, as it records a late Proterozoic collision during Pampean Orogeny (approximately 530 Ma) (Omarini *et al.*, 1999), subsequent extension during Cambrian and Ordovician times (Ramos, 1988a), and final amalgamation during Late Ordovician (Coira *et al.*, 1982, 1999; Dalziel and Forsythe, 1985), resulting in the Ocloyic Orogeny (Ramos, 1986; Bahlburg and Hervé, 1997).

The western Puna magmatic arc was active from Late Cambrian to Middle Ordovician times (Palma *et al.*, 1987; Bahlburg, 1990). The inception and northern extension of this Early Paleozoic subduction zone, is not that evident, and it could be underlying the thick Andean cover.

The Early Paleozoic proto-margin of northwestern and western Argentina is better constrained. The development of a calc-alkaline series of granitoid rocks of Late Cambrian to Middle Ordovician age in western Sierras Pampeanas, has permitted the identification of an Early Paleozoic magmatic arc (Ramos, 1988a; Toselli *et al.*, 1996; Pankhurst *et al.*, 1998). The distribution and magmatic characteristics show two different belts: one with volcanic and plutonic rocks in the Famatina Terrane (Toselli *et al.*, 1996; Quenardelle and Ramos, 1999), and another restricted to western Sierras Pampeanas (Rapela *et al.*, 1999). Figure 5 summarizes the tectonic evolution and subsequent amalgamation of these terranes. This magmatic arc was related to the accretion of

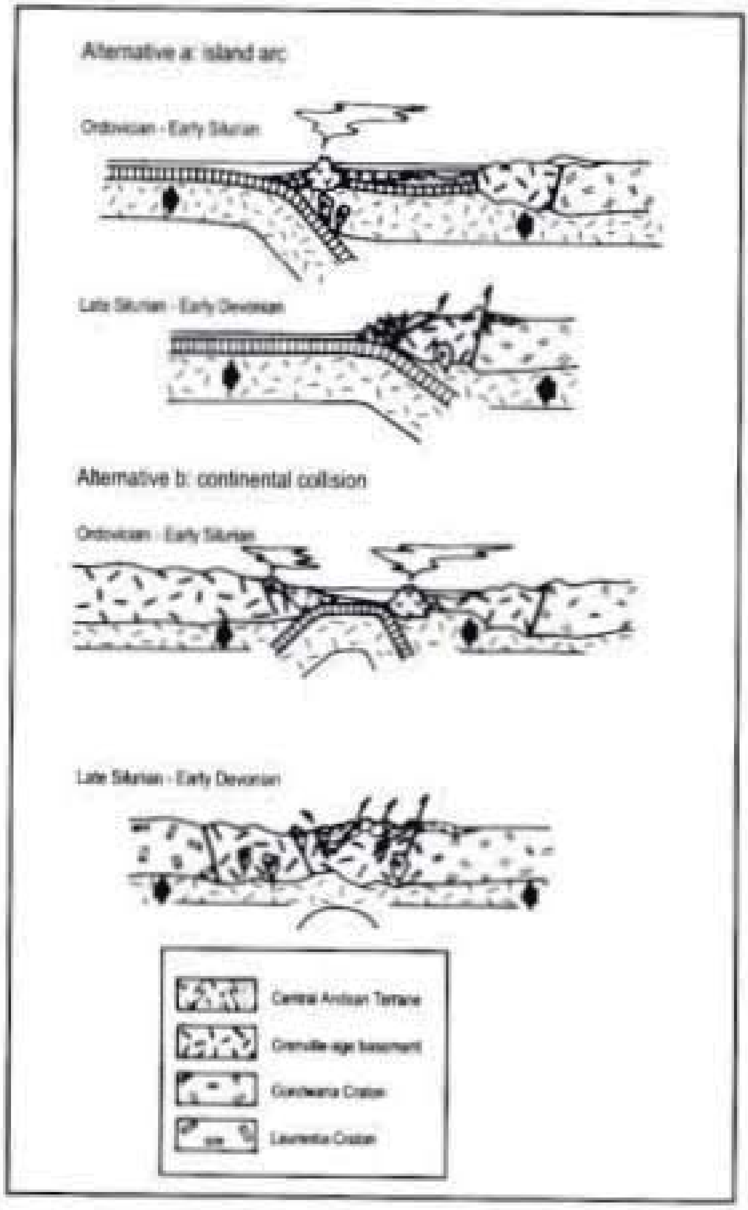


FIGURE 4: Different scenarios to explain the magmatic activity in the Chibcha (Central Andean) and related terranes of Colombia (modified after Restrepo et al., 1997).

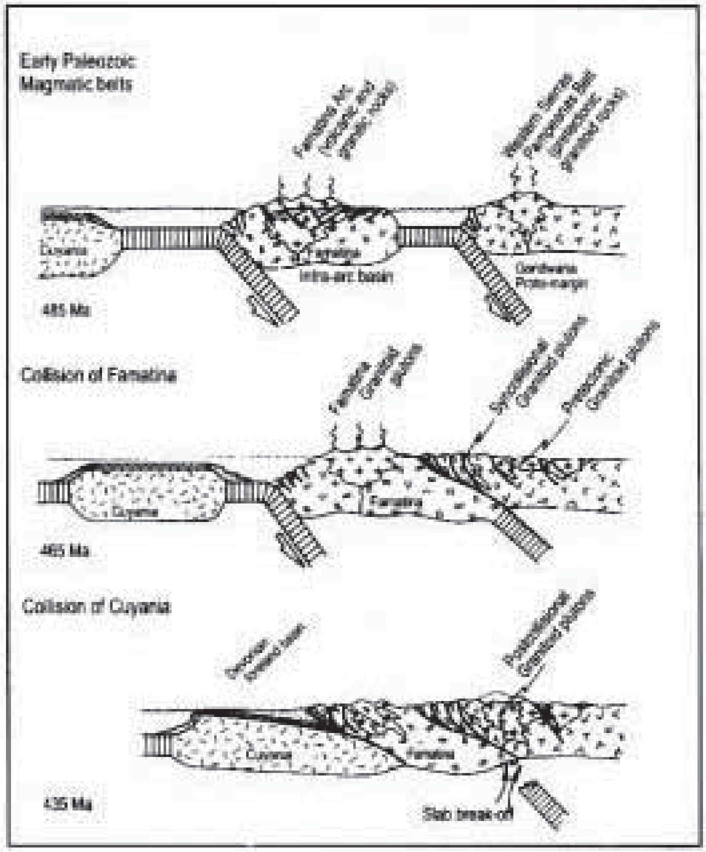


FIGURE 5: Tectonic evolution of western-central Argentina during Early Paleozoic times (modified after Quenardelle and Ramos, 1999).

FIGURE 6: Major acid and bimodal provinces developed along the western Gondwana margin during Late Paleozoic-Early Jurassic times. On the cratonic side mafic rocks dominated, while more acid rocks formed on a fringe of Early Paleozoic terranes along the proto-continental margin (modified after Mpodozis and Kay, 1992; Hervé, 1988).

the Cuyania composite terrane, containing the Precordillera, an allochthonous carbonate platform derived from Laurentia and accreted to the Gondwana proto-margin during Middle Ordovician times (Astini and Thomas, 1999). Another Grenville basement block, known as the Chilenia Terrane, collided against the western margin of Precordillera during Devonian times (Ramos *et al.*, 1986).

Farther S in the Patagonian Andes the Precambrian basement is scarce. Most of the gneiss and medium grade metamorphic rocks assumed to be Precambrian have U/Pb Paleozoic ages (Ramos and Aguirre-Urreta, this volume). These rocks are equivalent to the accretionary prism of Late Paleozoic age preserved along the Pacific coast (Hervé, 1988).

Alleghanian and Gondwanan Orogenies

The Late Paleozoic rocks of the Andes record two important mountain-building events: the Alleghanian Orogeny in the Northern Andes and the Gondwanan Orogeny in the Southern Andes.

The Alleghany Orogen was proposed by Woodward (1958) to encompass what in the old literature was known as the "Appalachian Revolution", responsible for the final deformation and uplift of the Appalachian System. As a consequence of the final closure of the Iapetus Ocean, Laurentia collided against western Gondwana, the present Africa and South America continents. The Alleghanides were diachronously formed between Late Carboniferous and Permian, encompassing a time span from 330 to 265 Ma (Hatcher *et al.*, 1989).

The Gondwanan Orogeny was first proposed by Keidel *(1916, 1921), to include the Late Paleozoic deformation* of Sierra de la Ventana in the extra-Andean province of Buenos Aires, and the previously adjacent Cape Town System (Keidel, 1916, 1917). The Gondwanides were reconstructed from eastern Australia, through western Antarctica, South Africa to finally join the Ventania System and the Precordillera de Cuyo in the present foothills of the Andes (Du Toit, 1937; Groeber, 1938).

The Northern Andes

The final collision and amalgamation of the Mérida and other terranes into the proto-margin of Gondwana, was one of the most evident effects of the Alleghanian Orogeny in the Andean basement of Venezuela (Figs. 2 and 4). This deformation predated the Late Permian and was interpreted by Pindell (1985) as the result of final docking of the Yucatán Terrane, trapped between Laurentia and Gondwana during the closure of the Iapetus Ocean. Early Carboniferous rocks of the Mérida Cordillera underwent important deformation (Marechal, 1983). The compressive deformation was followed by transtension with widespread granitic plutonism from Early Permian to Triassic (290-225 Ma). This transtension was related to counterclockwise rotation between western Gondwana and North America, as proposed by Rapalini and Vizán (1993) on paleomagnetic grounds.

Farther S in the Colombian Andes, the Alleghanian event is characterized by strike-slip deformation related to the final closure of the Iapetus Ocean, associated with some S-type granitoid plutons (McCourt and Feininger, 1984).

The Central Andes

The Eo-Hercynian deformation of southern Peru and Bolivia proposed by Mégard *et al.* (1971), is temporally restricted to a Late Devonian-Early Carboniferous transpression (Díaz Martínez, 1996). This transpression has been detected farther S in the Tarija Basin (Fernández Seveso *et al.*, 1993) and in the basins of west-central Argentina. In the Paganzo and Río Blanco basins, at both sides of the proto-Precordillera, Limarino *et al.* (1999) recognized during the Late Paleozoic two different stages. A foreland basin stage as a result of the Chanic deformation at the end of the Devonian and beginning of the Early Carboniferous, that persisted until the Late Carboniferous; an extensional phase associated with alkaline basalt in the Early Permian. The first stage could be related to transpression as the result of oblique convergence and block rotation as identified by Rapalini and Vilas (1991) based on paleomagnetic data.

The Late Paleozoic along the Pacific margin was characterized by an important magmatism. It is represented by alkali-basalt and associated leucogranite and rhyolite of the Mitu Group (260-190 Ma) in southern Peru and northern Bolivia (Kontak *et al.*, 1985). These Permian to Triassic rocks are associated with red-beds in rift basins.

There is also a series of important batholiths and volcanic rocks formed during Late Paleozoic-Early Triassic times in southern Peru, Chile and central Argentina (Kontak *et al.*, 1985; Kay *et al.*, 1989). The magmatic suites have been divided in a subduction related Carboniferous-Early Permian series, and an extensional mainly granitic and rhyolitic series known as the Choiyoi Province (Mpodozis and Kay, 1992; Llambías and Sato, 1995).

Along the Principal Cordillera a strong angular unconformity is related to the San Rafael Orogeny (Ramos *et al.*, 1996b), which caused strong deformation in the Late Carboniferous-Early Permian rocks, separated by an angular unconformity of the Choiyoi volcanic rocks. This diastrophism produced an important penetrative deformation in the Late Paleozoic rocks of the Frontal Cordillera, but in the adjacent Precordillera the effects are only seen by an increase in subsidence rates in the late Early Permian basins as described by Fernández Seveso *et al.* (1993).

The Late Paleozoic marked the beginning of the subduction in the Pacific margin for the first time along the present trench. Most authors have related the Late Paleozoic deformation to changes in the intensity and direction of the convergence vector as inferred by paleomagnetic data (Ramos, 1988b; Kay *et al.*, 1989). Another tectonic alternative was proposed by Mpodozis and Kay (1992). These authors related the deformation of the San Rafael Orogeny to the docking of an enigmatic *terrane X*, which is not presently preserved due either to tectonic erosion or strike-slip displacements along the continental margin.

The subduction related magmatism was followed by an extensional regime. The extensional magmatic activity can be tracked (Fig. 6), as well as the younger Jurassic associated rocks, from Peru all along the western margin of Gondwana trough Antarctica to eastern Australia (Mpodozis and Kay, 1992).

Southern Andes

The Gondwanides Orogeny on the western slope of the Patagonian Andes is characterized by an extensive accretionary prism where high pressure-low temperature metamorphic rocks are cropping out (Hervé, 1988). These rocks farther N are partially preserved due to the tectonic erosion of the fore-arc region (Stern, 1991). Late Paleozoic magmatic rocks are poorly exposed, mainly in the eastern slope of the cordillera, where scarce tonalite and other granitoid rocks have Late Carboniferous-Early Permian ages (Ramos, 1983).

Farther S along the present coast, minor exotic terranes have been identified in the Madre de Dios region by Mpodozis and Forsythe (1983), who confirmed the hypotheses proposed previously by Helwig (1972). Mafic and ultramafic oceanic rocks compose the accreted rocks, and fusulinid-bearing platform carbonates associated with flysch and pelagic chert facies, which were docked into the Gondwanan subduction complex. Recent paleomagnetic studies demonstrate conspicuous rotation after their emplacement (Rapalini *et al.*, 1999). These oceanic exotic rocks may have been derived from platform carbonate patches coeval with the Copacabana Limestone developed at the latitude of present Bolivia.

The Early Mesozoic extension

Most of the Andes were dominated by extension at the end of the Permian and Triassic times. This extension was the precursor of the Pangea break-up. The Northern Andes continental margin (Fig. 7) was the conjugate rifted margin of the Yucatán Block and of the extensive present platform of U.S. Gulf and eastern Mexico. The reconstruction of the North American margin is difficult due to the intense crustal attenuation distributed along 760 km-wide zone (Pindell, 1985). However, the South American rifted margin is even more difficult to reconstruct due to the superimposed deformation produced by the Cretaceous and younger oceanic terrane accretion in Colombia and Ecuador, and the Caribbean deformation during emplacement of the Bonaire Block. These regions were displaced during Cenozoic times by important wrenching that obliterated the original margin.

The Northern Andes continental margin was developed during Triassic-Jurassic times in a series of *en échelon* rifts of NE-SW trend, such as the Espino, Uribante, Barquismeto, Mérida, and Machiques (Fig. 7), among others, which concentrated active extensional faulting until Late Jurassic times (Parnaud *et al.*, 1995). The tectonic inversion of the Machiques Rift exposed in the Sierra de Perija felsic to mafic igneous rocks, interfingered with red-beds, as well as the characteristic red-bed sequences of the La Quinta Formation of the Barquismeto and Mérida rifts, cropping out in the Serranía de Trujillo and the Mérida Andes (Lugo and Mann, 1995).

Small plutons along the Magdalena Valley of Colombia are precursor of the Late Triassic extension. Conglomerate, breccia, and sandstone form the initial fill of half-graben systems such as the Bogotá Trough, with a northeastern trend similar to previously described rifts farther N (Mojica and Dorado, 1987; Mojica *et al.*, 1996).

Highly deformed S-type granites of Late Triassic age are coeval with continental to volcaniclastic sequences in Ecuador (Litherland *et al.*, 1994). Subsequent terrane collision and closure of back-arc basins have obliterated original paleogeography.

The foreland region of Peru shows an intricate pattern of Triassic rifts, superimposed on the Late Paleozoic extension. Thick sequences of evaporites and red-beds of Late Triassic to Jurassic age of the Pucará Group constitute the initial syn-rift deposits (Mathalone and Montoya, 1995). This rifting postdates the Mitu Group rifts and alkaline magmatism.

Farther S, minor alkaline basalt of Entre Ríos (233 Ma) and Tarabuco (171 Ma) in Bolivia are associated with the Triassic and Early Jurassic rifting (Sempere, 1995). Near Lake Titicaca the red-beds fill the Mitu Group grabens, and along the foreland Subandean areas fluvial deposits of the Serere Formation of Late Triassic-Early Jurassic age are linked to mild extension.

The Triassic rift system of the Andes of Argentina and Chile has a dominant NW-SE trend (Charrier, 1979; Uliana and Biddle, 1988). The Triassic extension followed the rifting that began with the Choiyoi Volcanics, and was developed over the fringe of Paleozoic terranes accreted to cratonic South America (Kay *et al.*, 1989). The age of the normal faulting progressed from N to S. It was Late Carboniferous to Permian in northern Chile, Late Permian in Central Argentina and Triassic to Early Jurassic in Patagonia, predating the opening of the South Atlantic Ocean (Mpodozis and Ramos, 1990).

The rift system was filled by red-beds, interfingered with volcanic rocks of bimodal composition in northern and central Chile (Suárez and Bell, 1993; Mpodozis and Cornejo, 1997). Thick sequences of red-beds and lacustrine deposits, interfingered with alkaline basalt (233 Ma) are well represented in the Cuyo Basin (Kokogián *et al.*, 1993). Farther S these occur in the subsurface of the Neuquén Embayment along both sides of the Colorado River and in the Huincul area.

The geometry of the Triassic-Early Jurassic rifting of the Andes as a whole, denotes three different regions controlled by the extension and characteristics of the basement terranes (Fig. 7). The Northern Andes trend NE-SW in an *en échelon* pattern, matching the extension with a counterclockwise rotation developed during the development of the continental margin. The central region of Peru and Bolivia has a rift system with a dominant NW-SE trend, developed in the hanging-wall of the suture between the Arequipa Terrane and the Amazonian Craton. This suture developed in Grenville times indicates that Arequipa subducted beneath the Amazonian Craton, that was the active continental margin at that time (Sadowski and Bettencourt, 1994). The NW-SE rifting direction in the southern region was heavily controlled by the basement fabric. The inception of most of the rifts has been located along the suture hanging-wall of the previous Paleozoic accreted terranes (Ramos and Kay, 1991).

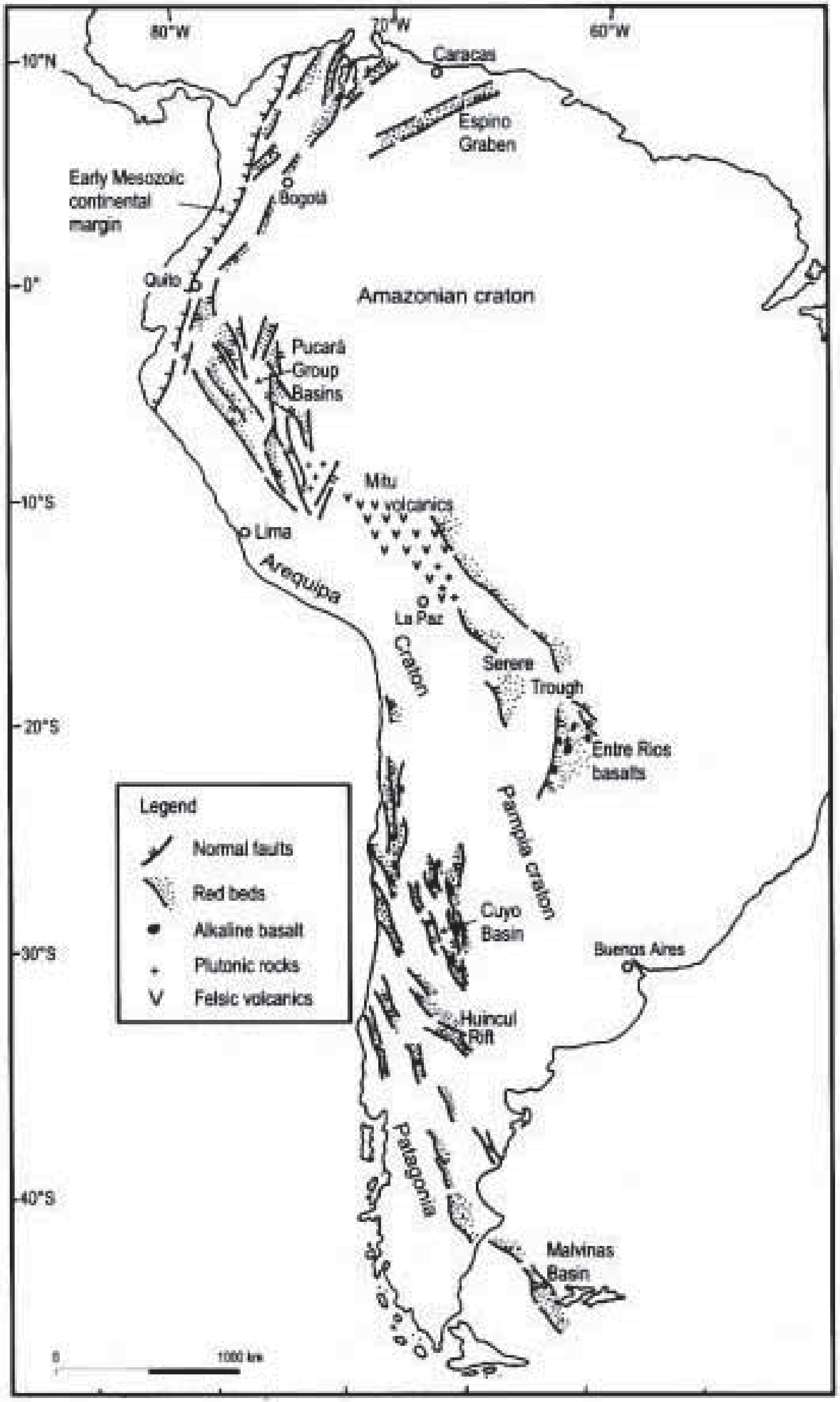


FIGURE 7: Major rift trends developed during the Pangea break-up along the margin of the Northern and Central Andes, as well as the generalized extension of the Southern Andes during Late-Triassic and Jurassic times. Configuration of the continental margin prior to the accretion of oceanic terranes is indicated in the Northern Andes. (modified after Parnaud et al., 1995; Mojica and Dorado, 1987; Daly, 1989; Mathalone and Montoya, 1995; Uliana and Biddle, 1988).

