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Tectonic Evolution of the Andes of Ecuador, Peru, Bolivia and Northernmost Chile

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This chapter was prepared under the co-ordination of E. Jaillard. Together with G. Hérail and T. Monfret, he wrote the Introduction. Enrique Díaz-Martínez prepared the section on the Pre-Andean evolution of the Central Andes. Again Jaillard, on the Pre-orogenic evolution of the North-Central Andes. E. Jaillard, P. Baby, G. Hérail, A. Lavenu, and J.F. Dumont wrote the text on the orogenic evolution of the North-Central Andes. And, finally, Jaillard closed the manuscript with the conclusions.

Introduction:

The Present-Day North-Central Andes (1°N - 23°S)

The Andean Chain is the major morphological feature of the South American Continent. This 8000 km long mountain belt extends along the western border of the South American Plate, and can be divided into three segments of distinct orientations, separated by two major bends. The NNE-SSW trending Colombian-Ecuadorian segment (12°N - 5°S) is 2000 km long and includes part of northernmost Peru and easternmost Venezuela. It is separated from the Peruvian segment by the Huancabamba Bend (Mégard, 1987). Its northern end (eastern Venezuela) exhibits a change to an E-W orientation, due to its connection to the South Caribbean dextral transform system. The Peruvian segment (5°S - 18°S) is 2000 km long and its orientation is close to NW-SE. It includes northern Bolivia, and is separated from the Chilean segments by the Arica Bend. The Chilean segment (18°S - 56°S) is 4000 km long and trends N-S. Its southern tip exhibits a change to an E-W direction along the Scotia sinistral transform system.

The width of the Andean Chain varies from 250 km in northern Peru (5°S) or southernmost Chile (52°S - 55°S), to as much as 750 km in Bolivia (18°S).

The area described in this chapter includes Ecuador, Peru, Bolivia and, therefore, northernmost Chile (N of 23°S). It includes parts of the three segments described above, the orientation of which played a significant role in the pre-orogenic period.

The lowest pass of the studied area is situated in northern Peru (Abra de Porculla, 2300 m), a zone where the chain is very narrow. The highest average altitude is reached between 15°S and 23°S, where the Altiplano of Bolivia and southern Peru reaches a nearly 4000 m of average elevation, and corresponds to the widest part of the chain. The Andean Chain is usually highly asymmetric, with a steep western slope, and a large and complex eastern side. In Peru, the distance between the trench and the hydrographic divide varies from 240 to 300 km, whereas the distance between the hydrographic divide and the 200 m contour line ranges between 280 km (5°N) and about 1000 km (Lima Transect, 8°S - 12°S). In northern Chile and Argentina (23°S), these distances become 300 km and 500 km, respectively. In southern Peru, as little as 240 km separates the Coropuna Volcano (6425 m) from the Chile-Peru Trench (- 6865 m). This, together with the western location of the Andes relative to the South American Continent, explains why the rivers flowing toward the Pacific Ocean do not exceed 300 km long, whereas those flowing to the Atlantic Ocean reach 4000 km long. We have to note two exceptions to this rule.

In Ecuador this asymmetry disappears, due to peculiar tectonic history and deformational processes. The trench-hydrographic divide distance roughly equals the distance between the water divide and the 200 m contour line, and ranges between 280 and 350 km. Between 13°S and 24°S (southern Peru-Bolivia), there are two hydrographic divides, delimiting a wide, flat, endorheic basin known as the Altiplano, which coincides with the zone of highest average elevation and largest width of the chain (Fig. 1).

The highest summits are usually recent or active volcanoes situated on the deformed chain or on metamorphic or granitic slices uplifted by reverse faults. Among the former are the Cotopaxi (5897 m) and Chimborazo volcanoes (6310 m) in Ecuador, the Coropuna (6425 m) and Ampato volcanoes (6310 m) in Peru, and the Sajama Volcano (6520 m) in Bolivia, near the Bolivia-Chile border. Among the latter, we may note the mainly granitic Cordillera Blanca in Peru, which culminates at 6768 m (Nevado Huascarán), and the metamorphic Eastern Cordillera of southern Peru and northern Bolivia (Nevado Illimani, 6682 m; Nevado Illampu, 6485 m). However, deformed and uplifted sediments may also form high summits in Peru, such as the Nevado Yerupajá (6632 m) in the Cordillera Huayhuash and the Nevado Ausangate (6384 m) in southern Peru or the Cordillera de Apolobamba (Nevado Cololo, 5975 m) in Bolivia.

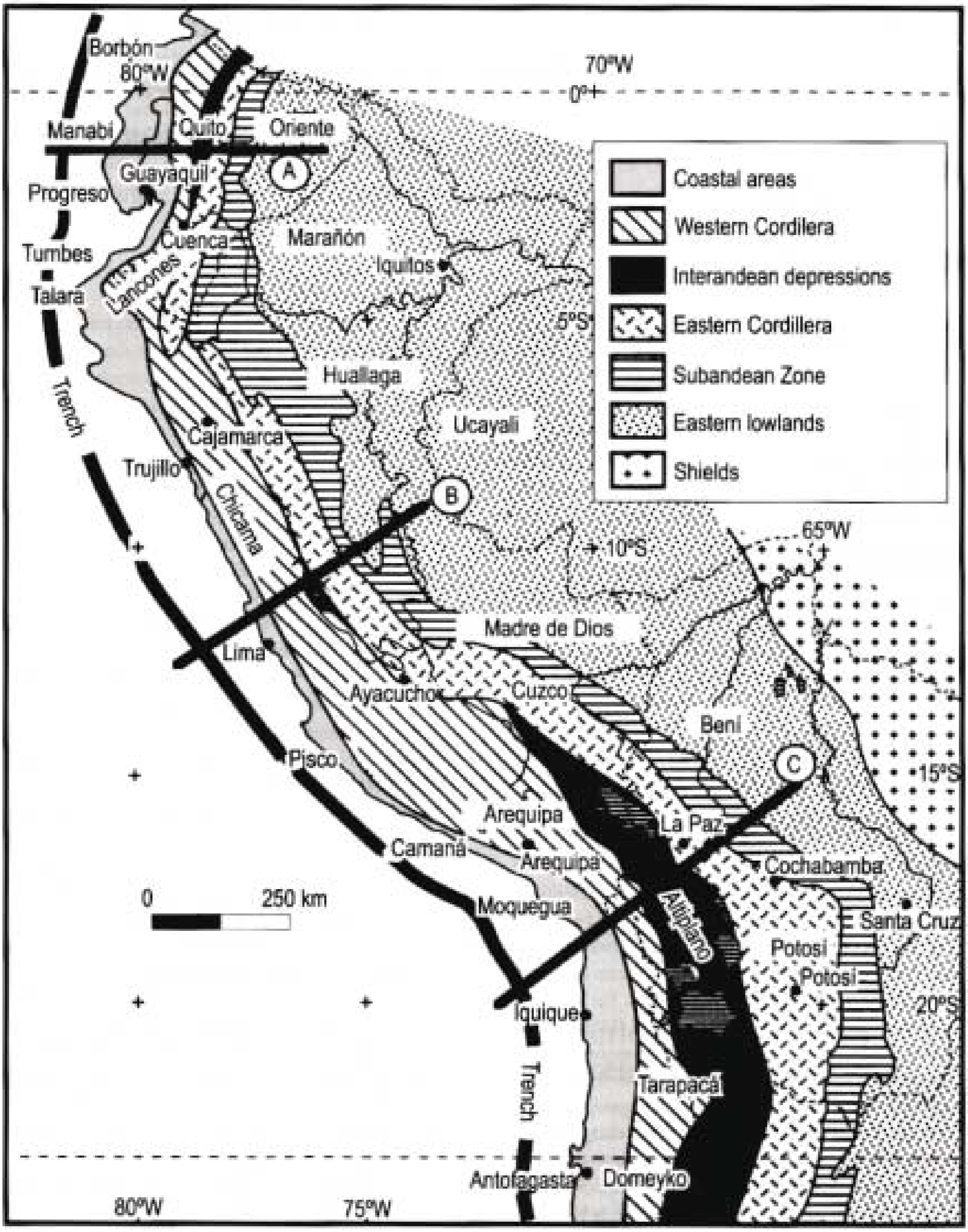