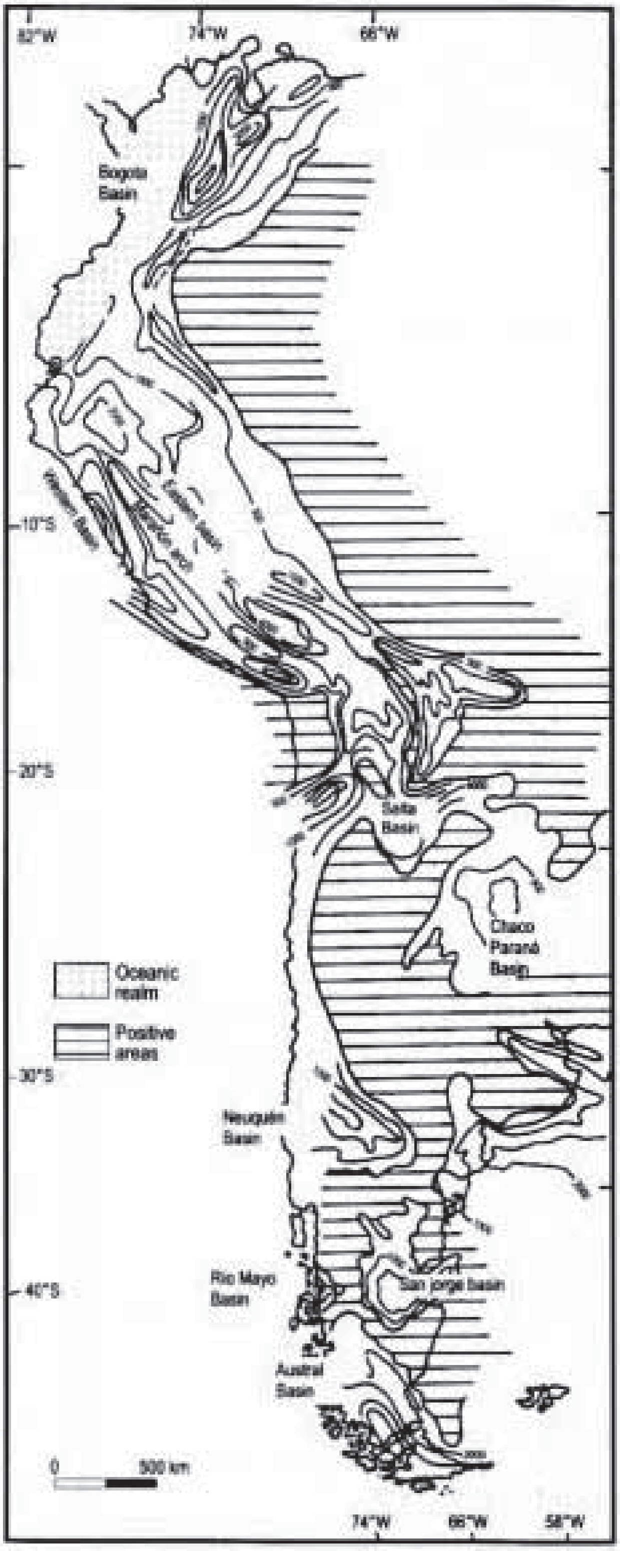


FIGURE 8: Cretaceous paleogeography of the Andes with isopach contours in metres (modified after Macellari, 1988; Ramos and Aguirre-Urreta, 1994 Salfity and Marquillas, 1994).

Passive margin and retro-arc Thermal Subsidence Stage

The entire Andean system records during Jurassic and Cretaceous times the development of a complex series of fore-arc, intra-arc and retro-arc basins that have been recently reviewed by several authors (Zambrano, 1987; Mojica and Dorado, 1987; Salfity, 1994; Benavides-Cáceres, 1999; Villamil, 1999; Jaillard *et al.*, this volume). The crustal fabric, heavily controlled by the previous accretionary history governs this complexity. Although the transition from active extensional faulting to thermal subsidence had occurred in Cretaceous times, the details and precise timing vary from one place to another. Some of these sag sequences have remained without deformation until Tertiary times, while others have had important precursor orogenies, during the Cretaceous, predating the Andean deformation.

Soon after the Late Jurassic most of the Northern Andes continental margin went into thermal subsidence developing a passive margin (Fig. 8), where the Cretaceous marine sediments were accommodated (Pindell, 1985). Isopach maps indicate that subsidence was maximum along the earlier depocenters such as Machiques, Barquismeto and Uribante rifts (Macellari, 1988). The end of the Early Cretaceous was marked by an important paleo-oceanographic change, consisting of a generalized drowning of the carbonate platform and deposition of intervals rich in organic matter in an anoxic environment that extended into the foothills at the Barinas-Apure basins (Erlich *et al.*, 1999). These environments were interrupted by the uplift of the Central and Eastern Cordilleras during Campanian to Maastrichtian times (Villamil, 1999). The platform deposits bounded the Brazilian Craton with important depocenters such as the Upper Magdalena Valley (Etayo-Serna, 1994) or the Bogotá Trough, where several thousand metres of Cretaceous sediments were deposited (Macellari, 1988).

Farther S in the Ecuador retro-arc area, marine platform sedimentation was important from Albian times (Macellari, 1988; Jaillard *et al.*, 1995). In Peru, two distinct longitudinal depocenters are recognized: the Eastern Basin along the Brazilian Craton and the Western Basin, separated by the Marañón Arch extended from 6°S to 14°S (Benavides-Cáceres, 1999). This shelf sedimentation was coeval with the arc volcanism along the margin.

Active rifting with continental deposition persisted from the Late Jurassic to the Early Cretaceous in Bolivia in a distal retro-arc setting (Sempere, 1995). Isolated depocenters of the Puca Group red-beds were interfingered with alkali basalt. The thermal sag sedimentation began in Albian times, when marine limestone beds overlap the Andean depocenters.

The active extensional faulting in eastern Bolivia and northwestern Argentina was younger than farther N in the Andes, and it was closely related to the opening of the South Atlantic Ocean. This Early Cretaceous extension was active from the volcanic arc in Chile (Mpodozis and Allmendinger, 1993), to central Argentina where the old Brasiliano sutures were reactivated, as well as along the Atlantic margin (Fig. 9).

This extension was coeval with the Salta rift basin in northern Argentina where thick syn-rift deposits and related

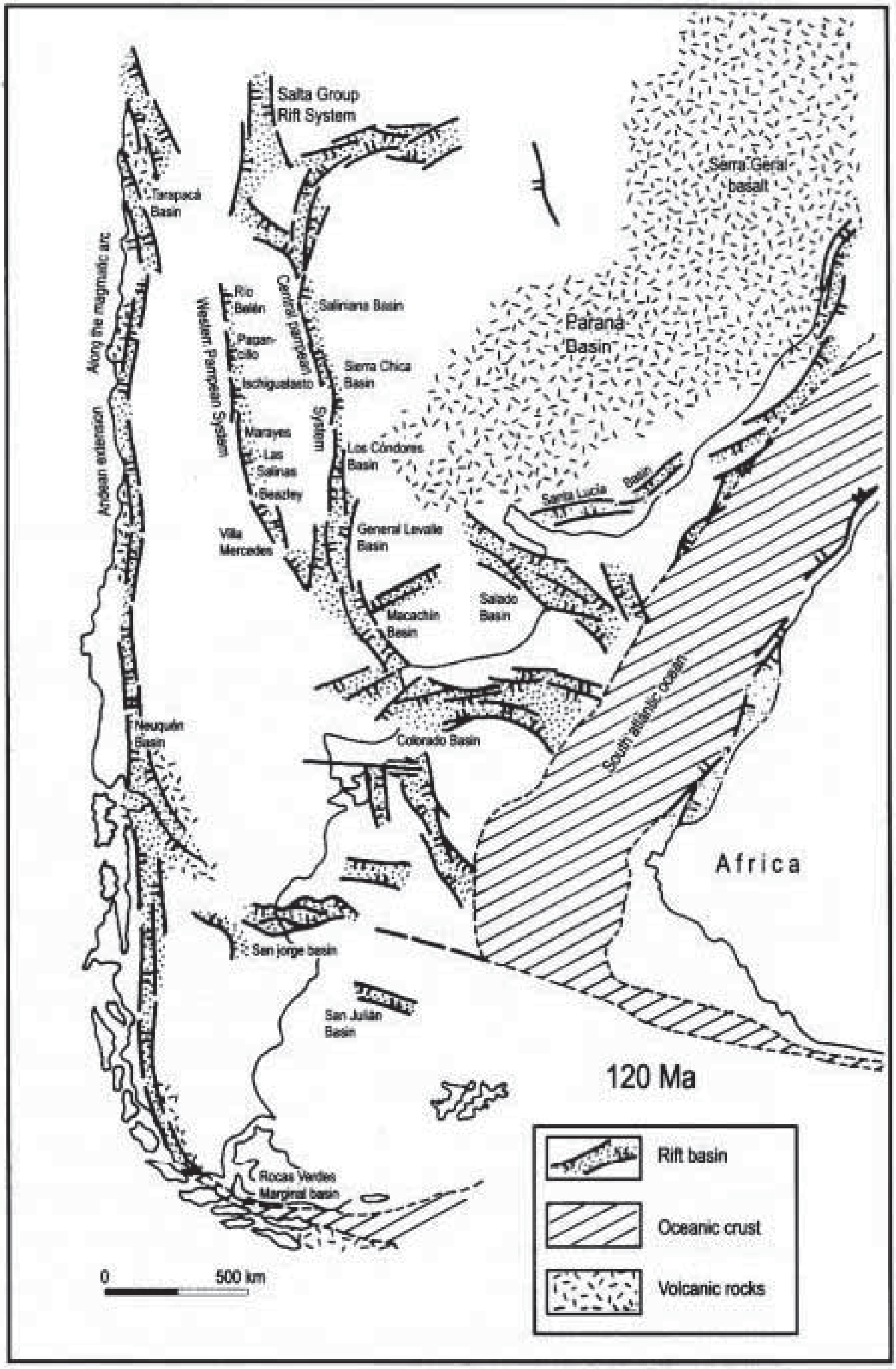


FIGURE 9: Early Cretaceous rifting in southern South America coeval with the early opening of the South Atlantic (modified after Salfity and Marquillas, 1994; Uliana and Biddle, 1988; Rossello and Mozetic, 1999).

alkali-basalt of the Pirgua Subgroup have been deposited in different sub-basins (Salfity, 1982; Salfity and Marquillas, 1994), with variable trend controlled by the Precambrian fabric of the Pampia Craton (Comínguez and Ramos, 1995). Thermal subsidence and widespread shallow marine to lacustrine sedimentation in these sub-basins took place in Campanian to Paleocene times.

Along the Pacific margin several retro-arc basins were filled with clastic and carbonate deposits (Fig. 8) such as the Neuquén Basin (Uliana and Legarreta, 1993), Río Mayo Basin (Ramos and Aguirre-Urreta, 1994), and the Austral or Magallanes Basin (Biddle *et al.*, 1986). The axis of the Austral Basin was the locus of the Rocas Verdes marginal basin (Dalziel *et al.*, 1974) that extended from Cordillera de Sarmiento at about 52°S to S of the Beagle Channel in Tierra del Fuego Island, at the southernmost Andes.

Accretion and collision along the Northern Andes

The western Northern Andes consist of several superimposed orogens as a result of the obduction of oceanic terranes (Restrepo and Toussaint, 1973), that may have been formed far from the continent. Their allochthonous nature was suspected since the early work of Barrero (1979) in western Colombia and by Henderson (1979) in Ecuador. Many authors have proposed different tectonic settings for the ophiolite sequences, tholeiitic basalt, and pelagic and hemipelagic sediments, that constitute these oceanic terranes (Feininger, 1982, 1987; McCourt *et al.*, 1984; Lebrat *et al.*, 1985; Mégard, 1987). The number and boundaries of these exotic terranes vary considerably from a maximum of 34 terranes (including the Paleozoic terranes, Etayo-Serna and Barrero, 1983) to two or three major composite terranes (Henderson, 1979; McCourt *et al.*, 1984; Restrepo and Toussaint, 1988). However, there is a general agreement in the allochthonous oceanic nature of western Colombia and Ecuador, and in the source area for these terranes in the Caribbean Plate during an early stage of its evolution (Burke, 1988; Meschede and Frisch, 1998).

The Caribbean Andes also record an important accretion of oceanic rocks derived from the Caribbean Plate that constitutes the termination of the Andes along the Atlantic Ocean.

There are some important differences among the oceanic terranes from N to S, due to the different petrological and geochemical characteristics and the timing of the accretion.

Accretion of the Caribbean Andes

The motion of the Farallon Plate changed during the Campanian from a northeasterly to a nearly eastward direction (Engebretson *et al.*, 1985). As a result, the relative movement between North America and South America became slightly convergent (Pindell and Barret, 1990), and due to the eastward displacement of the Caribbean Plate during Late Cretaceous to Eocene times, part of the oceanic floor of this plate was obducted in the Atlantic margin of northern South America.

The Caribbean Orogeny was related to the accretion of these oceanic rocks, emplaced orthogonally to the Mérida Andes during the latest Cretaceous, and continuing into the early Eocene. The late Eocene molasses were deposited in an E-W trending foreland basin that was reactivated during the Oligocene, developing a complex system of nappes, folded nappes, and deformed syn-orogenic deposits (Stephan, 1982).

The oblique subduction along the Pacific margin, increased the northward displacement of basement blocks such as the Santa Marta and the Guajira blocks, producing a complex system of strike-slip faults in the late Cenozoic (Fig. 10).

The interaction of the N-S trending Mérida Andes and the orthogonal trend of the nappes produced the stress partitioning manifest by the E-W right-lateral displacements of the Oca and El Pilar faults, and the left-lateral offsets of the Bucaramanga and Santa Marta faults.

Most of these faults, and mainly the Boconó Fault, display important neotectonic activity (Schubert and Vivas, 1993).

Accretion of western Colombia

Most authors have recognized two major accretionary episodes in the Cretaceous and another one during the Neogene (Restrepo and Toussaint, 1988; Mégard, 1987) that are recorded in the ophiolite sequences and in blue-schist exposed on the western side of the Central Cordillera, in the Western Cordillera and in the Serranía de Baudó (Fig. 11).

The first tectonic unit corresponds to several dismembered ophiolite complexes that are emplaced with an eastern vergence on the western slope of Cordillera Central (Bourgois *et al.*, 1985). There are exposed in Jambaló (Fig. 11), where these authors described high-pressure metamorphic rocks with a probable Paleozoic protolith, but with a whole-rock age of 125±15 Ma (Orrego *et al.*, 1980), which is interpreted as the Early Cretaceous age of obduction. Several other bodies emplaced roughly along the Romeral Fault have similar ages (Mégard, 1987).

The second tectonic unit corresponds to the Dagua Terrane exposed in the Western Cordillera, and in some tectonic klippes of the Central Cordillera. The thrust sheets are 1.5 to 5 km thick and are folded by later deformation and also affected by subvertical strike-slip faults (Bourgois *et al.*, 1987). These rocks known as the Diabase Group are oceanic low-K tholeiite, with ages based on paleontological and geochronological data, ranging from Albian to Maastrichtian (Barrero, 1979). Most authors favour an oceanic flood basalt origin for the Diabase Group of the Dagua Terrane (Millward *et al.*, 1984, Lapierre *et al.*, 1999). The cross-cutting relationships between the Antioquía Batholith (59-57 Ma) and the ophiolite in the Yarumal Complex constrain the emplacement of this second tectonic unit between 80 and 60 Ma (Restrepo and Toussaint, 1988; Toussaint and Restrepo, 1994).

The third tectonic unit consists of the Choco Terrane, that includes the oceanic terranes W of the Atrato Suture (Fig. 11), and mainly exposed in the Serranía de Baudó (Dengo, 1983). This terrane was formed by island-arc tholeiitic assemblages, emplaced in an oceanic plateau at approximately 78 - 72 Ma (Kerr *et al.*, 1997a). This oceanic

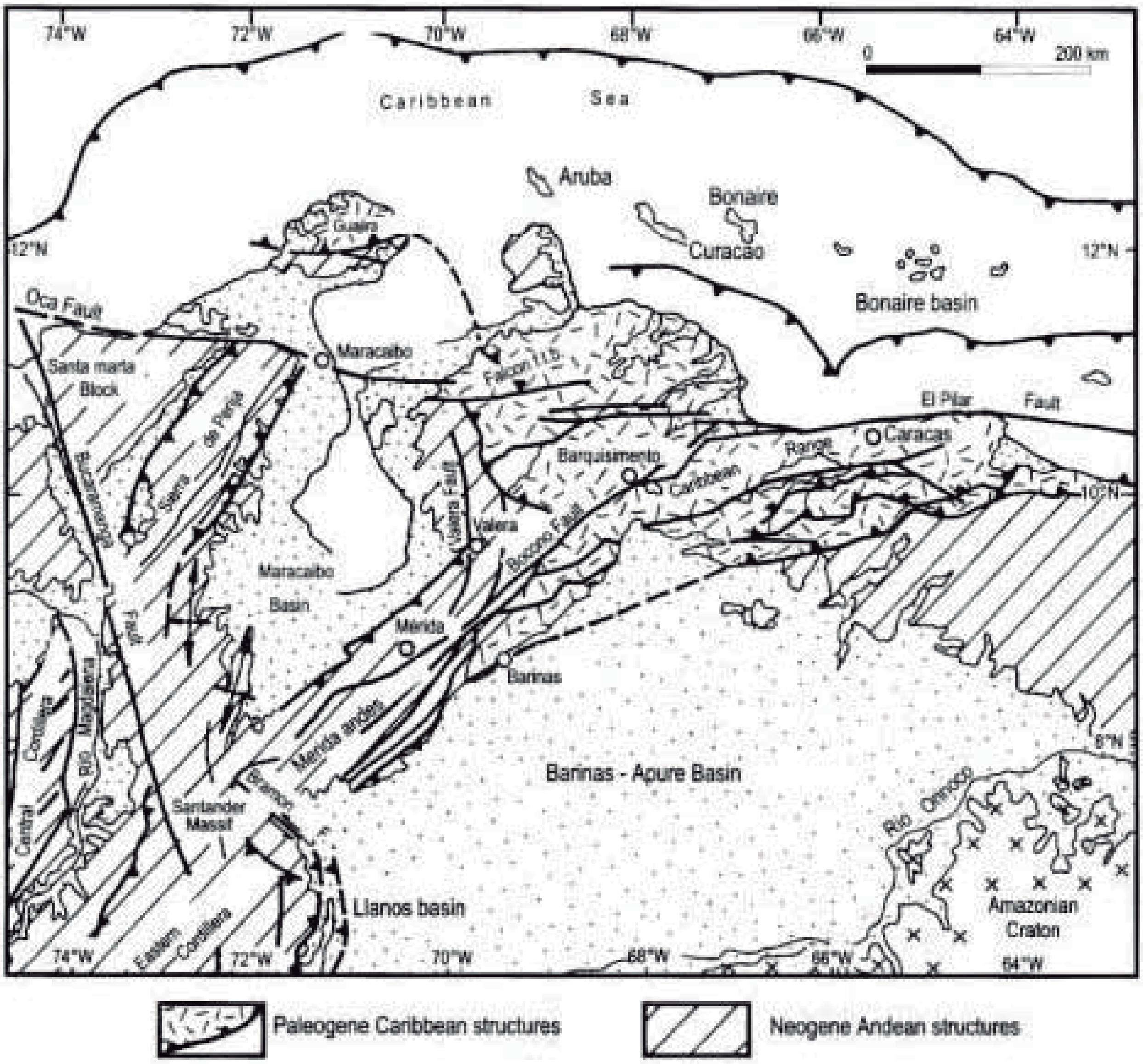


FIGURE 10: Present tectonic setting of the Caribbean Andes. Note the right-lateral offset produced in the nappes front by the Boconó strike-slip fault (modified after Colletta et al., 1997).

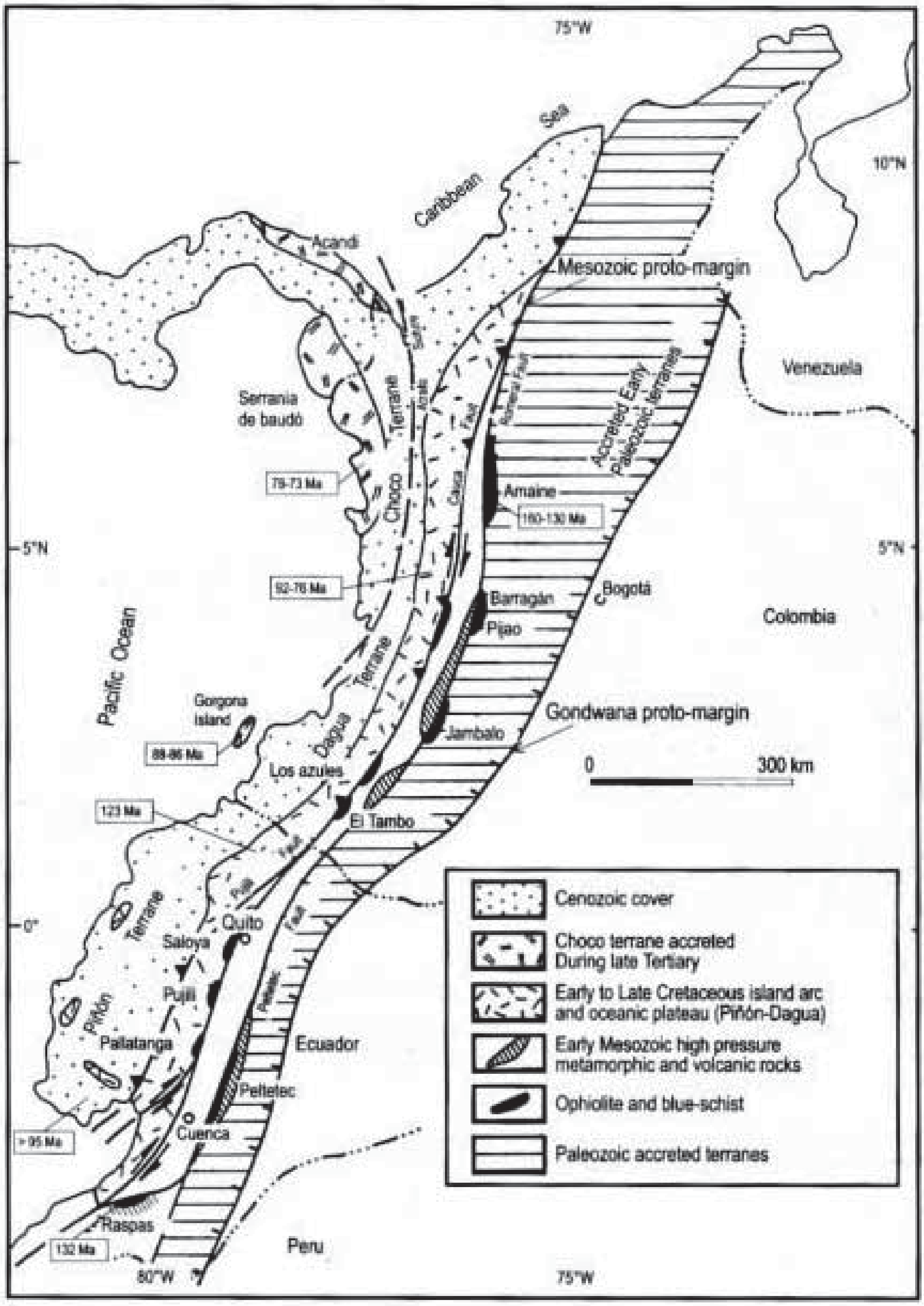


FIGURE 11: Accreted oceanic terranes of the Northern Andes with indication of main ophiolites and high pressure metamorphic rocks exposed along the Cauca Valley of Colombia and the Interandean Depression of Ecuador (modified after McCourt et al., 1984; Millward et al., 1984; Van Thornout et al., 1992). Radiometric ages correspond to the oceanic basement (modified after Reynaud et al., 1999).

basement, covered by pelagic to hemipelagic sequences of Late Cretaceous to Miocene age, was accreted to continental South America during a short episode of obduction between 12.9 to 11.8 Ma (Duque-Caro, 1990). According to this author the planktonic assemblages found in the Paleocene cover of this terrane have Central American affinities, which corroborate a source in the Caribbean Plate of this terrane (Meschede and Frisch, 1998).