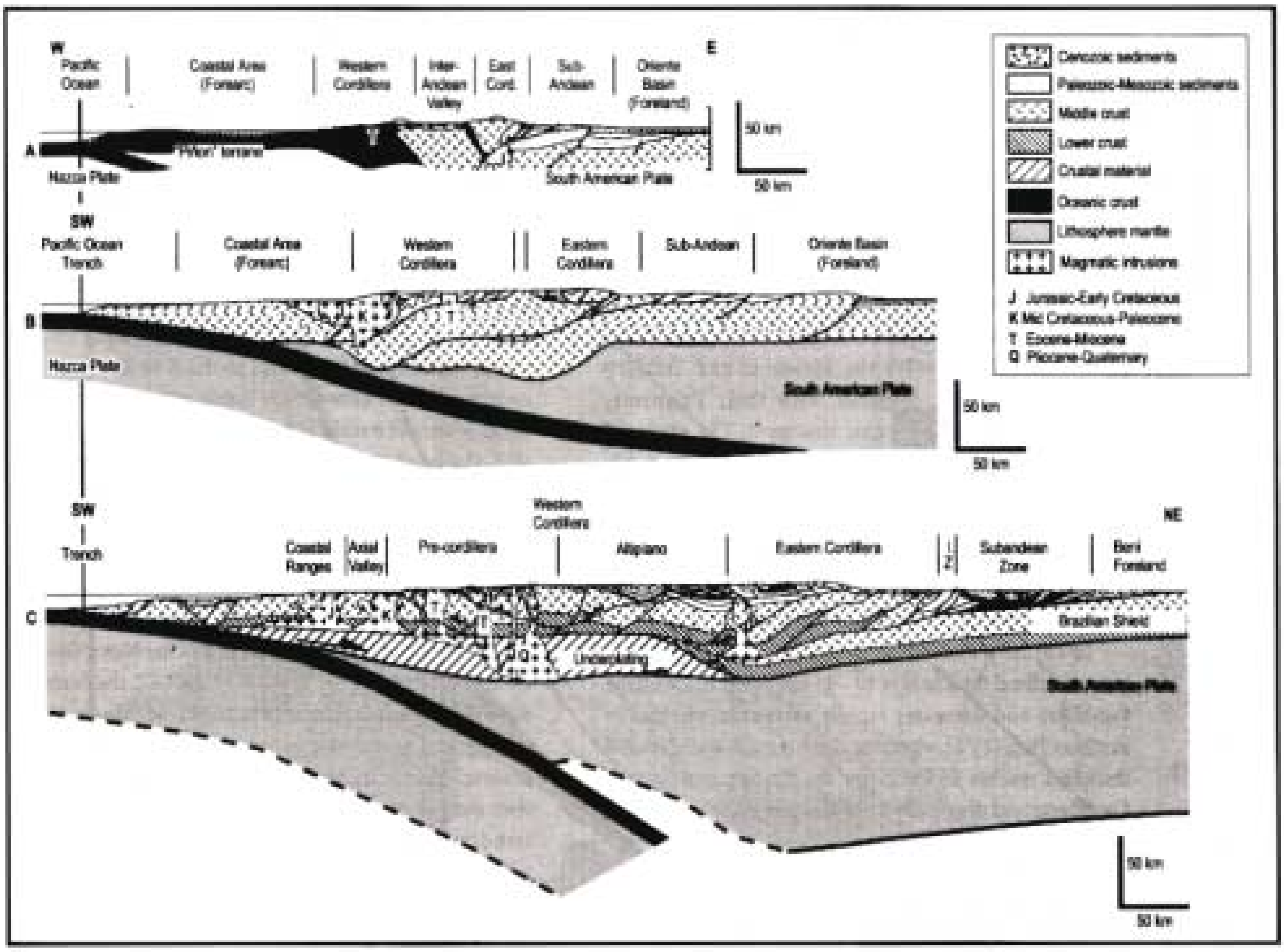