Based on the existing data it is interpreted that an island arc terrane, active during the Jurassic to Early Cretaceous time, was accreted to South America, as indicated by the calc-alkaline magmatic arcs developed along the Central Cordillera (Colletta *et al.*, 1990). Remnants of this oceanic terrane are now preserved on the western slope of Central Cordillera W of the Early Mesozoic margin, and are bounded by the Romeral Suture (Fig. 12).

A most prominent and best-preserved obduction corresponds to the Calima Orogeny (Barrero, 1979) that produced an important Alpine deformation during Late Cretaceous-Paleocene times. This deformation was transmitted through the Central Cordillera to the Magdalena Valley and coincides with the tectonic inversion of the western flank of Eastern Cordillera (Etayo-Serna, 1994; Restrepo- Pace, 1999). This deformation abruptly terminated with marine deposition in the retro-arc basins (Cooper *et al.*, 1995).

The Paleogene arc at this time was situated on the western slope of Western Cordillera (Toussaint and Restrepo, 1982). After a Paleogene period of oblique subduction the third obduction took place in the middle Miocene, with the emplacement of the Choco Terrane, that produced reactivation and folding of the older structures and intense deformation in the foothills of the Eastern Cordillera. The basin axes steadily migrated from the Maastrichtian in the Upper Magdalena Valley on the western margin of Eastern Cordillera to the Subandean foothills in the early to late Oligocene (Villamil, 1999). In middle Miocene times, deformation extended to the Llanos Basin developing a complex structure where thin-skinned deformation was followed by thick-skinned basement faults, as seen in the interaction of the Cusiana and Yopal fault systems (Cooper *et al.*, 1995).

After the emplacement of these oceanic terranes, Andean-type subduction in the present trench began in the late Miocene. The magmatic arc at first situated in the Western Cordillera migrated during Pliocene times to the Central Cordillera (Toussaint and Restrepo, 1982). However, probably due to the shallowing of the Wadati-Benioff Zone to the N, active volcanism in the late Cenozoic arc ends in Cerro Bravo, at 5°N (Méndez-Fajuri, 1989), developing the Bucaramanga flat subduction segment (Pennington, 1981).

Accretion of western Ecuador

Western Ecuador also recorded a complex history of accretion, preserved in the Western Cordillera and the adjacent coastal plains, W of the Pujili Fault, as well as in the Eastern Cordillera (or Cordillera Real) E of the Peltetec Fault (Fig. 11).

The most complex structures are in the Cordillera Real because Mesozoic and Paleogene deformation as well as late Miocene strike-slip obliterated the original relationships. Aspden and Litherland (1992) have reconstructed the tectonic settings of a series of metamorphic units and undeformed volcanic rocks of Triassic and Jurassic ages. These settings encompass ocean floor and island-arc assemblages for the Alao Terrane, and Jurassic arc rocks for the Misahualli andesite and dacite, as well as calc-alkaline granitoid rocks for the Abitagua Stock. A simplified tectonic evolution is shown in Figure 13.

The Peltetec Fault on the western flank of Cordillera Real is associated with a melange with island-arc signature, where the sheared metagabbro, metabasalt and foliated serpentinite has been interpreted as a subduction complex of Jurassic - Early Cretaceous age (Litherland *et al.*, 1994). These rocks are correlated with the ophiolite sequences exposed along the Romeral Suture in Colombia, as well as the Raspas metamorphic complex (132 Ma), and associated ophiolitic rocks exposed on the northern flank of the Amotape-Tahuin terranes (Feininger, 1982; Reynaud *et al.*, 1999).

The Western Cordillera oceanic rocks of the Piñón Terrane are interpreted as having developed in an oceanic plateau of 123 Ma (Sm/Nd age, Lapierre *et al.*, 1999). An island arc system was developed in this Early Cretaceous oceanic basement, known as the Macuchi Arc of Late Cretaceous - Paleocene age in northern Ecuador (Henderson, 1979; Van Thournout *et al.*, 1992). The Piñón Terrane had in the Manabi area along the coast in southern Ecuador, the San Lorenzo Arc (77-60 Ma), and to the E in the Guayaquil area the Cayo Arc (92-80 Ma), two oceanic island-arc systems of Late Cretaceous age (Jaillard *et al.*, 1995). These authors favour the development of a marginal basin between these two arcs, being the Cayo volcanics a remnant arc during latest Cretaceous.

The collision of the Piñón Terrane against the continental margin was diachronic. It started in the southern part by Maastrichtian times, and it was completely amalgamated during early Eocene time (Reynaud *et al.*, 1999). The andesite, dacite and breccia of the Tandapi Arc indicate a continental calc-alkaline setting (Cosma *et al.*, 1998). The Baudó oceanic rocks of the Choco Terrane are not present in the coastal plains of Ecuador.

East of the accreted terranes a magmatic arc developed in the Cordillera Real during Jurassic and Cretaceous times. The Celica Arc was developed during Late Cretaceous in central and southern Ecuador on the continental margin (Lebrat *et al.*, 1985). The metamorphic facies of the Celica Volcanics indicate its development in a highly attenuated crust during an extensional regime (Aguirre, 1992). Paleocene compression generated a thickened crust and the marginal basin inversion (Jaillard *et al.*, 1995), and since that time, the development of normal Andean-type volcanism along the margin.

During the Eocene, the arc developed on the accreted oceanic terranes, but after important late Eocene deformation (Jaillard *et al.*, 1995), the arc migrated to the Interandean Valley in Miocene times. The opening of the Guayaquil Gulf during Miocene times due to important strike-slip in the Pujili-Cauca and Peltetec-Romeral fault systems, generated a series of complex pull-apart basins in the coastal areas, and intermontane basins in the Andean areas (Jaillard *et al.*, 1999). Important deformation and basin

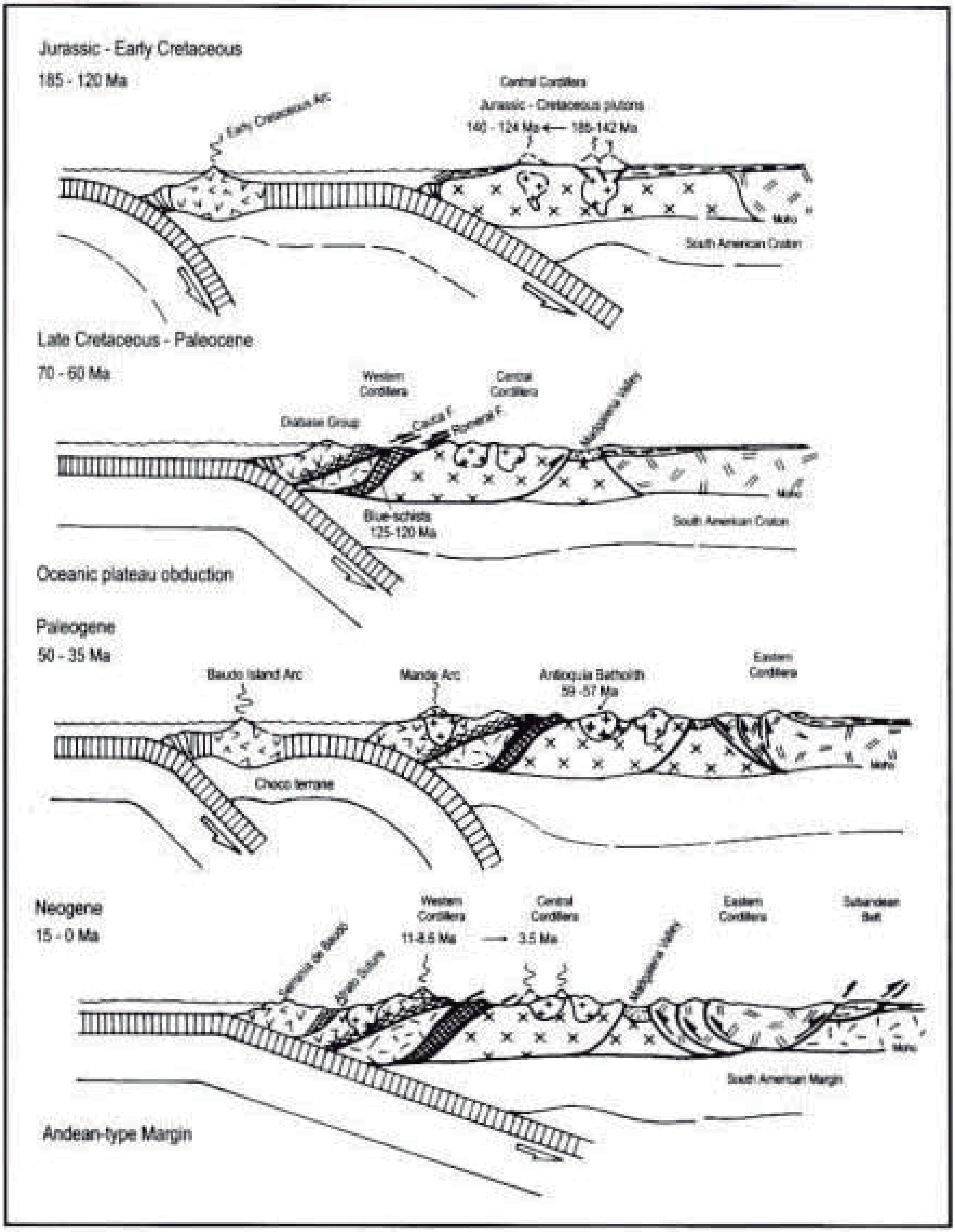


FIGURE 12: Schematic evolution of western Colombia showing the accretion of different terranes. Ophiolites related to the Romeral Fault represent the suture of the oceanic terranes with the Early Mesozoic continental margin. The obduction of an oceanic plateau emplaced the Diabase Group of the Dagua Terrane (modified after Barrero, 1979; Aspden and McCourt, 1986; Bourgois et al., 1987; Colletta et al., 1990; Kerr et al., 1997a,b).

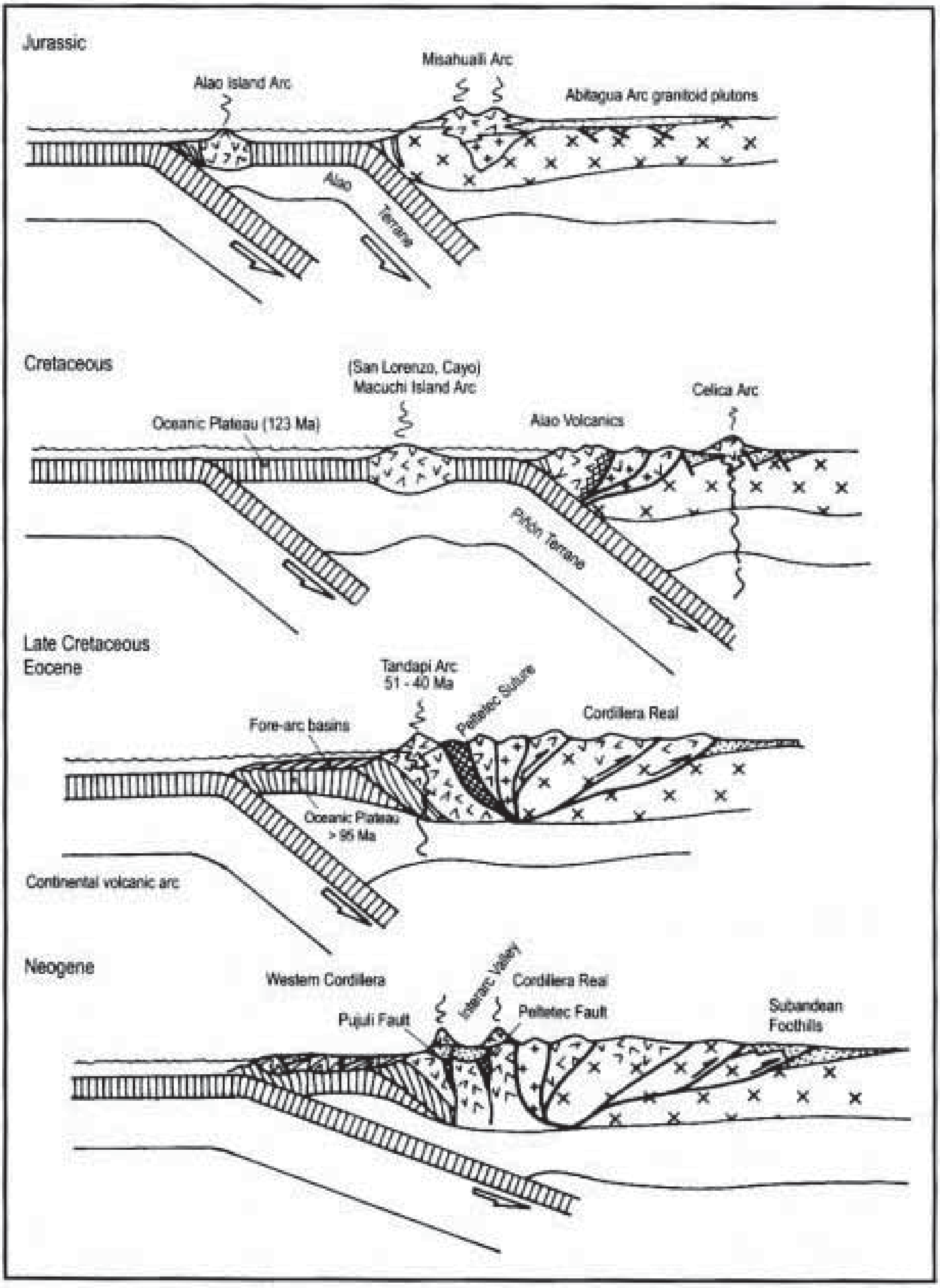


FIGURE 13: Generalized tectonic evolution of the accreted terranes of western Ecuador (modified after Aspden and Litherland, 1992; Van Thournout et al., 1992; Aguirre, 1992; Cosma et al., 1998; Reynaud et al., 1999).

inversion took place in middle to late Miocene times related to the onset of the Carnegie Aseismic Ridge (Daly, 1989). As a consequence of this, an uplift-rate of 0.7 mm/y occurred in the last 9 Ma, accounting for more than 6 km of uplift (Steinmann *et al.*, 1999).

Intra-arc and arc magmatism

The Andes between of the Guayaquil (3°S) and Penas (46°30'S) gulfs have a long history of Andean-type subduction, without accretion of oceanic terranes. This segment of the Andes known as the Central Andes after Gansser (1973) has a series of varying tectonic settings through time, prior to the present Andean-type subduction. The Early Mesozoic was a time of extension along most of the Andean margin. The first stages were related to active rifting during the Pangea break-up, and soon after that, in the latest Triassic or Early Jurassic times, to a peculiar type of subduction that developed a poorly evolved magmatism with intra and retro-arc extension.

This Early Mesozoic subduction was related to a negative trench rollback velocity as described by Uyeda (1983) and Daly (1989), as a consequence of the relative motions between the Farallon and South America plates. The change in true polar wandering path of South America at about 115 Ma, as calculated from the hot-spot reference frame (Somoza, 1995), modified the trench rollback to a positive velocity. As a result of that the South America Plate overrode the previous trench line beginning an important period of compression (Ramos, 1989a).

The timing and intensity of the extension was mainly controlled by the obliqueness of the convergence vector with the continental margin trend as inferred in northern Chile by Scheuber *et al.* (1994).

Northern Central Andes

This changing Early Mesozoic scenario has been recognized in central Peru (Fig. 14), where a marginal basin formed along the continental margin during Early Cretaceous subduction (Atherton *et al.*, 1983). This marginal basin, here interpreted as an intra-arc basin, was responsible of the Copara-Casma volcanism developed during Aptian-Albian times (Soler and Bonhomme, 1990). The basin was filled by several kilometres of pillow lava, sheet lava, hyaloclastite sediments, tuff and minor chert, siliceous and calcareous ooze. There are tholeiitic basalt flows and andesite along the axis with more acid high-K rocks toward the E (Atherton and Webb, 1989). Geochemical affinities indicate a destructive margin, with some rocks showing some MORB-type tendencies. This is in agreement with basin deepening to the W, and a general trend of deeper facies to the N, where the basin continues into the Celica Arc Basin (Jaillard *et al.*, 1999). The Casma Group Basin has been considered an aborted ensialic basin developed on highly attenuated continental crust (Aguirre and Offler, 1985) compatible with an extensional intra-arc setting (*sensu* Dickinson, 1974).

Localized Middle Albian Mochica compressional deformation was coeval with early batholith emplacement (± 105 Ma; Soler and Bonhomme, 1990). Softening of the lower crust and weakening of the lithosphere along the continental margin during the extension of the Casma Group volcanism favoured the compressional deformation. This phase was also associated with an increase in convergence rates, and was responsible of basin closure, compressional deformation, tectonic inversion and emplacement of the Coastal Batholith from Late Albian to Paleocene times (102 to 59 Ma). Several episodes of emplacement have been recognized, as well as a variety of facies from typical I-type medium-K to high-K granitoid rocks, described in detail by Pitcher *et al.* (1985). Oblique convergence was dominant during Coastal Batholith emplacement, with a low orthogonal convergence rate (Soler and Bonhomme, 1990). Compressional deformation of the Peruvian phase propagated eastwards during the Late Cretaceous, changing the volcanism from generally subaqueous to subaerial. The Late Cretaceous deformation was also well developed in the Western Cordillera of southern Peru, where the allochthonous thrust sheet of the Arequipa-Paracas Arc overrode the Early Mesozoic platform facies (Vicente, 1990).

Subsequent middle to late Eocene (47-32 Ma) compression, produced the fold and thrust belt of the Western Cordillera, E of the Cordillera Blanca Lineament (Benavides-Cáceres, 1999). A series of Paleogene pulses between the middle Eocene and the early Oligocene are known as the Incaic Phase, which produced the main orogenic uplift of the Peruvian Andes (Vicente *et al.*, 1979). This coincides with a period of rapid orthogonal convergence rates, and expansion of the magmatism as shown by Pilger (1984) and Soler and Bonhomme (1990). The expansion and migration of the magmatism was related to a shallowing of the geometry of the Wadati-Benioff Zone at that time (Fig. 15).

Diachronic Miocene compression, generally attributed to some of the Quechua phases, uplifted and deformed the Eastern Cordillera and the Subandean foothills. The emplacement of the Cordillera Blanca Batholith (13-3 Ma) in the highest sector of Western Cordillera was controlled by an important crustal discontinuity (Atherton and Sanderson, 1987). The Peruvian flat-slab segment was developed in Pliocene times, and a magmatic lull extended from the Ecuador boundary to the latitude of Arequipa, with the only exception of some minor Pliocene peralkaline stocks in the Subandean Basin (Sébrier and Soler, 1991).

Southern Peru was characterized by some sort of delamination as proposed by Sandeman *et al.* (1995), after the Paleogene shallowing of the subduction zone, associated with the Incaic deformation. A thermal effect with no associated magmatism in Paleogene times was recognized along 450 km of the Eastern Cordillera parallel to the Altiplano (Farrar *et al.*, 1988). This area, known as the Zongo-San Gabán Zone, has produced a thermal event detected by Ar^{39}/Ar^{40} at 38 Ma, associated with uplift, erosion and southwestward vergent thrusting on the western flank of the Eastern Cordillera. This deformation was prior to the northeastward vergent thrusting along the eastern foothills. Sandeman *et al.* (1995) favoured the formation of a slab-window linked to the detachment of the subducting plate. An alternative model could be the transition from flat to normal subduction that resulted in interaction of hot

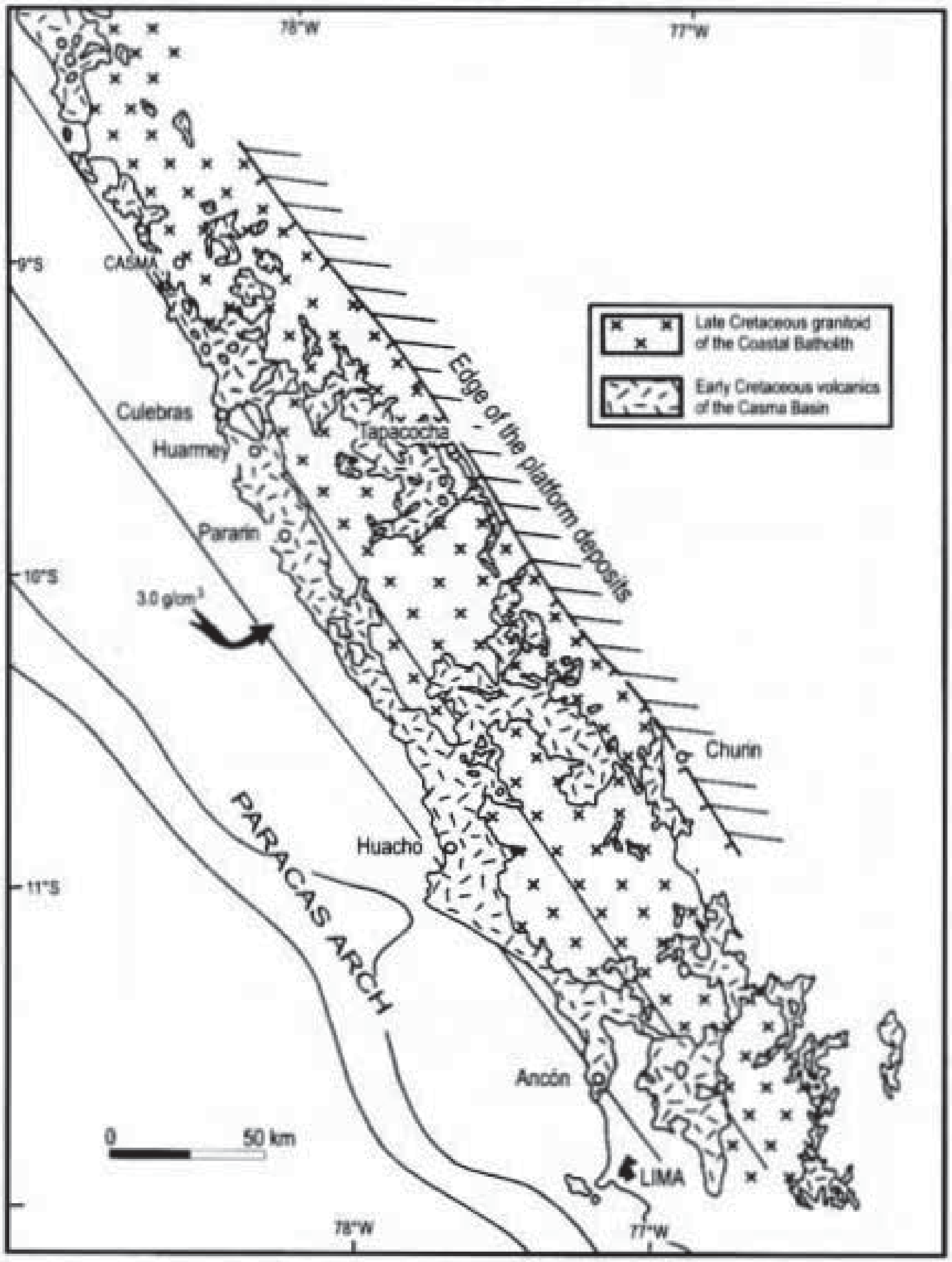


FIGURE 14: Surface exposures of the Casma Group volcanics, and offshore possible extension E of the 3.0 g/cm³. Basin is bounded by the Paraca Arch, possible offshore extension of the Arequipa Massif. The Coastal Batholith was emplaced along the Casma Group Basin (modified after Atherton and Webb, 1989; Soler and Bonhomme, 1990; Benavides-Cáceres, 1999).

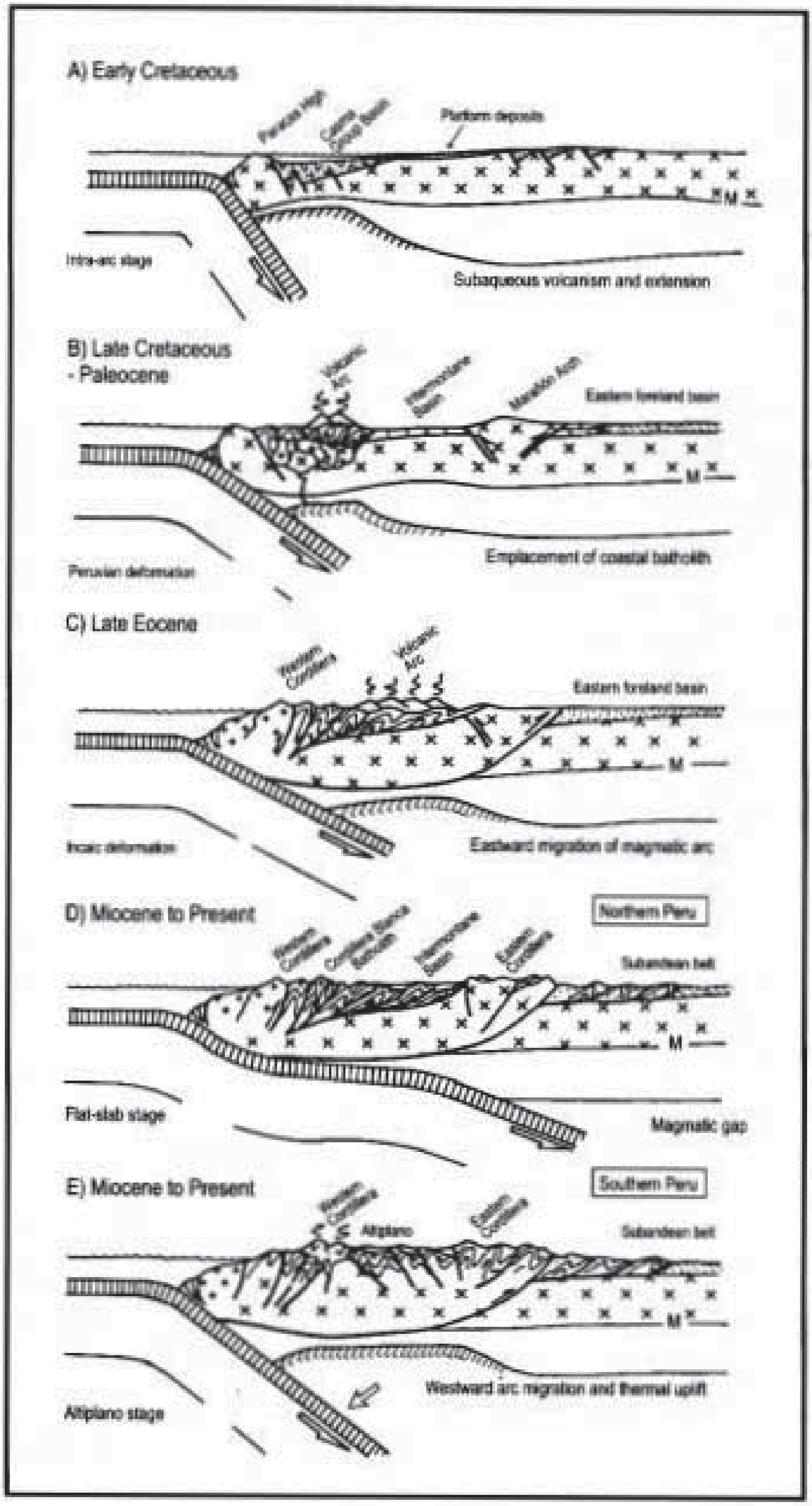


FIGURE 15: a) to c) Tectonic evolution of the Peruvian margin from an intra-arc setting and generalized rifting in the Early Cretaceous to a compressive regime in the Cenozoic; d) Late Cenozoic in Northern and Central Peru; e) Late Cenozoic in Southern Peru (modified after Soler and Bonhomme, 1990; Sébrier and Soler, 1991; Sandeman et al., 1995).

asthenosphere and hydrated mantle. This would cause mantle melts and hot fluids that may explain the thermal effect as being proposed farther S by Kay *et al.* (1999).

Late Cenozoic in the southern Peru was characterized by important arc magmatism along the Western Cordillera, with some shoshonitic volcanic rocks occurring in the Eastern Cordillera. The northern flank of the Altiplano was affected by important normal faulting in late Cenozoic times. This normal faulting was explained by the high topography produced after uplift of the Andean Cordillera and the Altiplano, and the reorganization of the stress that resulted in N-S rending extension (Sebriér *et al.*, 1985).

Central Andes

Early Mesozoic evolution of this segment differs from the previous segment. Although there was a generalized extension, during Jurassic and Early Cretaceous times between 21°S and 27°S, a deep retro-arc basin developed behind the arc (Mpodozis and Ramos, 1990). Volcanic activity was concentrated in the La Negra Arc, a poorly evolved tholeiitic series of basalt flows, basandesite, and mafic dykes, developed along the continental margin during an extensional regime. The western part of La Negra Arc was tectonically eroded (Rutland, 1971), due to the high relief horst-and-graben structure of the outer slope of the oceanic subducting plate at these latitudes (Von Huene *et al.*, 1999).

The highly oblique convergence during Early Mesozoic times (Scheuber *et al.*, 1994), was also responsible for the development of the Atacama strike-slip system (Fig. 16). This trench-parallel fault system, which affected the Coastal Cordillera, controlled the ascent and emplacement of the arc granitoid of the Coastal Batholith during the Jurassic and earliest Cretaceous (Naranjo *et al.*, 1984). The negative rollback velocity of the trench expanded the previous Triassic normal faults, until about 132-125 Ma, when a left-lateral strike-slip system generated important transtension and enhanced the emplacement of granitoid plutons (Grocott *et al.*, 1994).

The transtensional regime ended at about 106 Ma, when the Coastal Cordillera Arc was abandoned and the arc migrated toward the Central Valley (Mpodozis and Ramos, 1990). This 50 km shifting of the magmatic arc coincides with major compressional reactivation of the strike-slip system, where a series of compressional duplexes deformed the Coastal Cordillera granitoid rocks. Rotation associated with this strike-slip displacements was confirmed by paleomagnetic studies (Taylor *et al.*, 1998).

Late Cretaceous compressional deformation, approximately coeval with other Cretaceous phases described in the Peruvian segment, thrust and uplifted the proto-Cordillera de Domeyko (Mpodozis and Ramos, 1990), and generated the Purilactis-Salar de Atacama foreland basin (Muñoz *et al.*, 1997).

The early Paleogene was characterized by an intense explosive volcanism (Fig. 17), with huge calderas and associated intrusives, developed in an extensional regime (Cornejo *et al.*, 1994). During late Eocene and early Oligocene times a lull in arc activity, was associated with the Domeyko Fault System, another strike-slip system developed parallel to the Atacama, but further to the E in the present Cordillera de Domeyko (Tomlinson and Blanco, 1997). Magmatic activity reduced to some acid porphyries was emplaced along the Domeyko Fault System, where giant porphyry copper ore deposits are found (Zentili and Maksaev, 1995).

Deformation and magmatism shifted to the Altiplano (N of 22°S) in Bolivia and to the Puna (S of 22°S) in Argentina, initially linked to the tectonic erosion of the fore arc (Stern, 1991), and subsequently associated with shallowing of the subduction zone (Kay *et al.*, 1999).

Paleogene magmatic rocks are scarce in the Altiplano, where a few volcanic fields (*c.* 34-23 Ma) are known with a retro-arc setting (Avila Salinas, 1991). Rocks of this age are rare in the Argentine Puna, and are associated with acid porphyries such as Taca Taca, in the western sector.

Most of the Altiplano and Puna was affected by active extensional faulting until Albian times with the development of the Salta Rift, and equivalent systems in the Altiplano, as well as in the Subandean region (Sanjines-Saucedo, 1982; Salfity, 1982, 1994). Thermal subsidence in these basins persisted until early Paleogene times, although some authors interpreted part of this Paleogene subsidence as a distal response of the flexural loading produced by the Peruvian phase farther W (Sempere, 1995).

Shifting of the deformation to the Altiplano and Puna produced during late Oligocene-early Miocene (Sempere *et al.*, 1990) the tectonic inversion of the Cretaceous N-S trending grabens, and the uplift of the Eastern Cordillera. As a result, the Neogene flexural loading produced two distinctive settings, the Altiplano intermontane basins and the foreland Subandean basins. This tectonic subsidence was enhanced in the Altiplano by western vergent-thrust systems in Eastern Cordillera (Hérail *et al.*, 1993). Calc-alkaline volcanic rocks expanded between 17 to 12 Ma to the E in the Altiplano and Puna, associated with a shallowing of the subduction zone (Kay *et al.*, 1999).

After an important tectonic shortening in late Miocene, where deformation produced important uplift of the Altiplano-Puna and Eastern Cordillera, deformation shifted to the Subandean System, where thin-skinned thrust belt starts developing (Baby *et al.*, 1995).

By the late Miocene to Pliocene, voluminous ignimbrite sheets, with ages ranging from 11 to 3 Ma, were erupted from huge calderas. Centers with ages from 11 to 6.5 Ma are found across the Altiplano-Puna and into the Eastern Cordillera, whereas younger centers are restricted to Western Cordillera and the western flank of the Altiplano (Kay *et al.*, 1999). This change in the character of the volcanism, and the migration toward the trench was interpreted as evidence of steepening of the subduction zone, that concentrated during late Cenozoic times the magmatic activity in Western Cordillera (Coira *et al.*, 1993). The different geological processes associated with the steepening of a subduction zone have been analyzed by James and Sacks (1999), who noted the important influx of asthenosphere material from depth into the growing mantle wedge. As a result of this there occurred widespread melting beneath the older hydrated arc.

Tectonic evolution of the region continued through foreland propagating thrusts in the foothills of the Subandean System and adjacent Chaco plains, and out-of-sequence thrust reactivation in the Altiplano-Puna and Eastern Cordillera.

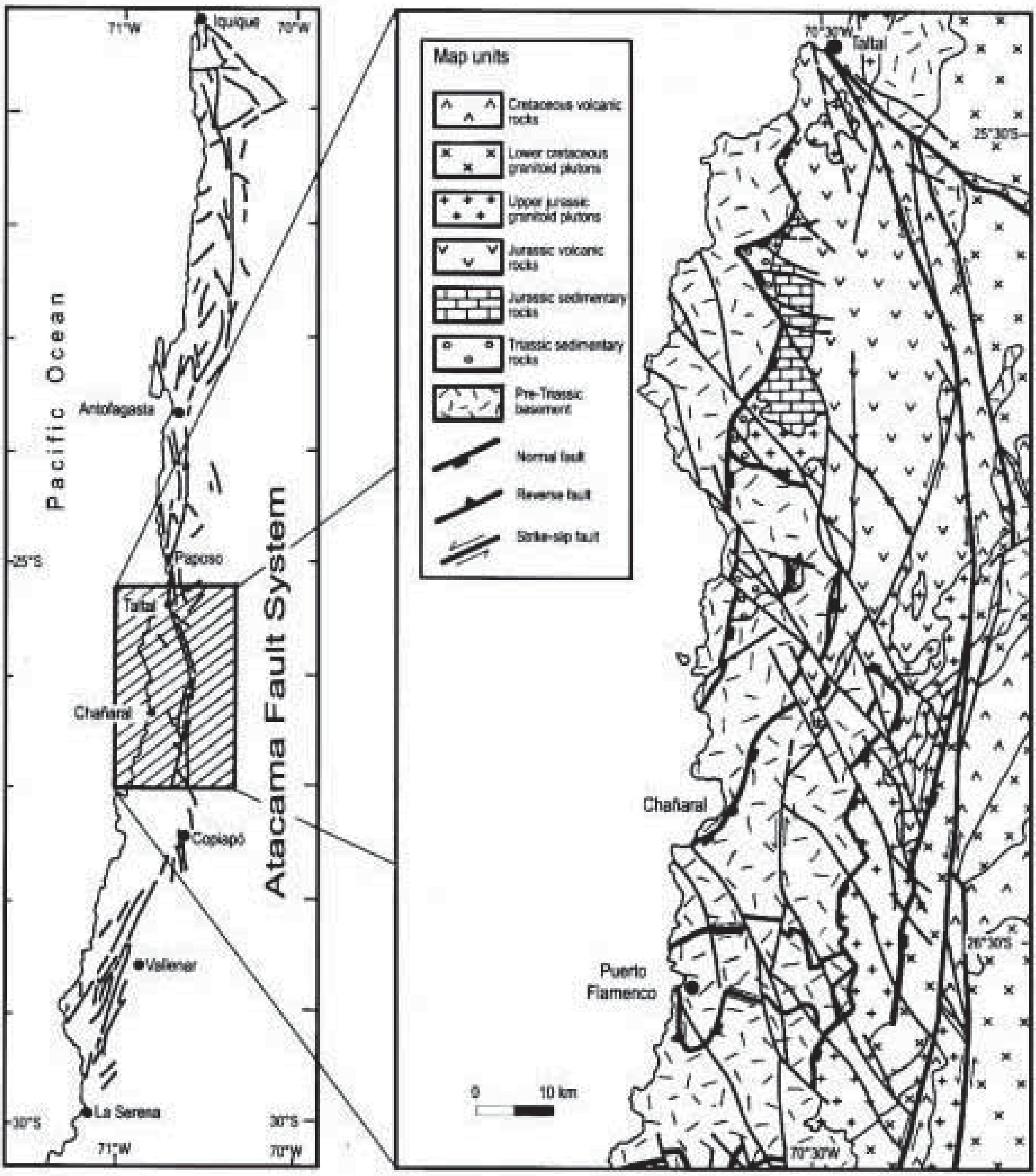


FIGURE 16: a) The Atacama strike-slip fault system. b) Detailed structural scheme between Taltal and Puerto Flamenco with location of early normal faults, and subsequent strike-slip faults with the different granitoid plutons emplaced during left lateral transtension (modified after Naranjo et al., 1984; Taylor et al., 1998).

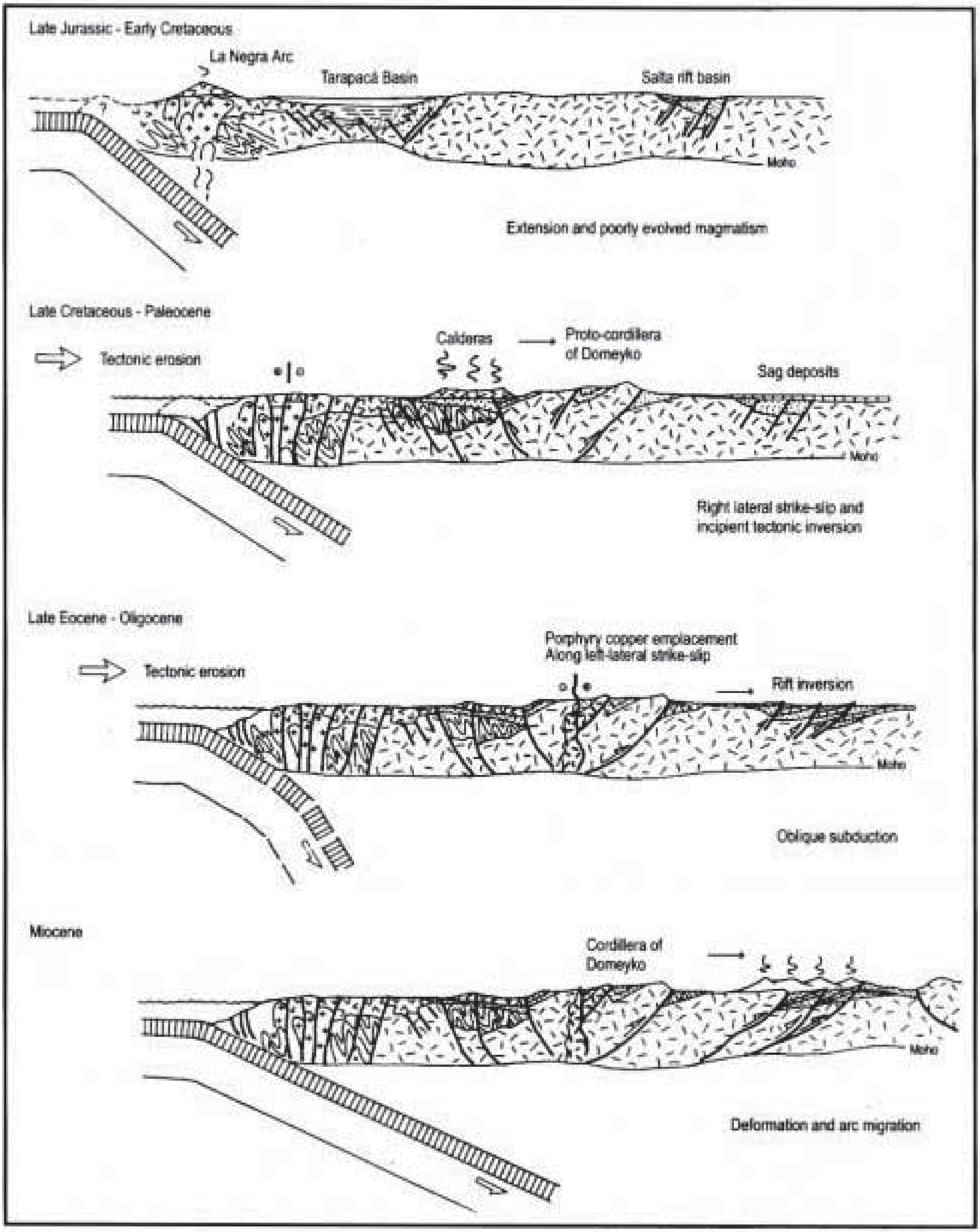


FIGURE 17: Tectonic evolution of the Central Andes since Mesozoic times (modified after Mpodozis and Ramos, 1990).

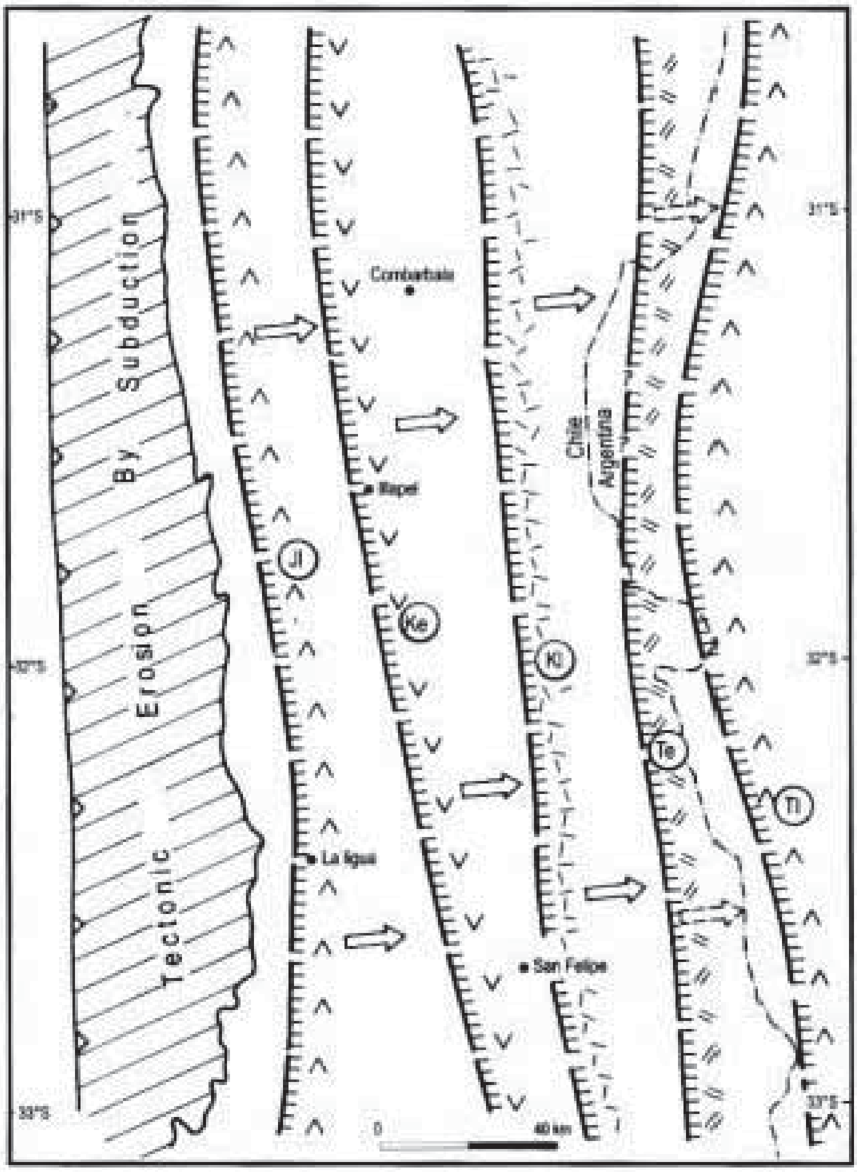


FIGURE 18: Eastward migration of the volcanic front related to tectonic erosion of the continental margin up to Paleogene times (modified after Ramos, 1988b). Miocene volcanic front migration and magmatic arc expansion were controlled by changes in the Wadati-Benioff geometry (modified after Jordan et al., 1983).