FIGURE 1 - Structural sketch of the Andes of Ecuador, Peru, Bolivia and northern Chile, showing the main morphostructural units. A, B, C is the location of profiles of Fig. 2.

FIGURE 2 - Structural sections of the Andes of A - Ecuador (after Mégard, 1987), B - Central Peru (after Moulin, 1989), and C - Bolivia (after Rochat et al., 1999). Location on Fig. 1.



Morphostructural units and crustal structure

The Ecuadorian segment (1°N - 3°S) exhibits four main longitudinal morpho-structural units (Fig. 1). In the 50 to 180 km wide coastal area relief is low (< 300 m), and sediments are accumulated in rather wide alluvial plains by rivers descending from the Andes. The Andean Chain is made of a Western Cordillera constituted by deformed oceanic rocks, and an Eastern Cordillera (Cordillera Real) made of Paleozoic to Mesozoic metamorphic rocks (Litherland *et al.*, 1994; Fig. 2). The eastern areas are made of a 50 - 80 km wide uplifted zone (500 to 1000 m) made of Mesozoic deformed sedimentary rocks (Subandean Zone), bordered to the E by gently dipping lowlands covered by rainforest, where Quaternary sediments derived from the Andes are presently being deposited.

North of 3°S, the Interandean Valley filled by Tertiary-Quaternary volcanic and volcaniclastic deposits separates the cordilleras. High, active or Recent volcanoes are built on the Cordilleras, the Interandean Valley and the eastern Subandean Zone. South of 3°S, the Interandean Valley, the western oceanic terranes and the active volcanoes disappear, and the Andean Chain is made of metamorphic rocks unconformably overlain by pre-Pliocene volcanic and subordinate sedimentary rocks. The Guayaquil Gulf indentation (3°S) roughly separates the northern fore-arc units (North Andean Block), made of oceanic terranes, from southern fore-arc zones made of continental blocks. In northern Peru (3°S - 5°S), the latter constitutes an isolated NE-SW trending coastal Cordillera (Amotape Massif). This zone corresponds to an important change in the chain orientation (Huancabamba Bend).

Farther S (5°S - 18°S), the NW-SE trending Peruvian margin is basically made of the same units, which exhibit, however, distinct features. Except in northern Peru (Sechura Plain), the coastal zone is narrow and disappears locally in southern Peru. It is made of the deformed edge of the South American Continental Plate, locally covered by fore-arc marine deposits, mainly of Tertiary age. The southern tip of this segment is defined by a sharp direction change from NW to N-S (Arica Bend, Fig. 1).

The Western Cordillera constitutes the continental water divide. It is mainly made of deformed metamorphic rocks and Mesozoic sediments, intruded to the west by Mesozoic plutons (Mesozoic arc) and unconformably covered to the E by abundant volcanic rocks (Tertiary arc), associated with localised Tertiary plutons (Cordillera Blanca). Active volcanoes are only known S of 16°S, thus defining a zone without present-day arc volcanism between 2°30'S and 16°S.

The Eastern Cordillera is chiefly made of pre-Mesozoic metamorphic rocks and subordinate Paleozoic to Tertiary intrusions and volcanic rock (inner arc of southern Peru; Kontak *et al.*, 1985). It is bordered to the W or SW by a nearly continuous E-verging thrust and fold belts (Marañón, Ticlio, Mañazo FTB) involving mainly Mesozoic sedimentary rocks. In southern Peru, the Eastern Cordillera is thrust over the Altiplano by means of SW-verging reverse faults. Longitudinal valleys, the trend of which is controlled by the main Andean structures, separate N of 13°S the Eastern and Western Cordilleras. S of 13°S, the Eastern and Western Cordilleras are more distant and are separated by large valleys (13° - 14°S) evolving southeastwards into an endorheic basin (Altiplano), which considerably enlarges in Bolivia.