As a result, one of the most complete section of the Andes is observed in the Central Andes of Bolivia, with an active orogenic front propagated as far as 800 km from the trench.

Southern Central Andes

The central region of Chile and Argentina underwent an important extension during early Mesozoic times. This generalized extension was interpreted as the result of continental spreading of an ensialic aborted marginal basin where several kilometres of Middle Cretaceous volcanic rocks bearing an important burial metamorphism were deposited (Aguirre *et al.*, 1989). This tectonic scenario has also been explained as an intra-arc setting by Charrier (1984) and Ramos (1985), where an inner and outer arc with different petrographic characteristics were developed.

The tectonic setting of this central Chile basin is similar to the Casma Group basin described in the Peruvian segment, although the sedimentary facies associated with the intra-arc basin do not record deep or basinal facies as in the Peruvian Basin (Atherton and Webb, 1989). Most of the carbonate platform and associated clastic sediments indicate shallow water deposition, and interfingering with subaerial volcanism (Charrier, 1984). These intra-arc basins were coeval with the development of retro-arc basins in the eastern side of the orogen. Several cycles of carbonate and clastic sediments were deposited, controlled by tectono-eustatic sea-level changes (Uliana and Legarreta, 1993; Legarreta and Uliana, 1996).

Extension was driven by low-angle normal faults associated with subhorizontal detachments in the Copiapó region of central Chile (27°S, Mpodozis and Allmendinger, 1993). The retro-arc basins on the eastern side of the cordillera show older extensional structures that are characterized by high-angle faults.

The plate-kinematic changes observed along the entire South American Pacific margin are also noticeable in central Chile. By the end of the Early Cretaceous, a major plate reorganization took place, and as a response the extensional regime of the marine intra-arc and retro-arc basins ended (Mpodozis and Ramos, 1990). The new stress field was responsible of the sea withdrawal, and the inception of continental retro-arc basins. The volcanic front migrated to the E (Figs. 18 and 19) and a series of volcanic and

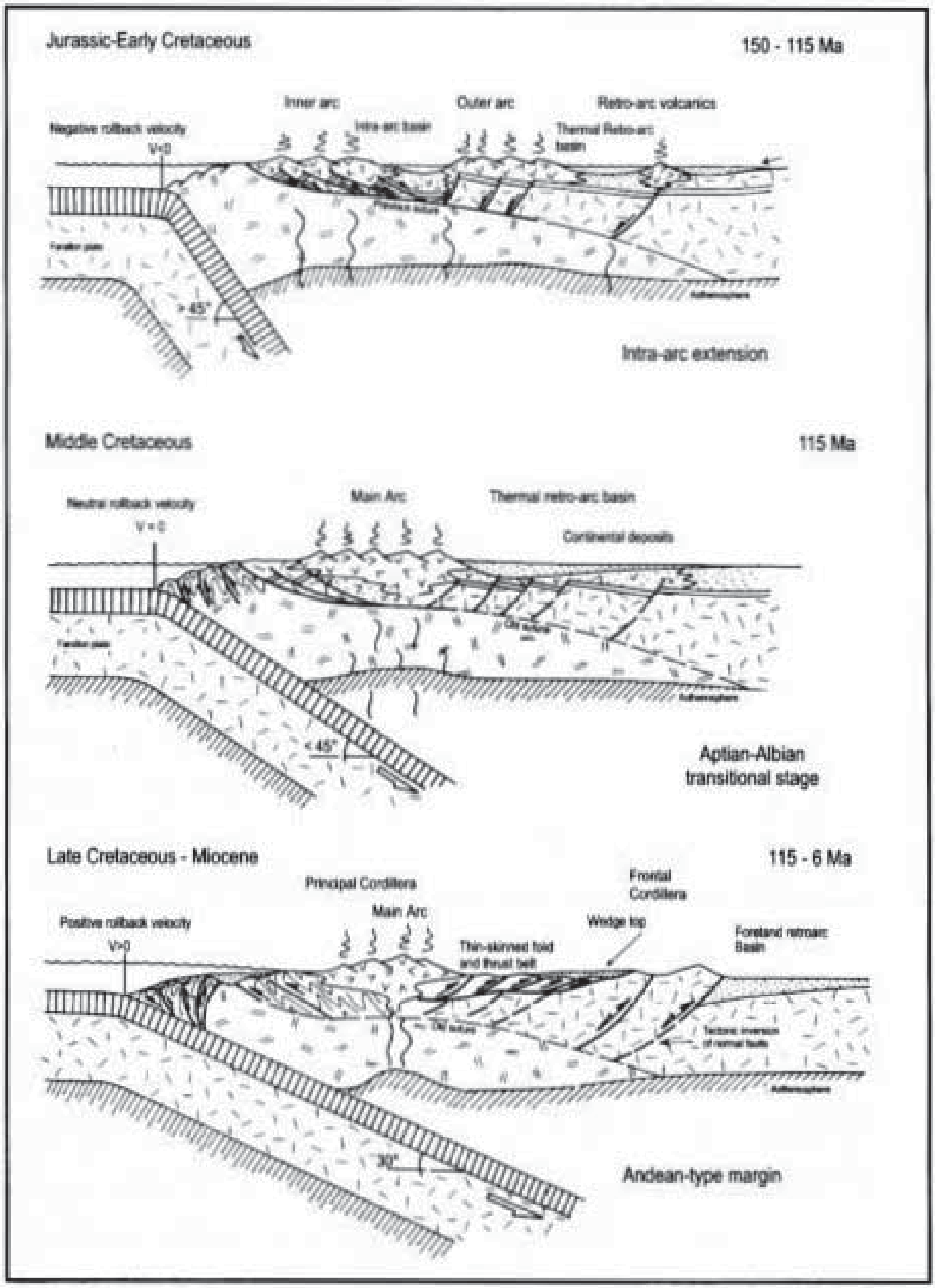


FIGURE 19: Different tectonic regimes in the southern Central Andes, showing the change between an extensional and compressional tectonic regime at about 115 Ma (modified after Ramos, 1988b; Mpodozis and Ramos, 1990; Somoza, 1995).

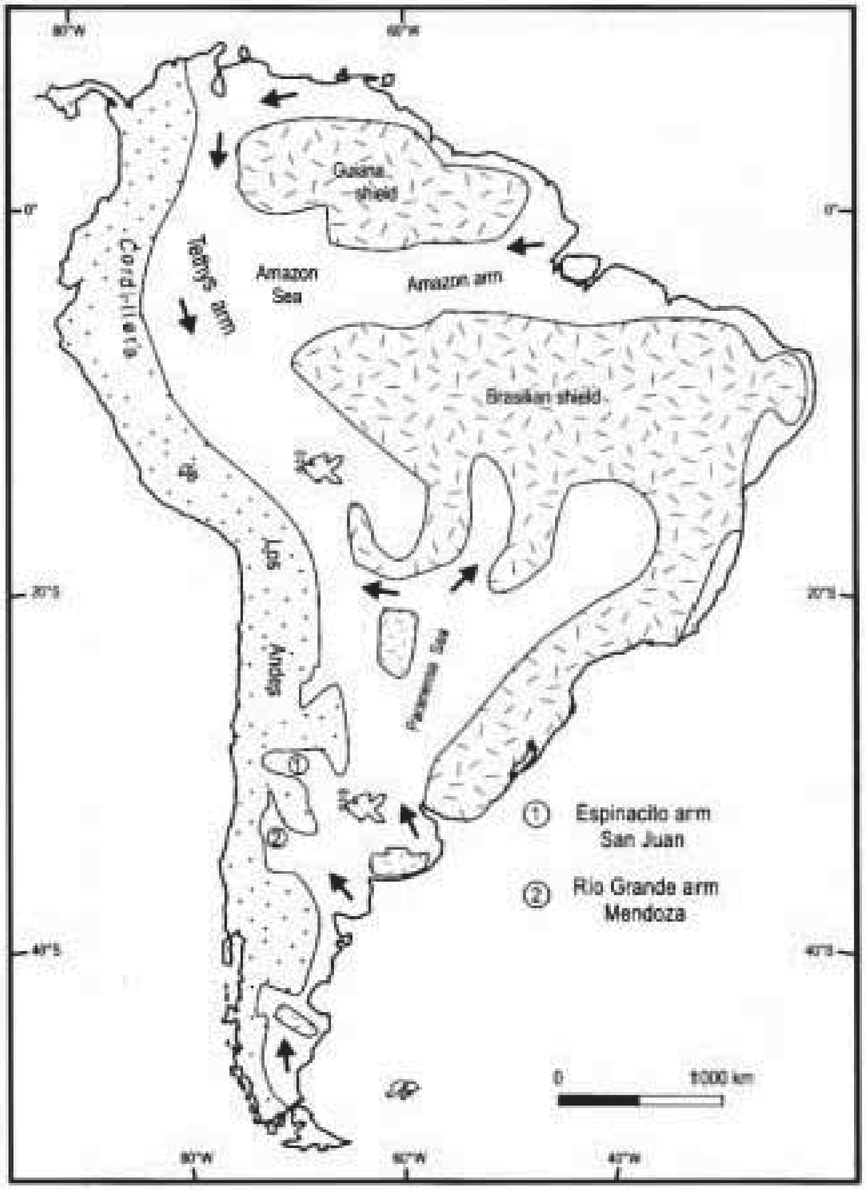


FIGURE 20: Miocene marine transgression into the Andean foothills as a result of rapid tectonic loading of the Andes. Records of these marine deposits are known in central and northern Argentina, subandean region of Bolivia and Peru, and the foothills of Eastern Cordillera of Ecuador and Colombia (modified after Ramos and Alonso, 1995; Pérez et al., 1996).

volcaniclastic rocks were accumulated in a set of eastwards-migrating depocenters. This steady migration since Jurassic times was attributed to tectonic erosion of the continental margin (Ramos, 1988b; Stern, 1991).

A rapid migration of the volcanic front by early to middle Miocene times was related to shallowing of the subduction zone (Jordan *et al.*, 1983). The eastwards shifting of the volcanism was also related to deformation that started *c.* 22 to 20 Ma in the westernmost Chilean sector and prograded at *c.* 16 - 15 Ma on the Argentine slope of the cordillera.

Maximum flexural subsidence between 15 and 13 Ma was responsible of the marine transgression of the Paranense Sea, which diachronically invaded most of the Andean foothills between these latitudes and N of Colombia up to the Maracaibo area (Fig. 20). Testimonies of this Miocene transgression are preserved as isolated patches at 3200 m a.s.l. in several localities of the Principal Cordillera of central Argentina and in the Puna foothills of northern Argentina, where middle late Miocene foraminifera have been found and the age was also constrained by radiometric dating and magnetostratigraphy (Ramos and Alonso, 1995; Pérez *et al.*, 1996). These marine deposits continue farther N with the middle to late Miocene green shale beds of Santa Cruz de la Sierra in Bolivia where foraminifera have been found (Yecua Formation of Marshall *et al.*, 1993). These marine deposits have been recognized in the foreland basin of Peru (15 Ma Pebas Formation, Hoorn, 1993); in southern Ecuador, with possible Pacific connections through the Guayaquil Seaway in the middle Miocene (Steinmann *et al.*, 1999); and the middle Miocene foreland basin of Colombia (16 to 10.5 Ma León Formation in Cooper *et al.*, 1995; Villamil, 1999). The late middle Miocene sporadic marine transgression in this highly subsiding foreland basin can be traced up to the Maracaibo area (Hoorn *et al.*, 1995). This marine transgression of the Paranense Sea could be correlated with the Amazon Sea of Räsänen *et al.* (1995), if the peak of flooding is correlated with the middle Miocene highstand of 12 Ma (Haq *et al.*, 1987), as discussed by Webb (1995). This connection can explain the Caribbean affinities of the middle to late Miocene foraminifera fauna described by Boltovskoy (1991) in the Argentine offshore platform, which was noticeable absent on the Brazilian platform at that time.

The Andes of central Chile also record at these latitudes an important Miocene deformation favoured by the thermal weakening of the continental crust, that inverted the previous Oligocene intra-arc basin (Godoy *et al.*, 1999).

The shallowing of the subduction zone in the Pampean flat-slab, ends the last stages of volcanism at about 7 - 4 Ma in the Principal Cordillera, and resulted in subduction related magmatism as far as 750 km away from the trench at 1.9 Ma in the eastern Sierras Pampeanas (Ramos *et al.*, 1991; Kay *et al.*, 1991). The migration of the magmatism was accompanied by diachronic deformation and uplift: First, the Frontal Cordillera as a single mountain block (9 - 8 Ma); and subsequently, the thin-skinned Precordillera fold and thrust belt and the basement blocks of Sierras Pampeanas, during Pliocene to Quaternary times (Ramos *et al.*, 1996a, b). Present tectonic activity is shown in the highly seismic area of the Precordillera thrust front and in the western Sierras Pampeanas.

The Neogene was a time of important subsidence in the foothills of the central Argentina foreland basins (Jordan, 1995). During the early Miocene, it was a single extended foreland basin that was cannibalized during middle Miocene times. Subsequent deformation in the late Miocene, and out-of-sequence thrusting at about 7 Ma, originated and reactivated a series of intermontane basins such as the Uspallata and Iglesias piggy-back basins between the Principal and Frontal Cordilleras (Beer *et al.*, 1990; Cortés, 1993). The uplift of Precordillera was not coeval along strike, as demonstrated by magnetostratigraphic analyses in different sectors of the Precordillera (Jordan *et al.*, 1988, 1997). The uplift of western Sierras Pampeanas at about 3 Ma originated the Bermejo broken foreland basin (Fernández and Jordan, 1996; Zapata and Allmendinger, 1996) and the Los Colorados Basin farther N, that respectively accumulated more than 9000 m and 10 000 m of clastic deposits (Fernández and Jordan, 1996; Ramos, 1970).

The eastern Chaco plains of Bolivia and Argentina and the Pampas of central Argentina were affected by widespread dynamic subsidence responsible of few hundred metres of latest Cenozoic sedimentation as far as 1000 km of the Andes thrust front.

Southern Andes

The Patagonian Andes also show the generalized extension during early Mesozoic times with the development of foreland embayments such as the Neuquén and Río Mayo basins. Along the axis of the cordillera an intra-arc setting is observed at the latitude of Lago Fontana (45°S), where an inner and partially subaqueous arc interfingered with Neocomian and older marine deposits, in a complex graben system. The early Mesozoic extension encompassed most of the foreland region, almost to the Atlantic coast (Uliana and Biddle, 1988).

Middle Late Jurassic arc magmatism consists of andesite and dacite along the main Patagonia Cordillera, that after a pulse of lull magmatism during Neocomian times resumed in Aptian to Albian times (Ramos and Aguirre-Urreta, 1994). At this time vast volcanism of dacitic to rhyolitic composition was dominant in the extra-Andean region.

Mild inversion tectonics took place during Late Cretaceous times and partially uplifts the Patagonian Cordillera and the adjacent Precordillera. Oblique subduction in the northern sectors was responsible of the strike-slip and pull-apart basin formation in Paleogene times.

The Ñirihuao Basin between 41°S and 43°S is a retro-arc basin developed during late Eocene times, filled with continental deposits and sporadic marine transgression from the Pacific side (Ramos, 1982). This basin was reactivated as a pull-apart basin formed by transtension in Oligocene times (Dalla Salda and Franzese, 1987). The basin was subsequently inverted during Miocene times, developing a narrow fold and thrust belt (Giacosa and Heredia, 1999).

The Liquiñe-Ofqui Fault (Hervé, 1994), parallel to the Pacific margin along more than 750 km in the present fore-arc region of the Patagonian Andes, is an active fault that shows right-lateral displacement, at least since the Miocene (Cembrano *et al.*, 1996). The wrench fault trace is approximately aligned with the chain of active volcanoes from the Hudson (46°S) to the Lonquimay Volcano (39°S). Strike-slip decreases from S to N due to transfer of displacements to synthetic systems. The Cenozoic stress is partitioned in the fore arc, favoured by the thermally weakened crust of the continental margin (Cembrano *et al.*, 2000), and is not transmitted to the retro-arc (Dewey and Lamb, 1992). The decoupling of the deformation along the active arc by oblique subduction, as suggested farther N by Scheuber *et al.* (1994), left the region E of the arc with almost no deformation during Late Cenozoic times.

The situation dramatically changes S of the present Chile triple junction at about 46°30'S (Fig. 21). The effects of the ridge collision, ophiolitic emplacement, and the consequent triple junction migration show a connection between ridge-trench collision, the cessation of arc volcanism with deformation in the foreland fold and thrust belt and retro arc basaltic plateau volcanism in late Miocene to present times (Stern *et al.*, 1976; Ramos and Kay, 1992; Gorring *et al.*, 1997). Prior to the subduction of the ridge was emplaced the exceptional Cerro Pampa adakite due to partial melt of hot and young oceanic lithosphere (Kay *et al.*, 1993b). Oceanic plate reconstruction (Cande and Leslie, 1986, Tebbens *et al.*, 1997) suggests that Eocene retro-arc plateau magmatism may also be associated with ridge-trench interactions.

The geological and magmatic history of the eastern foothills at these latitudes shows a striking coincidence between shortening and uplift of the southern Patagonian Andes (Fig. 22), the arc gap and the distribution and chemistry of the volcanic arc rocks to the N and S, and the timing and chemistry of alkali plateau basaltic activity (Gorring *et al.*, 1997).

The southern Patagonian Andes S of 46°30'S due to these ridge collisions, have a 2 km higher topography, an important orogenic shortening generated by a fold and thrust belt and extensive foreland basin subsidence as seen in the Austral or Magallanes Basin (Ramos, 1989b; Biddle *et al.*, 1986). This sector is also characterized by the late orogenic emplacement of Miocene stocks of granitic composition such as San Valentín, San Lorenzo, Fitz Roy and Torres del Paine (Ramos *et al.*, 1982).

The distribution and timing of Eocene arc and plateau magmatism also appears to correlate with a ridge-trench

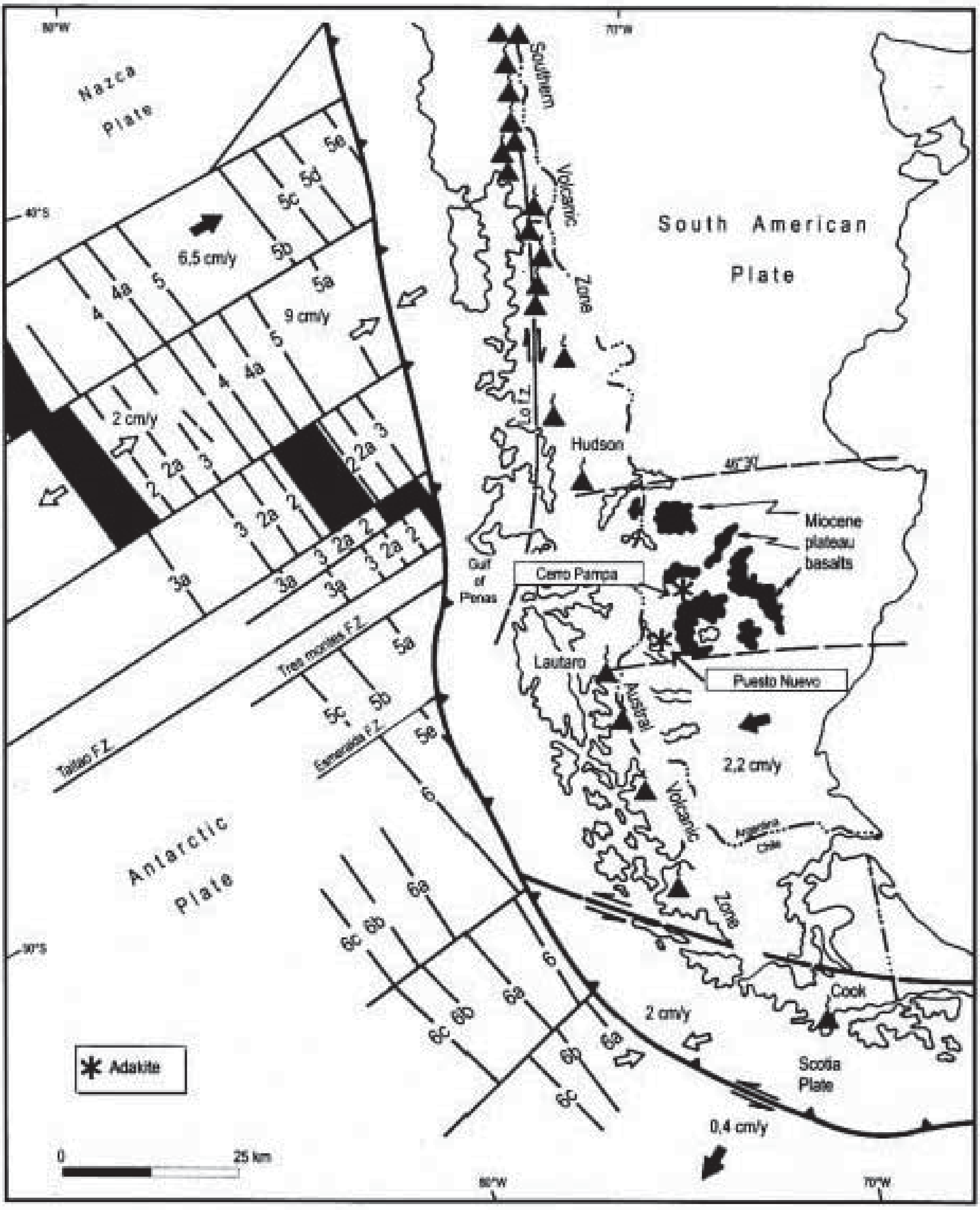


FIGURE 21: Plate tectonic framework of the Patagonian Andes N and S of the triple junction with location of the Southern and Austral volcanic zones (modified after Ramos and Kay, 1992; Gorring et al., 1997). LO F.Z.: Liquiñe-Ofqui Fault Zone.

collision (Fig. 23). In this case the ridge that separated the Farallon and Aluk plates collided at a much lower angle to the coast. The effects of the ridge collision appear only S of 43°S where extensive alkali plateau basalt was developed and the volcanic arc was absent (Ramos and Kay, 1992). Comparison of the magmatic rocks erupted in the retro-arc behind three different collided Chile ridge segments suggests that the volume of magmatism and the percentage of melting increases with the length of the ridge segment and the size of the slab window (Gorring *et al.*, 1997).