The deformed subandean zone is much larger than in Ecuador (120 to 250 km) and enlarges southeastwards (Fig. 2). Deformation involve the Mesozoic and Tertiary sedimentary rocks, together with their Paleozoic, sedimentary or metamorphic basement. The eastward migration of deformation influenced the course of the rivers, which trend mainly parallel to the Andean structures. The eastern lowlands are made of either wide alluvial plains receiving Quaternary sediments (northern Peru), or slightly uplifted, fault-controlled zones underlain by the Brazilian Shield (southern Peru, Bolivia).

In northern Peru (7°S), where the chain is relatively narrow, crustal thickness is 40 - 45 km below the Western Cordillera and decreases rapidly eastwards, whereas in southern Peru (13°S), where the chain is much wider, crustal thickness reaches 65 km below the Western and Eastern Cordilleras and drastically decreases below the Subandean Zone (Fukao *et al.*, 1989).

The northern part of the Chilean-Bolivian segment is more complex. To the W, the 50 km wide Coastal Cordillera is made of chiefly Jurassic-Early Cretaceous magmatic arc rocks, cut by the major N-S trending Atacama wrench fault system. The Longitudinal Valley, underlain by Middle Cretaceous arc rocks separates it from the Precordillera. The 3000 m high Precordillera is made of latest Cretaceous-early Paleogene volcanic and sedimentary rocks overlying Jurassic-Early Cretaceous back-arc sedimentary rocks, and overlain by Neogene-Quaternary tuff and piedmont deposits. To the N of 21°S, it merges with the Western Cordillera, whereas to the S of 21°S, it is separated from the latter by the Preandean Depression occupied by the Salar de Atacama. As a whole, these four desert units are 70 to 250 km wide.

The Western Cordillera is made of Neogene-Quaternary arc rocks, and comprises active volcanoes culminating at more than 6000 m. Isolated outcrops show that Tertiary arc rocks directly overlie Precambrian basement, at least locally. The Western Cordillera forms the Chilean-Bolivian border and limits to the W the 200 km wide, flat Bolivian Altiplano, in-filled by thousands of metres of Tertiary to Quaternary rocks. The latter exhibits an average altitude of more than 3900 m and contains endorheic lakes evolving locally into wide *salars*. The Eastern Cordillera is made of Paleozoic folded metasediments which are thrust to the SW or W onto the Altiplano border, and to the NE or E onto the Subandean Zone. At 19°S, crustal thickness exceeds 60 km and reaches 70 to 75 km below the Western and Eastern Cordilleras (Beck *et al.*, 1996; Fig. 2). The Subandean Zone is made of faulted and folded sedimentary rocks of Paleozoic to Miocene age, which are thrust to the NE or E onto the eastern lowlands. The latter is a large alluvial plain where products of the erosion of the Andes are presently being deposited.

Plate tectonic setting

As first proposed by Wegener early in this century, the Andean Orogeny is related to the convergence between the Pacific Ocean Plate and the South American Continent, the latter being pushed westwards by the opening of the South Atlantic Ocean. We know nowadays that not only South America, but also the Pacific oceanic Plate moves in an absolute reference frame. South America migrates roughly westwards at a rate of 3 cm/y. Between 2°S and 23°S, the Nazca Oceanic Plate migrates to the E to ENE, and the convergence rate between the Nazca and South American plates is around 8 cm/y, although it seems to be lower at 0° (Fig. 3). The whole Andean margin is, therefore, submitted to a roughly E-W convergence between the subducting Nazca Plate and the South American continental margin. This phenomenon is long regarded as the major cause of the Andean Orogeny (Rutland, 1971; James, 1971).