Back-arc basin formation and tectonic inversion in the Fueguian Andes

The generalized Early Mesozoic extension observed through all the Andean chain was maximum in the Fueguian Andes where the Rocas Verdes marginal basin was formed (Fig. 24) during latest Jurassic-Early Cretaceous times (Dalziel *et al.*, 1974). To the S of 52°S, along the Patagonian channels, the ophiolite rocks of the Cordillera de Sarmiento are exposed. The Sarmiento Ophiolite as well as the Tortuga Complex in the Fueguian Archipielago indicate the longitudinal development of oceanic crust along the axis of the back-arc basin (Stern *et al.*, 1976b; Dalziel, 1981).

This extension was also detected in recent seismic lines as basement-level normal faults (Klepeis and Austin, 1997). The regional extension produced the attenuated crust and the rhyolitic rocks and subordinated clastics of middle to Late Jurassic age that filled the half-graben system across the extra-Andean region (Suárez, 1976; Uliana and Biddle, 1988). Thermal subsidence produced the extensive Springhill clastic platform that expanded during the early Cretaceous throughout the entire extra-Andean region.

Closure of the marginal basin in middle Cretaceous times was followed by intense deformation and thrusting with northerly vergence (Mpodozis and Ramos, 1990).

Rocks containing upper amphibolite facies assemblages reflecting Mesozoic-Cenozoic metamorphism are exposed in the Cordillera Darwin (approximately, 55°S). The exhumation of these unique high-pressure rocks has been interpreted as having been produced by extension in a metamorphic core complex (Dalziel and Brown, 1989) after the important crustal thickening of the Late Cretaceous; low-angle extensional crustal shearing bounding the high grade metamorphic complex along the Beagle Channel, would be responsible of the tectonic denudation in latest Cretaceous-early Paleocene times. This hypothesis was challenged by Klepeis (1994) who interpreted the Cordillera Darwin as a 8 kbar basement uplift produced by thick-skinned thrusting that followed the thin-skinned deformation of the Magallanes Basin (Fig. 25). This late thrusting was produced out-of-sequence between 65-40 Ma, compatible with cooling ages and deformation history of Cordillera Darwin (Nelson, 1982; Kohn *et al.*, 1995). More recent seismic data permit the calibration of the final uplift of Cordillera Darwin during the shift away from compression towards transtension during late Tertiary crustal relaxation (Klepeis and Austin, 1997).

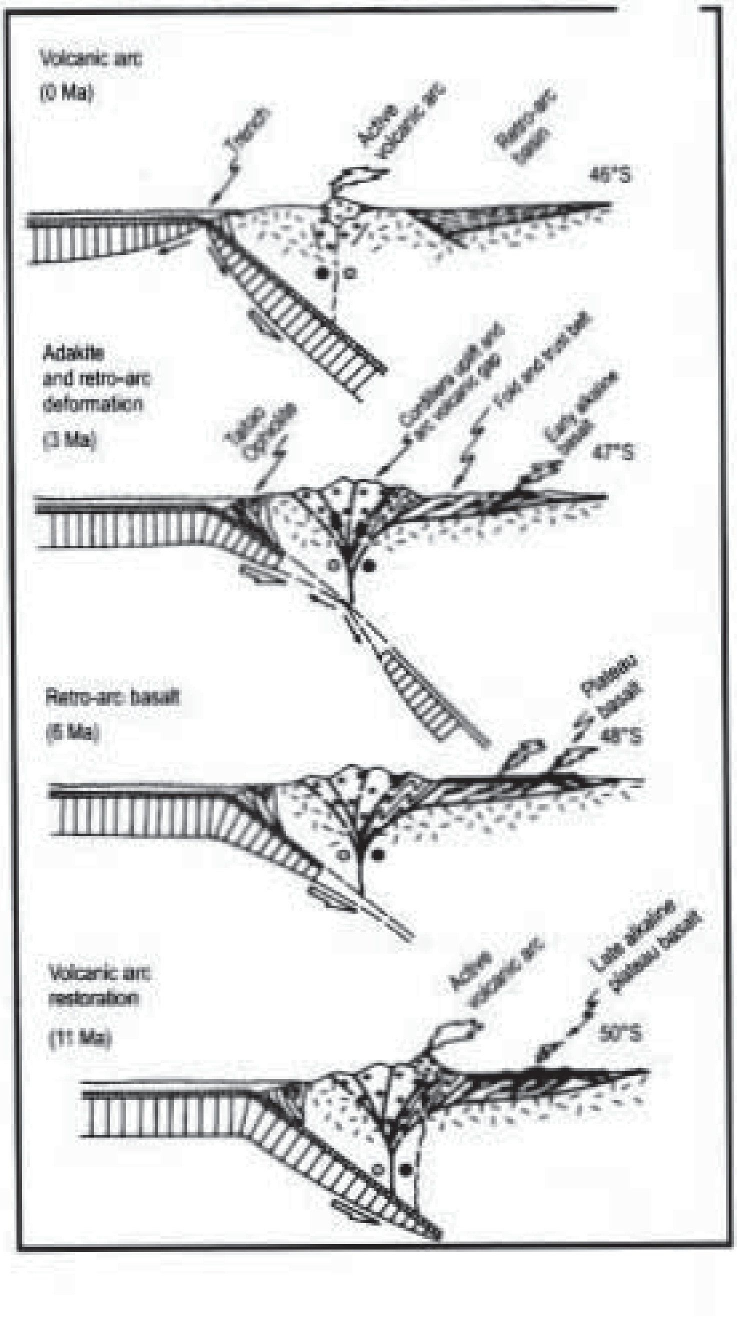


FIGURE 22: Tectonic evolution of the southern Patagonian Andes during ridge collision of a seismic ridge (modified after Stern et al., 1976a; Ramos and Kay, 1992).

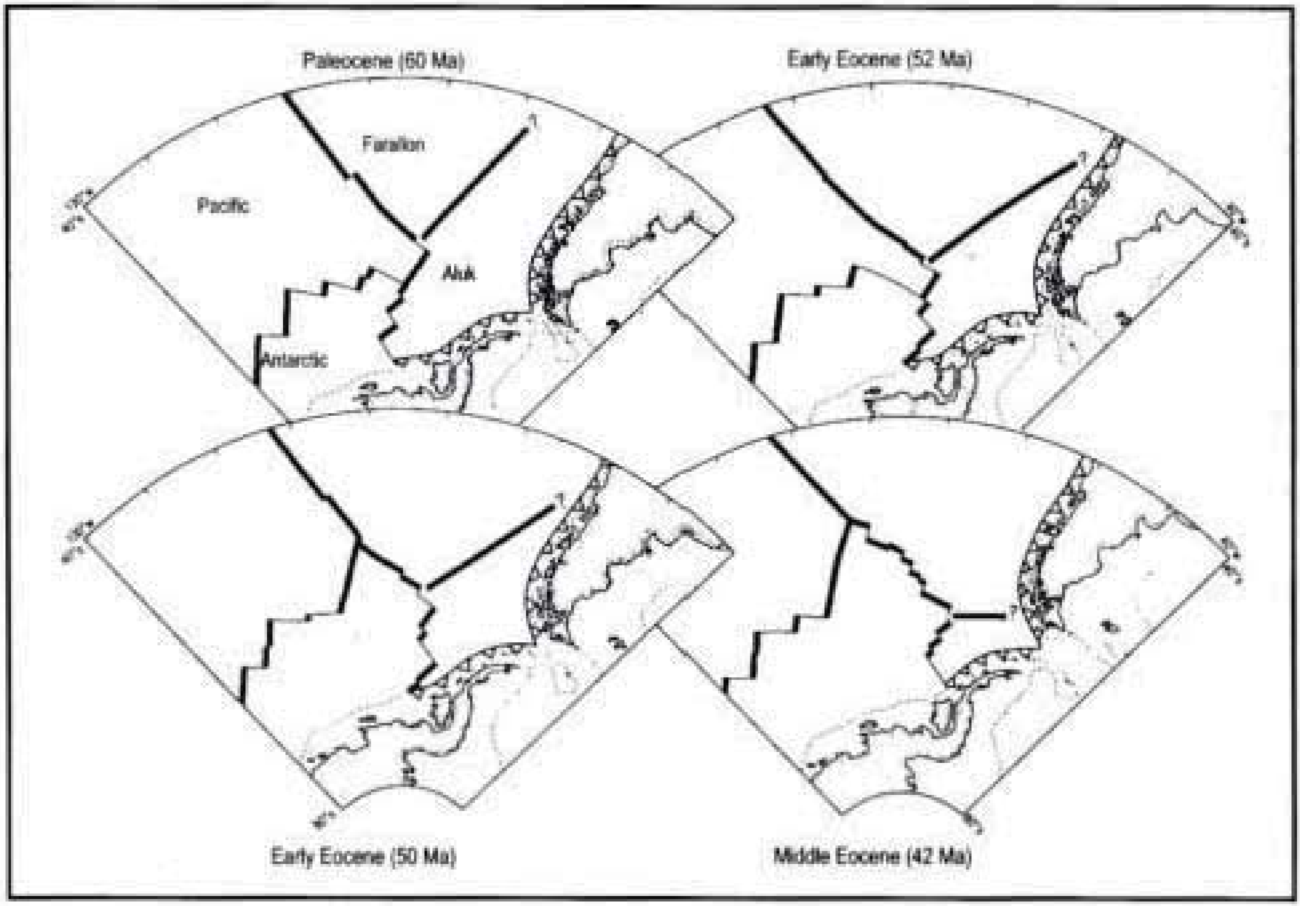


FIGURE 23: Ridge collisions along the Pacific margin based on the Paleogene and Neogene plate reconstruction of Cande and Leslie (1986).

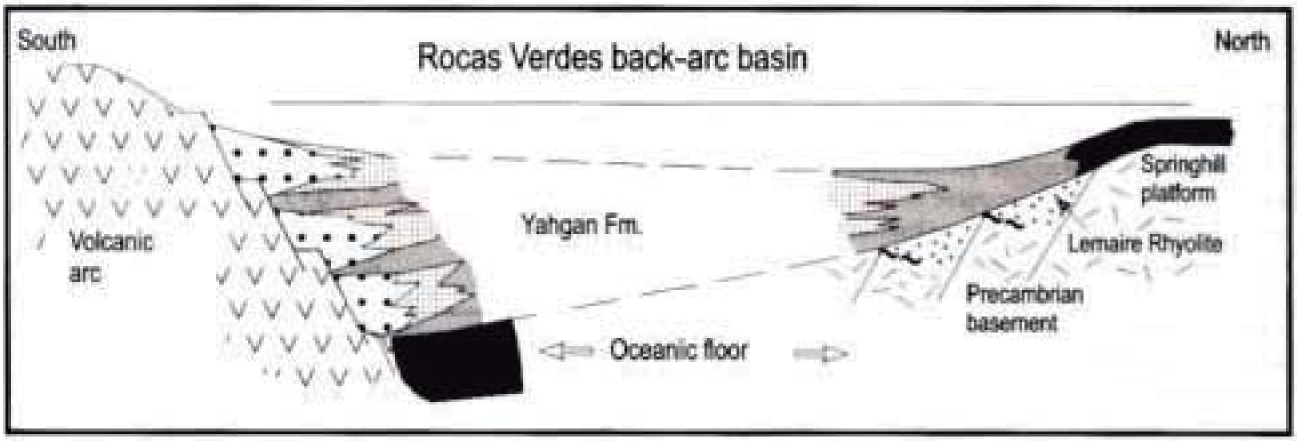


FIGURE 24: Rocas Verdes back-arc basin during Cretaceous times (modified after Dalziel et al., 1974; Olivero, 1998).

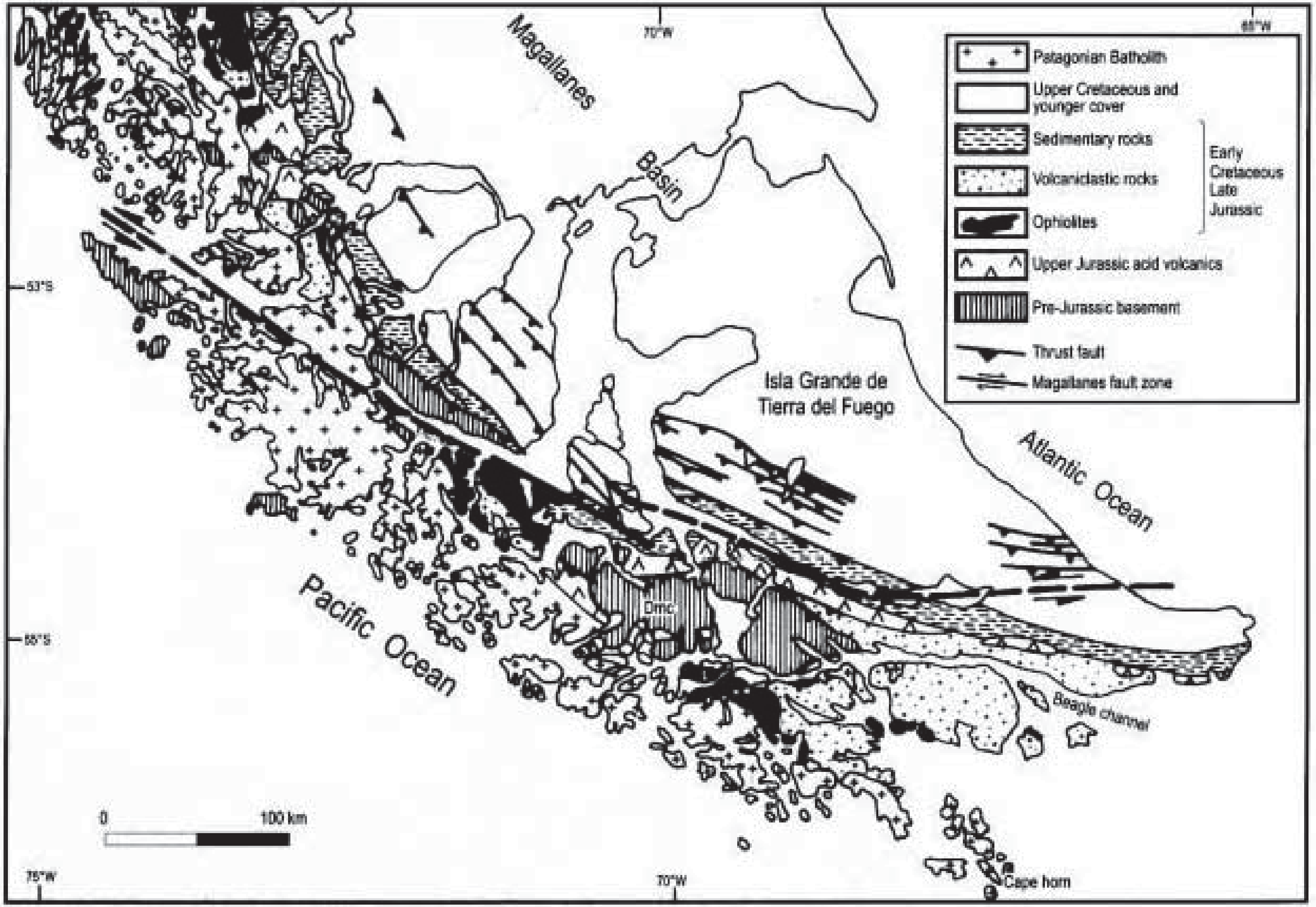


FIGURE 25: Tierra del Fuego tectonic setting with major strike-slip faults and fold and thrust belts with location of the ophiolitic assemblages of the Rocas Verdes Basin. DMC corresponds to Cordillera Darwin Metamorphic Complex (modified after Klepeis and Austin, 1997).

Important shortening in the E-W trending segment of the Fueguian Cordillera occurred in middle to late Eocene times, as demonstrated by the growth strata and limb rotation along the Atlantic coast of the Fueguian thin-skinned fold and thrust belt (Ghiglione *et al.*, 1999; Olivero and Malumián, 1999). Miocene deformation produced the present uplift of the Patagonia fold and thrust belt (Biddle *et al.*, 1986; Ramos, 1989b).

Neogene transtensional and transpressional deformation are still active along the present plate boundary between South America and Scotia plates. The Patagonia orocline appears to be the product of broad interplate shearing accommodated by strike-slip faulting, block rotation, and contraction (Cunningham, 1993).

Present plate tectonic setting of the Andes

As result of the complex and distinctive geological history of the different segments, and the oceanic plate interaction along the margin, the present tectonic setting of the Andes shows a great variety of plate tectonic conditions.

Seismicity and subduction geometry

The feature that most strongly characterizes the subduction geometry beneath the Andean Cordillera is the along-strike variation in dip of the subducted Nazca Plate from subhorizontal flat-slab segments to normal subduction (Barazangi and Isacks, 1976; Pennington, 1981; Cahill and Isacks, 1992). The seismic energy released above flat-slab segments is on average 3 or 5 times higher than in the adjacent steeper areas (Gutscher and Malavieille, 1999). Based on the interplate seismicity, several segments of flat-slab subduction have been recognized along the Andes (Fig. 26).

The Northern Andes N of 5°N correspond to the Bucaramanga segment, where flat-slab subduction has been recognized along the Colombian margin (Pennington, 1981). In this area important intraplate seismicity is recorded in the upper plate, and crustal thickening is combined with significant strike-slip motion. The large Bucaramanga earthquakes characterize the notable intraplate activity of this segment (Kellogg and Bonini, 1982) as well as the lack of active volcanism (Hall and Wood, 1985).

The other flat-slab segment corresponds to a 1500 km long sector of Peruvian Andes and is preserved between the gulfs of Guayaquil and Arequipa (5°-14°S). Along this segment of the Central Andes occurs the Cordillera Blanca tectonic uplift, one of the highest regions in the Andes. Important intraplate shallow seismicity is detected in the Eastern Cordillera and the Subandean Zone as reported by Suárez *et al.* (1983) and Dorbath *et al.* (1991), that accounts for present 4 mm/y shortening in central Peru. This segment is also characterized by the lack of volcanism and by a subhorizontal oceanic slab, dipping about 5° to the E and NE for several hundred kilometres (Barazangi and Isacks, 1976). Near 14°S there is an abrupt transition to a more steeply inclined zone (Cahill and Isacks, 1992). This flat-slab segment has been explained by the combined buoyancy of the Inca Plateau and the Nazca Aseismic Ridge subducted beneath the continental margin of Peru (Gutscher *et al.*, 1999a).

The third and southern subhorizontal subduction zone corresponds to the Pampean flat-slab segment, and is developed between 27°S and 33°S. The geometry of this segment is well established through a local seismic network (Smalley and Isacks, 1990) that defined a 300 km horizontal segment, with resumption of eastward descent below 125 km farther E (Cahill and Isacks, 1992). This segment focuses a high intraplate seismicity and is characterized by the lack of volcanism (Fig. 27) and the foreland uplift of Sierras Pampeanas (Jordan *et al.*, 1983). As a result, it concentrates the highest mountains along the Main Andes, such as the Aconcagua Massif (Ramos *et al.*, 1996b, Cristallini and Ramos, 2000).

Between the Bucaramanga and the Peruvian flat-slab segments there is a normal subduction sector where the slab dips 35° in the Cauca and Ecuador segments (Pennington, 1981; Gutscher *et al.*, 1999b). Farther S, between Arequipa and northern Argentina there is the most important normal subduction sector, where the slab inclines 30° to the E. This sector of southern Peru-Bolivia-northern Argentina, has an abrupt dip change between 14° and 16°S in the N, and a smooth transition between 24° and 27.5°S in the S. This central sector encompasses the best developed mountain chains, and is the paradigm of Andean-type orogen (Allmendinger *et al.*, 1997). To the S of 33°S the Andes have small variations in the Wadati-Benioff Zones, and a decreasing seismic activity. The normal subduction extends from 33°S to 46°30'S up to the Chile triple junction between the Nazca, South American, and Antarctic plates.

The shallow seismicity located within the Andes decreases drastically to the S of the Northern Andes, where shallow epicenters are restricted to two seismic belts, one in the fore arc and the other in the Subandean Belt (Ego *et al.*, 1996). These authors related the shallow seismicity and the absence of important seismicity in the Subandean Belt to the obliquity of the subduction. Sectors with more orthogonal subduction developed an active Subandean Belt, with little seismicity beneath the main Andes (Suárez *et al.*, 1983).

The Wadati-Benioff geometry dips at 30° to the E between 33°S and 36°S, and is deepening up to about 40° farther S (López *et al.*, 1997).

To the S of the triple junction seismicity resumes, as well as the convergent rate, that drops from 9 to 2.2 cm/y (Corvalán, 1981). The almost orthogonal subduction gives place to a more oblique subduction around 50°S with dominant strike-slip focal mechanisms, to end in a strike-slip transform boundary with the Scotia Plate (Dalziel, 1986).

Volcanic arc and related magmatism

The subduction geometry defines four distinctive zones of active volcanism: the Northern Volcanic Zone (5°N to 2°S) developed along the Cauca and Ecuador segments; the Central Volcanic Zone (16°S to 26°S) between southern Peru and Northern Chile; the Southern Volcanic Zone (34°S to 46°30'S); and the Austral Volcanic Zone, S of 47°S (Thorpe and Francis, 1979; Thorpe, 1984; Stern and Kilian, 1996). Each volcanic zone has its own peculiarities.