The NE trending Grijalva Fracture Zone (GFZ) located between 3°S and 5°S (Fig. 3) separates the Nazca Oceanic Plate into two portions. To the N of the GFZ, the currently subducting oceanic plate is 23 to 10 Ma old (Miocene) and the depth of the oceanic bottom is 2800 to 3500 m below sea level. Most of the Ecuadorian -Colombian trench is less than 4000 m deep. The Nazca Plate is overlain by the E-W trending Carnegie aseismic ridge, which is presently subducting offshore Ecuador between 0° and 2°S. The Carnegie Ridge is regarded as formed by the Galápagos hotspot, situated about 1000 km W of the Ecuadorian coast.

South of the GFZ, the subducting oceanic plate is 50 to 30 Ma old (Eocene - early Oligocene) and 4000 to 5600 m deep. The Peruvian-Chilean trench reaches depth of 6600 m (11°S, 17°S), 7500 m (19°S) and even 8055 m (23°S). The Nazca Oceanic Plate comprises two submarine highs. The NE-trending Nazca aseismic ridge, entered into the trench 3 to 10 Ma ago and is presently subducted at 15°S - 16°S. It presents summits culminating at -330 m, -841 m and -2059 m, from SW to NE, thus showing a NE plunge of its crestal line. To the S of 18°S, a N-S trending submarine ridge runs alongside the trench of northern Chile, and culminates locally at -1800 m.

Recent improvement in hypocenter relocation at the scale of South America has been done, giving a good definition of the Benioff-Wadati Zone geometry (Engdahl *et al.*, 1998). It shows along-strike segmentation of the Andes but without tears even around the latitude 15°S where the slab has a severe contortion due to changes in its dip. And only a very small portion of the observed seismicity originates within the Andean crust.

From N to S, the Nazca Plate subducts at an angle of 19° - 30° to a depth which varies from 250 km to 650 km. N and S of the Bolivian Orocline, at latitudes 2°S - 15°S and 26°S - 33°S, seismic activity indicates that the Nazca Plate is being subducted nearly horizontal at a depth of about 100 km for a distance of over 300 km before descending steeply into the mantle. There are no volcanic arcs because of the lack of asthenosphere wedge between the subducting flat slab and the upper continental plate (Barazangi and Isacks, 1976). These horizontal segments partially correlate with the oceanic ridges (Nazca Ridge, Juan Fernandez Ridge) or totally