The Northern Volcanic Zone comprises a series of active volcanoes developed in the Western and Central Cordilleras

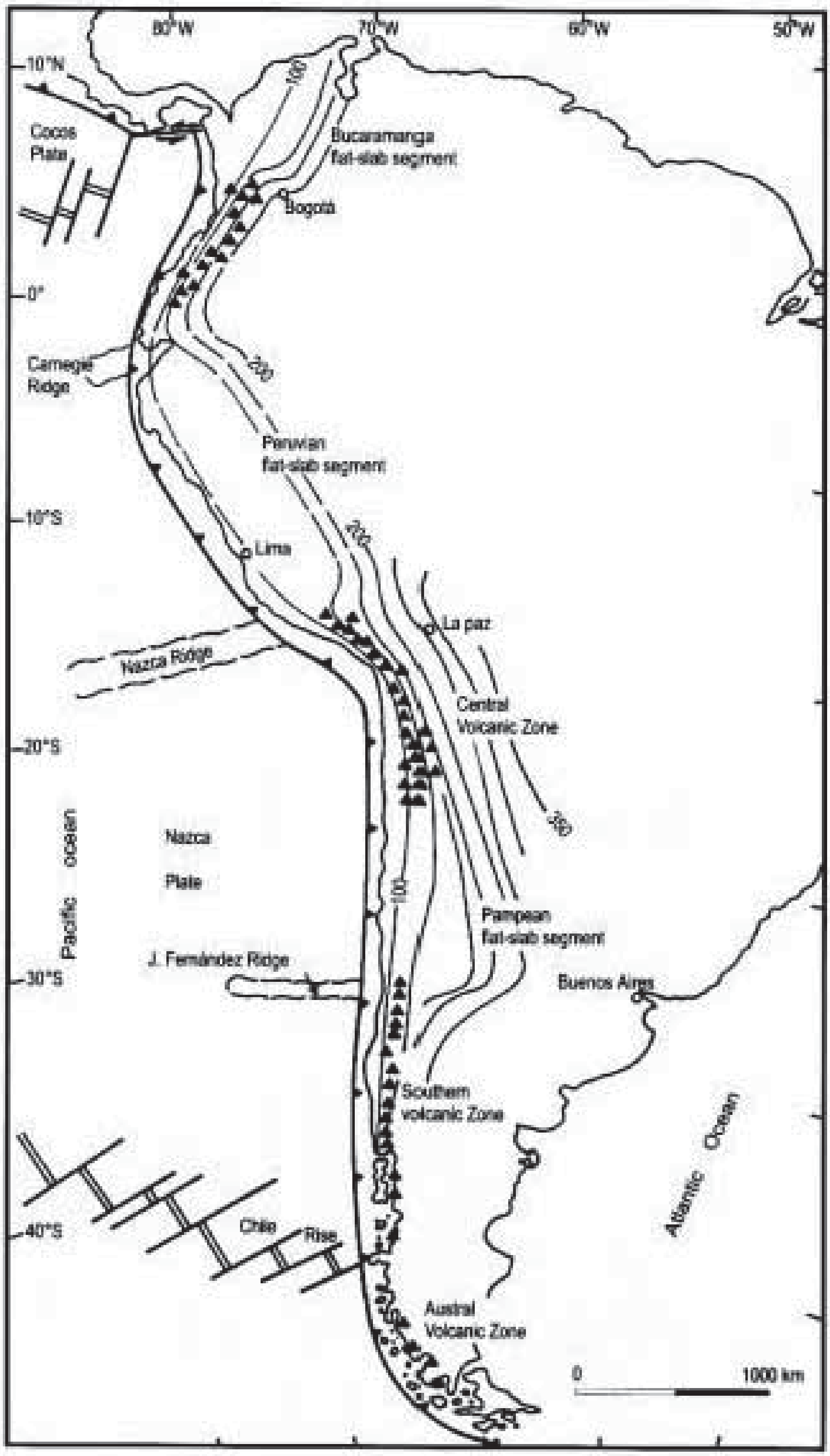


FIGURE 26: Geometry of the subduction zone along the Andes with indication of the major segments (modified after Pennington, 1981; Cahill and Isacks, 1992; López et al., 1997; Gutscher et al., 1999a).

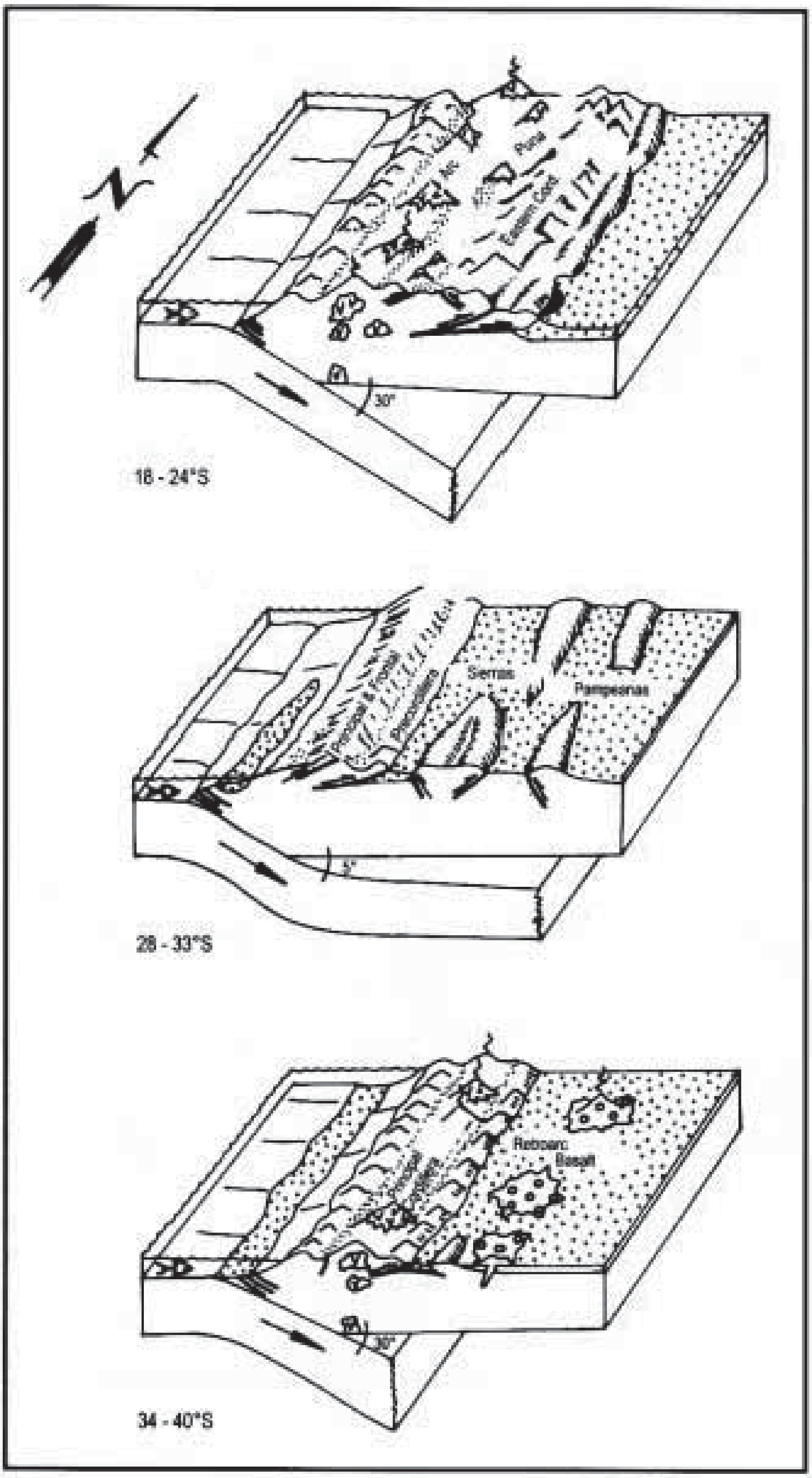


FIGURE 27: Present segmentation of southern Central Andes showing the main morphostructural units of the normal subduction and Pampean flat-slab segments (modified after Jordan et al., 1983).

of Colombia to the S of 5°N, such as the Nevados de Ruiz, Galeras, and Cerro Bravo (Méndez Fajury, 1989). This zone continues on both sides of the Interandean Valley of Ecuador where there occur several volcanoes such as the Mojanda, Chimborazo, and Pichincha persist up to 2°S (Hall and Beate, 1991; Robin *et al.*, 1997). Some other volcanoes are developed in the eastern foothills of the Andes, such as the Reventador and Sumaco (Hall and Calle, 1982).

The volcanic rocks of these volcanoes range from basaltic andesite to andesite, and are more primitive than the products of the Central Volcanic Zone (Thorpe, 1984). They are situated in the vicinity of the suture between the Piñón-Dagua oceanic terrane and other Jurassic terranes accreted during Early and Late Cretaceous to the Paleozoic proto-margin of Gondwana. Their volcanic rocks are consistent with derivation from fractional crystallization of basaltic magma produced from partial melting of the asthenosphere wedge containing components from oceanic lithosphere. However, their low to moderate Sr^{87}/Sr^{86} ratios (0.7036 to 0.7046), as well as the higher Pb and O-isotope ratios may indicate some assimilation of younger continental crust (Harmon *et al.*, 1984). Lava flows, small pyroclastic flows, and large lahar deposits, associated with ashfall deposits widely cover the Interandean Valley of Ecuador (Hall and Calle, 1982).

The Central Volcanic Zone is widely developed between Arequipa and northern Chile, along the Western Cordillera, which is bounding the Altiplano-Puna high plateau. Hundreds of volcanoes are widely spread along this region. These volcanoes are characterized by their eruption in a thick crust, in places over 70 km thick, and record high degrees of differentiation. Stravolcanoes constructed largely from andesitic and dacitic lava are dominant in additon to which there occur significant volumes of dacitic ignimbrites of latest Cenozoic age (Davison *et al.*, 1993). A Sr^{87}/Sr^{86} ratio varies from 0.7056 to 0.7149 (Harmon *et al.*, 1984). There is also a striking correlation between the Pb-isotopes of the volcanic rocks and those of the underlying basement (Wörner *et al.*, 1994). These facts, together with the chemical and isotope composition, led Hildreth and Morbath (1988) and Davidson *et al.* (1993), to assume that extensive modification of the mantle-derived magma took place as they ascended through an exceptionally thick crust. The amount of contamination has been evaluated in the order of 35% to 70% (Hawkesworth and Clarke, 1994). On the other hand, subduction of oceanic and terrigenous sediments into the asthenosphere wedge, as well as subduction erosion, could have contribute to the differentiation of the mantle-derived magma (Stern, 1991). Subduction geometry changed during Late Cenozoic times as recorded by the magmatism and by an extensive episode of rhyolitic and dacitic calderas and ignimbritic flows formed during an episode of steepening of the Wadati-Benioff Zone (Coira *et al.*, 1993; Kay *et al.*, 1999). This extensive silicic volcanic province produced in the late Miocene, known as the Altiplano-Puna volcanic complex, covers more than 50 000 km^2 and is one of the largest ignimbrite concentration in the world (De Silva, 1989). Teleseismic broadband records of earthquakes in this province at Bolivia have identified an active crustal magma flat body at 19 km depth, very extensive and 750-810 m thick (Chmielowski *et al.*, 1999). The southern part of this Central Volcanic Zone records a crustal delamination associated with mafic magma and extension (Kay *et al.*, 1994).

The Southern Volcanic Zone is developed between 33°30'S and 46°30'S, and corresponds with the southern part of the Central Andes. It comprises late Cenozoic and active volcanoes such as Tupungato, San José, Lonquimay, and Hudson, mainly developed in the Chilean slope of the main Andean Cordillera. The northern sector of this 1000 km long volcanic chain has more crustal influence and is characterized by andesite and dacite (López Escobar *et al.*, 1995). South of 37°S the volcanic province consists of rocks of basalt to rhyolite composition, but with a predominance of basalt and basaltic andesite with low Sr^{87}/Sr^{86} ratios (0.7037 to 0.7044). This volcanic province is heavily controlled by the onset of important strike-slip faults as the Iquiñe-Ofqui (Hervé, 1994). The increased angle of subduction S of 35°S from 30° to near 40° beneath the volcanic zone as well as the migration toward the trench recorded since Pliocene times, may account for a minimum coupling between the Nazca and the South American plates, and the dominant poorly differentiated volcanism (Stern, 1990).

The Austral Volcanic Zone has been recently defined by Stern and Kilian (1996). This volcanic zone consists of a few volcanoes developed in the Southern or Patagonian Andes S of the volcanic gap (Fig. 21) associated with the ridge subduction (Stern *et al.*, 1976b). Adakitic volcanic rocks of low Sr^{86}/Sr^{87} ratio, formed by components of the asthenosphere wedge, plus a partial melting of the subducted slab constitute the poorly evolved lava of the Lautaro, Aguilera, Diablo, Burney and Cook volcanoes (Stern and Kilian, 1996).

The areas of slab-flat subduction record the shifting, expansion, and the cessation of the volcanic arc through time, with striking compositional changes, declining volumes of volcanic rocks, and unusual petrological characteristics (Kay *et al.*, 1991; Ramos *et al.*, 1991; Kay and Abbruzzi, 1996).

The normal subduction southern segment of the Central Andes records an important retro-arc basaltic magmatism of alkaline composition. It is associated with trenchward migration of the volcanic front (Muñoz and Stern, 1988); a decreasing age of the oceanic crust being subducted (Ramos and Barbieri, 1989) and the presence of transient hot spots (Kay *et al.*, 1993a). Farther S, along the Southern Andes, asthenosphere windows formed after by ridge subduction controls the near trench magmatism, adakite and retro-arc plateau basalt (Ramos and Kay, 1992; Kay *et al.*, 1993b).

Crustal thickening and orogenic shortening

The convergence vector and subduction rate between South America and the Farallon Plate during the Paleogene (Pilger, 1984) favoured a more orthogonal convergence on the Peruvian margin and in the Fueguian Andes. The continental margins in these sectors have northwestern trends, and therefore the effects of the high convergence rates by the end of the Eocene were more severe. These two segments show an important compression during the Incaic deformation (Vicente, 1972; Galeazzi, 1996). On the other hand, along the Chilean margin stress partitioning of this oblique convergence reactivated strike-slip displacements (Mpodozis *et al.*, 1994) along the Domeyko Fault System and the western fissure (Tomlinson *et al.*, 1994).

The break-up of the Farallon Plate into the Cocos and Nazca plates, which occurred at about 25 Ma, seems to mark the beginning of a period of higher and more orthogonal convergence rates in most of the Central and Southern Andes (Pardo Casas and Molnar, 1987). This age coincides with the initiation of widespread Miocene magmatism and is a milestone in the geodynamic evolution of the area. The convergence vector and subduction rate were more important along the Chilean margin, during the middle and late Miocene (26 - 8 Ma), time of onset of major Miocene deformation generally assigned to the Quechua Orogeny (Mpodozis and Ramos, 1990).

The Northern Andes underwent an important stress partitioning during the Paleogene and the Neogene, due to the northeastern trend of the continental margin. As a result of that the oblique component of plate motion is taken up by dextral slip displacements in a series of crustal discontinuities (Fig. 11), such as the Pujili-Cauca and Peltetec-Romeral faults (Dewey and Lamb, 1992; Mégard, 1987).

As a result of several pulses of compression, generally grouped in the Incaic and Quechua deformations, important crustal thickening took place along the Andes. Maximum crustal thickening is observed along the central sector of the Central Andes, where orogenic shortening was the largest (Allmendinger *et al.*, 1997). New crustal balances made across Northern Chile and Bolivia, combined crustal shortening constrained by refraction seismic and partial seismic reflection surveys, and orogenic shortening derived from structural cross-sections (Fig. 28) (Schmitz, 1994; Kley *et al.*, 1999). These balances indicate up to 320 km of total shortening during the Cenozoic. This area also has the thickest crust of the Andes beneath the Western Cordillera at about 20°S. Broadband seismic analyses indicate a crust some 70-74 km thick (Beck *et al.*, 1996).

There is a well-defined gradient to the N and S of this shortening, coherent with fore-arc rotation constrained by paleomagnetism (Beck, 1998). The crustal thickening and orogenic shortening gradients are well established in the southern segment of the Central Andes (22°S to 46°30'S). There is a continuous decline in the thickening of Andean roots with orogenic shortenings from 160-40 km at 30°S-32°S (Introcaso *et al.*, 1992; Ramos *et al.*, 1996b) up to 44 to 20 km at 37°S-39°S (Martínez *et al.*, 1997).

This decrease in crustal thickness and orogenic shortening is also observed from the Altiplano to the Peruvian Andes (Cabassi *et al.*, 1999). Both gradients, from the Central Andes to the N and S, can be correlated with a decrease in the age of oceanic crust being subducted. There is a continuous trend of younger oceanic crust along the trench as the Chile Ridge gets closer to the present triple junction (Tebbens *et al.*, 1997).

Orogenic shortening and tectonic styles

The estimates of crustal thickening and the consequent orogenic shortening impose important constraints to the tectonic style of the Andean Cordillera. Studies in the last 20 years have shown contrasting attempts to understand the mode of structural deformation along the different foreland thrust belts of the Andes. Models with high angle basement thrusts (Mégard, 1987; Zeil, 1979) alternate in similar regions with thin-skinned deformation (Vicente, 1972; Ramos, 1988b; Allmendinger *et al.*, 1990). Tectonic inversion of previous normal faults in recent years has become one of the significant mechanism of thrusting (Daly, 1989; Grier and Dalmeyer, 1990; Manceda and Figueroa, 1995; Ramos *et al.*, 1996b; Cristallini *et al.*, 1997; Colletta *et al.*, 1997; Kley *et al.*, 1999). Those areas where seismic control is suitable either show: (1) unequivocal evidence of thin-skinned thrusting as the Subandean fold and thrust belt of Bolivia and Northern Argentina (Baby *et al.*, 1992; Mosquera, 1999); or (2) salt detachment of the Santiago fold and thrust belt of Peru (Alemán and Marksteiner, 1997); or (3) tectonic inversion with subordinate thin-skinned tectonics as in the Eastern Cordillera of Colombia (Cooper *et al.*, 1995). However, in areas where the seismic control is inappropriate, crustal balance can give a further constraint for the amount of shortening. For example, in the Subandean Neuquén Basin (37-39°S), several authors have proposed different tectonic styles from thin to thick-skinned deformation varying from hundred to ten of kilometres. As indicated by geophysical studies the crustal shortening does not exceed 44 km (37°S) to 20 km (39°S). These data constrain the structural style to thick-skinned systems such as proposed by Kozlowski *et al.* (1993) or Zapata *et al.* (1999), with less than 40 km shortening.

In the Northern Andes, where stress partitioning superimposed important strike slip-displacements, crustal thickening of Cenozoic deformation is more complex to evaluate. Shortening in the most active Neogene region of Eastern Cordillera of Colombia shows a minimum of 68 km (Cooper *et al.*, 1995) of the same order than the Mérida Andes with 50 to 60 km (Colletta *et al.*, 1997).

In spite of the overwhelming evidence of the importance of tectonic inversion of normal faults in the Andes, there is good testimony in some areas for thin-skinned deformation (Ramos *et al.*, 1996b). Stress partitioning under oblique subduction as proposed by Dewey and Lamb (1992) produced notable strike-slip displacements as detected in the Northern Andes (Campbell, 1968; Dengo and Covey, 1993), or in the Southern Andes (Hervé, 1994; Diraison *et al.*, 1998).

Neotectonics and orogenic shortening

As an active orogenic belt, the Andes display significant neotectonics along the thrust front of the Central Andes and strike-slip displacements in the Northern Andean block and in the Fueguian Andes (Fig. 29).

Active faults with notable strike-slip displacements have been documented in the Boconó Fault of the Mérida Andes by Schubert (1982) and Schubert and Vivas (1993). Similar settings have been described in the Eastern Cordillera of Colombia with active faulting in the Garzón Massif and in the Guaicáramo Fault (Van der Wiel, 1991). Farther S, active faulting is recorded in the Ecuadorian Andes in the Interandean Valley, associated with Quaternary volcanism (Hungerbuehler *et al.*, 1996).

Inversion of focal mechanisms in the Northern Andes, together with neotectonic analyses, have shown differences

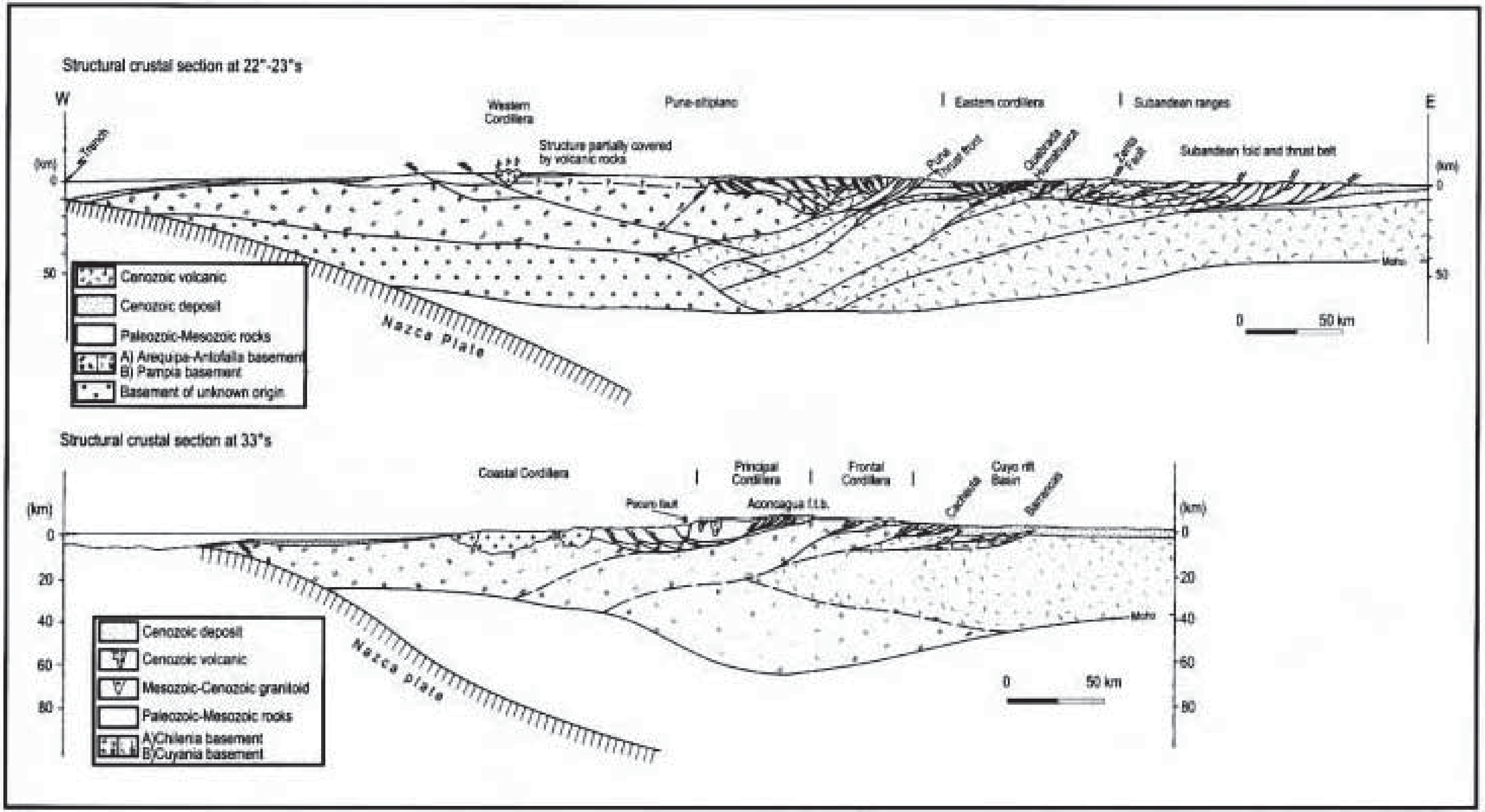


FIGURE 28: Crustal sections of the Andes showing the decrease in crustal thickness from the Altiplano region to the Southern Andes (22°S to 41°S)(modified after Schmitz, 1994; Introcaso et al., 1992; Ramos et al., 1996b).

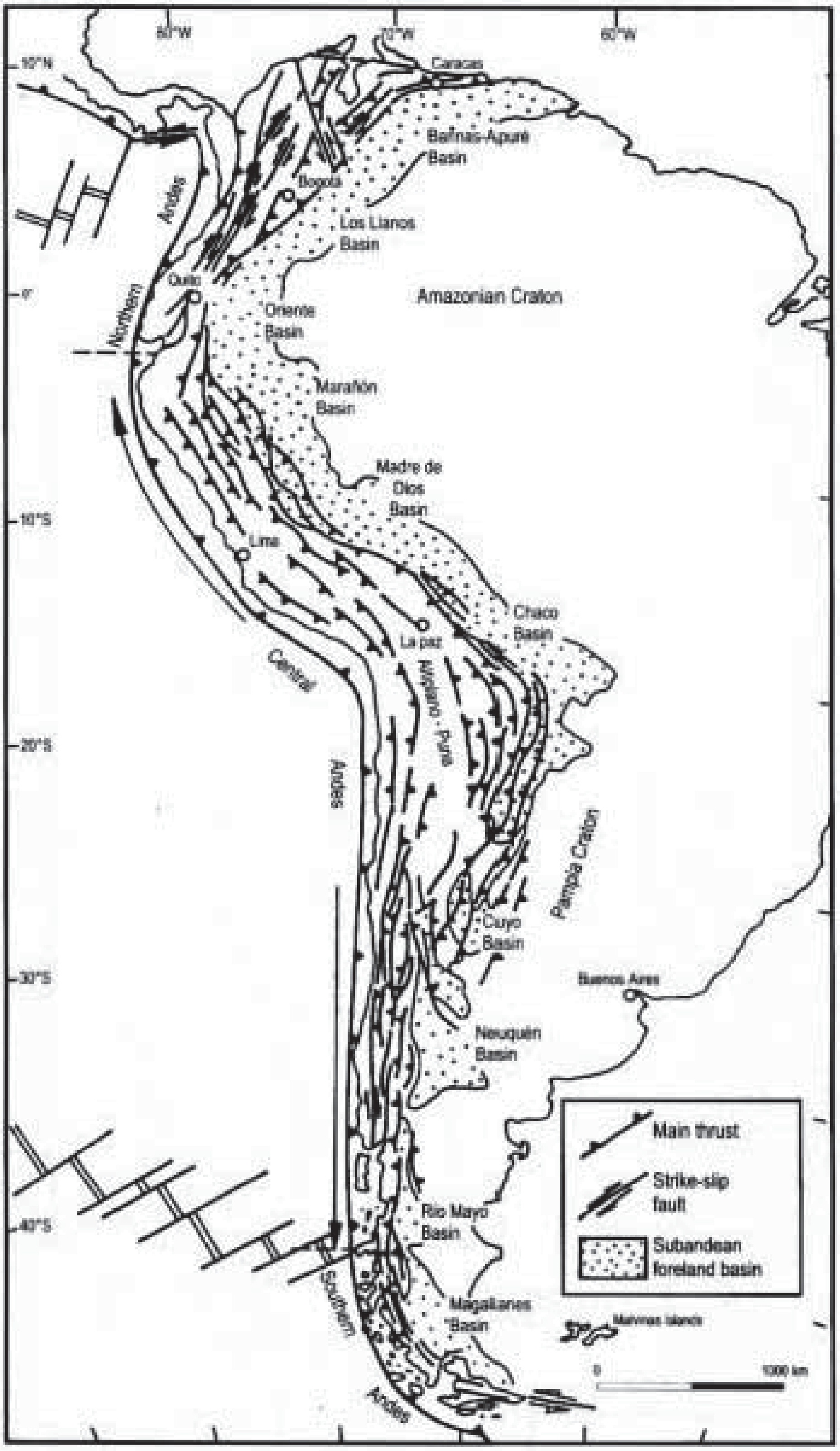
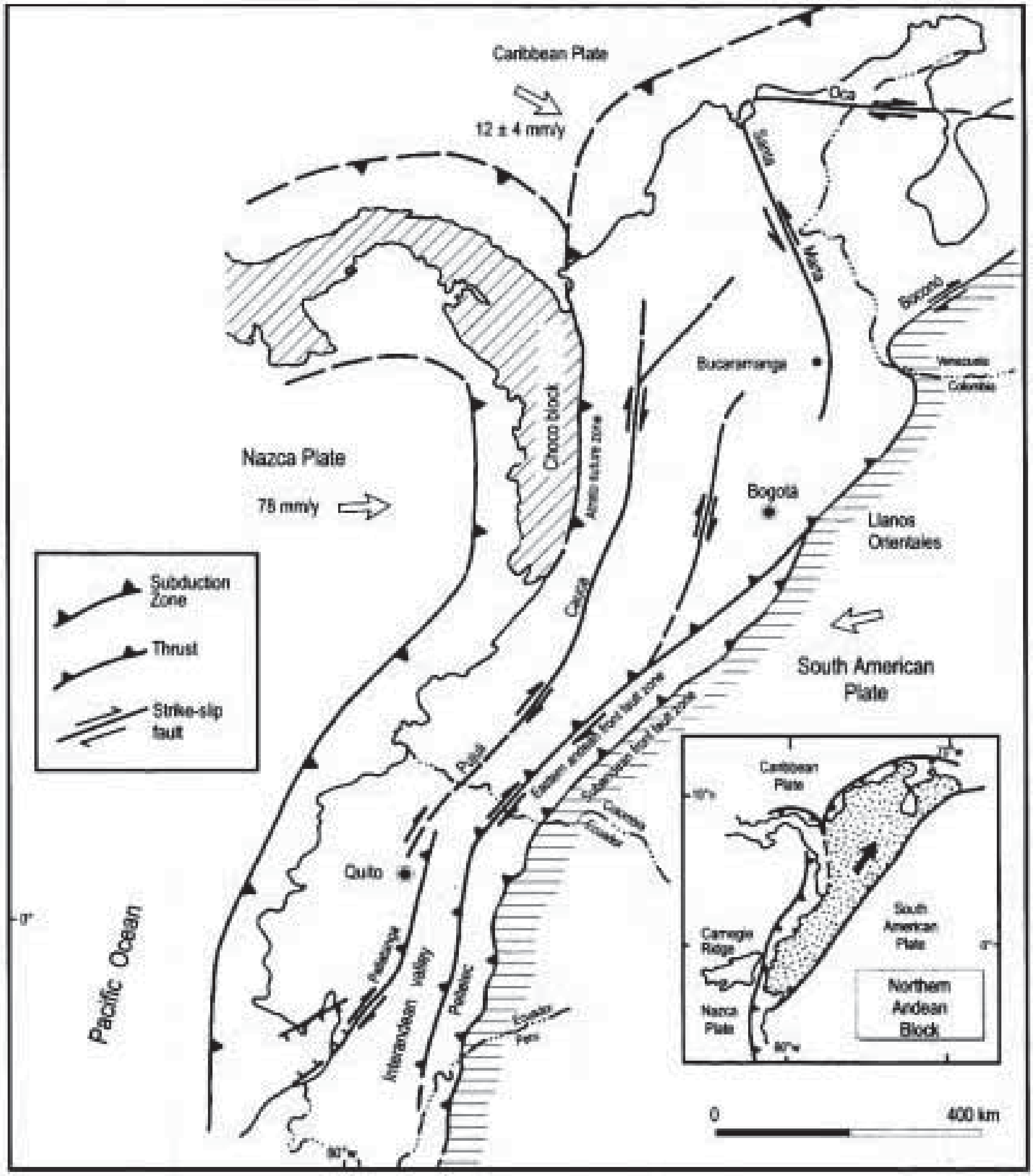


FIGURE 29: Major structural features of the Andes with major thrust fronts. Note the strike-slip faulting of the Northern Andes, and the dominant wrenching and dismembering of the Fueguian Andes (modified after Ramos, 1999).

FIGURE 30: Major active fault in the Northern Andean Block (modified after Kellogg and Bonini, 1985; Ego et al., 1996). Inset a) shows the North Andean block or microplate.



in the state of stress N and S of 5°N (Ego *et al.*, 1996). The northern sector seems to be controlled by the interaction between the Caribbean and South American plates (Fig. 30), coherent with dextral motion of the Northern Andean Block relative to stable South America as proposed by Kellogg and Bonini (1985).

This relative dextral motion of the Northern Andean Block was intensified in the last 9 Ma by the collision of the Carnegie Aseismic Ridge (Daly, 1989). As a result, a vertical uplift of 6 km took place since the late Miocene (Steinmann *et al.*, 1999), as well as important dextral strike-slip along the Peltetec-Romeral and Pujilí-Cauca fault systems. Topographic evidence along the coast indicates that collision with the trench took place at least 8 Ma ago (Gutscher *et al.*, 1999b).

Along the Peruvian Andes contrasting active tectonics took place in the flat-slab segment with contraction and reverse faults (Schwartz, 1988; Sébrier *et al.*, 1988), in comparison with normal active faults at the northern end of the Altiplano (Sébrier *et al.*, 1985). While continuous foreland blind thrusting is seen in the Chaco plains E of the Subandean Belt (Mujica and Zorzín, 1996), at the southern end of the Altiplano-Puna high plateau active faulting is dominated by normal faulting (Allmendinger *et al.*, 1997). This normal faulting was explained by collapse originated in body forces by Froidevaux and Isacks (1984) or by stress partitioning of oblique strike-slip faults by Dewey and Lamb (1992).

Farther S, active faulting is occurring in the Pampean flat-slab segment (27°S-33°S) where important shortening has been described in the Precordillera and Sierras Pampeanas (Bastías *et al.*, 1990; Costa, 1992). This activity declines to the S, where in the southern normal subduction sector of the Central Andes (34°S-38°S), is less conspicuous, until it disappears between 39°S and 46°30'S. Active faulting is again important at the latitude of the Chile triple junction,

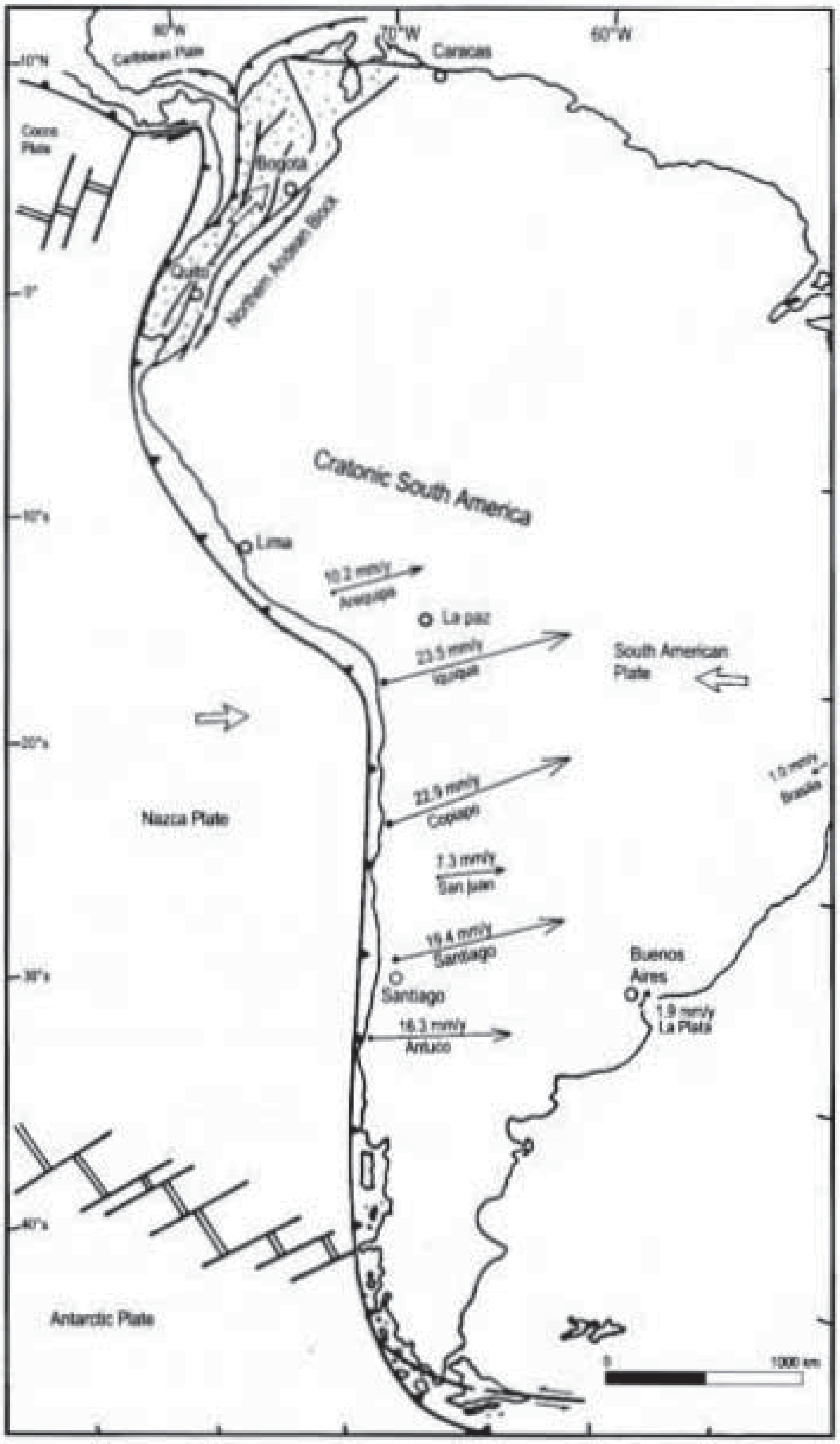


FIGURE 31: Space-based geodetic measurements along the Pacific border of the southern Central Andes in respect to stable South America. The differences between Santiago and the San Juan stations may account for the present shortening of the Andes, and between San Juan and La Plata for the Sierras Pampeanas shortening (modified after Kendrick et al., 1999).

where ridge collision is associated with Quaternary faults (Ramos, 1989b). Farther S, active strike-slip faulting controls the onset of intraplate basaltic flows along the northern part of the Magallanes Strait.

New space geodetic data record rates and direction of motion across the Andes, mainly between the continental margin affected by the Nazca Plate convergence and stable South America (Norabuena *et al.*, 1998). Recent data presented by Kendrick *et al.* (1999) and illustrated here in Figure 31 show significant shortening rates between the western slope of the Andes and the average position of stations situated in cratonic stable areas of Argentina and Brazil. The total relative motion is the result of several components such as (1) transient elastic deformation on the lock portion of plate interface, that can be released during large thrust earthquakes; and (2) permanent deformation through crustal shortening and mountain uplift. In the assessment of this permanent deformation it is important to weight the Andean shortening. The amount of shortening between Santiago de Chile (19.4 mm/y), San Juan (7.3 mm/y) and La Plata (1.9 mm/y) permits the evaluation of the active shortening within the Main Andes-Precordillera and Sierras Pampeanas in the Pampean flat-slab of Argentina and Chile. These figures indicate a shortening between both slopes of the Andes of 12 mm/y, and within Sierras Pampeanas of 5.4 mm/y. If these figures are compared with those derived from crustal balance of Andean roots (7.65 mm/y) or from structural cross sections (5.25 mm/y) in the Main Andes at these latitudes, the G.P.S. results are higher. This fact may indicate either a concentration of elastic deformation along the continental margin or an increase in recent years of the average Neogene shortening. More suitable figures are obtained when the structural shortening computed for Sierras Pampeanas, mainly the Pie de Palo area (5 mm/y in the last 3 Ma), one of the most active areas, is compared with the G.P.S. data (5.4 mm/y). The similar values may indicate that elastic deformation accumulated in this region is minimum, and if it existed, it was released by the large Pie de Palo earthquake in 1977 (Smalley *et al.*, 1993).

Although these values are still preliminary, they illustrate that space-based geodesy is opening a new era for studies of plate convergence along the Andes.

Concluding remarks

The previous analyses of the formation of the Andes and the description of their present tectonic setting show some contrasting characteristics among the different segments.

Early in the evolution of the proto-Andean margin of western Gondwana during Early Paleozoic times a great variety of basement blocks are observed, presently hidden under a thick Andean cover. Some of them are para-autochthonous basement blocks derived from Gondwana, and therefore considered by different authors as peri-Gondwanan terranes. On the other hand, some of the blocks have a distinct basement in comparison with western Gondwana, as indicated by isotope analyses, paleontological evidence, paleomagnetic data, and paleoclimatic conditions. These exotic blocks, such as the Cuyania Terrane, have well defined Laurentian affinities, and therefore are interpreted as allochthonous terranes accreted to the proto-margin of Gondwana during the Ocloyic and Chanic orogenies. Prior to the final amalgamation of these peri-Gondwanan and Laurentian blocks, a proto-Andean margin stage has been recognized. The Sierras Pampeanas basement records the best evidence of an Early Paleozoic magmatic arc, with time constraints that show a parallel evolution to the sedimentary record in the approaching Cuyania Terrane.

An intriguing characteristic of these basement terranes is the dominant Grenville age signature that almost all of them share. This is being interpreted as a confirmation of the common source in the Rodinia Supercontinent, and their participation in the Grenville Orogeny that may have amalgamated this supercontinent in middle Proterozoic times.

The nature of the different Paleozoic orogenies varies from N to S. The Northern Andes are the result of terrane accretion during Early Paleozoic times, that ended with a continent-to-continent collision between Laurentia and Gondwana to form the Alleghanides in Late Paleozoic times. The Southern Andes resulted from the collage of peri-Gondwanan and exotic Laurentian terranes amalgamated in Early Paleozoic times, without continent-to-continent collision. There is good evidence to show that the successive proto-margins were always facing to an open ocean, after docking a series of terranes. The Gondwanides mountain chain was the result of a Late Paleozoic orogeny that occurred after the final amalgamation of these terranes to the southwestern margin of Gondwana in Devonian times.

There is partial evidence that a magmatic arc occurred along the present continental margin during Late Paleozoic times. However, by the end of the Permian subduction there was a lull, and a well-developed generalized extension that predates the final break-up of the Pangea Supercontinent. The distribution of the rift systems and the associated magmatism shows that the extension was mainly concentrated in the hanging-wall of the suture between large cratonic basement terranes.

The opening of the North and South Atlantic seas induced another period of renewed subduction, but in the early stages linked to extension. Negative trench rollback velocities are in agreement with the northeastern true polar wandering of South America still connected with Africa. A reorganization of the stress state in the Andes produced a shift to compression that occurred between 115 and 105 Ma, and which is generally explained as the result of a change in the absolute motion of South America. The drift stage was completely obtained after the final break-up of South America and Africa at about 80 Ma. Oblique convergence vectors, regulated by the local orientation of the continental margin controlled strike-slip faulting in the fore arc.

Accretion of oceanic terranes in Colombia and Venezuela, most of them associated with oceanic plateaux derived from the Caribbean Plate, was the result of the interaction of this plate with South America. Island arc terranes have been accreted in Ecuador. These accretions took place in three different stages: in Early Cretaceous, Late Cretaceous to Paleogene, and in middle Miocene times. Obduction of oceanic basement, associated metamorphism and deformation characterized these stages.

The Incaic late Eocene-early Oligocene orogeny was important in some segments of the Andes, such as the Peruvian and Fueguian margins, with more orthogonal orientation to the slip vector of the oceanic plates.

The break-up of the Farallon oceanic plate in the late Oligocene-early Miocene, and the formation of the Cocos and Nazca plates was another major reorganization in the tectonics of the Andes, that led to the present plate tectonic setting of the different Andean active margin segments.

Although in most cases the Andean uplift was controlled by these oceanic factors, it is possible that uplift may also have resulted from the tectonic inversion of a complex extensional setting not directly related to subduction and arc magmatism as in the case of the Mérida Andes. When analyzed through time the stress regime in normal Andean-type segments shows:

1) The magmatic arc was either stable or oscillatory within few kilometres. Deformation in these cases shows minor shortening.

2) The magmatic arc and deformation migrated tens of kilometres following tectonic erosion of the continental margin.

3) The magmatic arc and deformation steadily migrated to the foreland in function of the changes in the Wadati-Benioff geometry.

4) The magmatic arc expanded and migrated to the foreland until cessation of magmatism and the development of a flat-slab setting. This produced strong deformation with foreland basement block-faulting in mature settings.

5) The magmatic arc is associated with near-trench magmatism, followed by strong deformation, and the development of retro-arc basaltic plateaux. Arc magmatism resumed after few millions years with an adakitic character.

6) The magmatic arc is retreated to the trench development in an extensional regime during the steepening of the angle of subduction, locally accompanied by the development of huge calderas with acid rocks, or extensive basaltic flows.

These variations are closely related to the interaction with the adjacent oceanic plates, the age of the oceanic crust being subducted, the collision of aseismic and seismic ridges, as well as the convergence vectors. Basement fabrics and earlier local geological history impose additional constraints on the resulting deformation.

There is general agreement that shallowing of the subduction zone is related to an increase in deformation and shifting of the magmatic arc. However, the effects of the steepening of the subduction zone are not well established. The trenchward migration of the magmatic arc is one of the first assumed pieces of evidence for this steepening. In some cases this is closely linked with extension and basaltic magmatism on a vast scale in the retro-arc settings. In other cases, there is no evidence for extension and we have generalized ignimbritic flare-up in huge calderas.

The space-based geodetic observations are permitting the evaluation of instantaneous deformation and its comparison with long-term deformation rates. This is opening an exciting field in the understanding of the relationships between the Andean uplift and its interaction with adjacent oceanic plates.

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