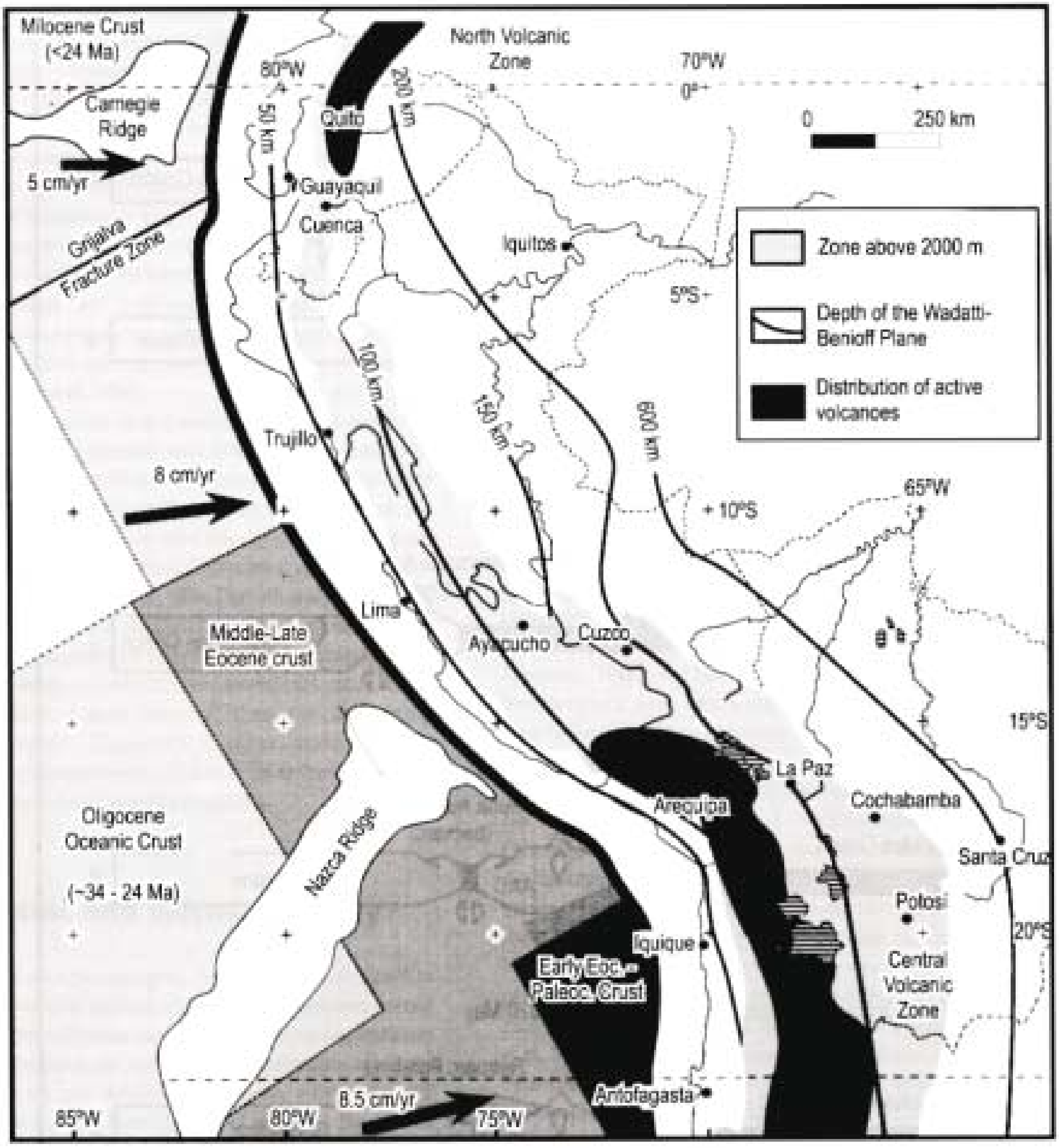


FIGURE 3 - Plate tectonic setting of the Andes of Ecuador, Peru, Bolivia and northern Chile.

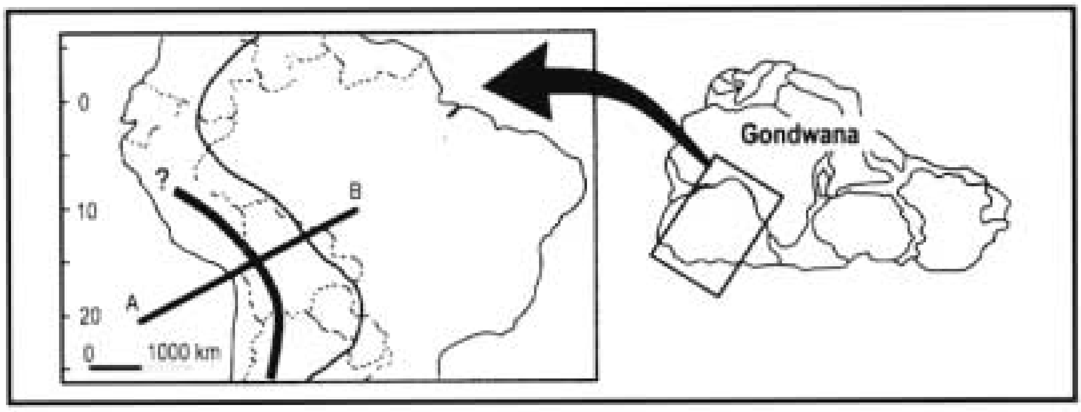


FIGURE 4 - Location of the Central Andes within the Gondwana. Thick grey line - boundary zone between the Arequipa-Antofalla Craton and the Amazonian Craton. Dashed line - eastern boundary of the inferred maximum extent of Paleozoic sedimentation in the "Peru-Bolivia" backarc/retroarc basin. A-B - location of the schematic geodynamic cross-sections of Fig. 5. (all are approximate).

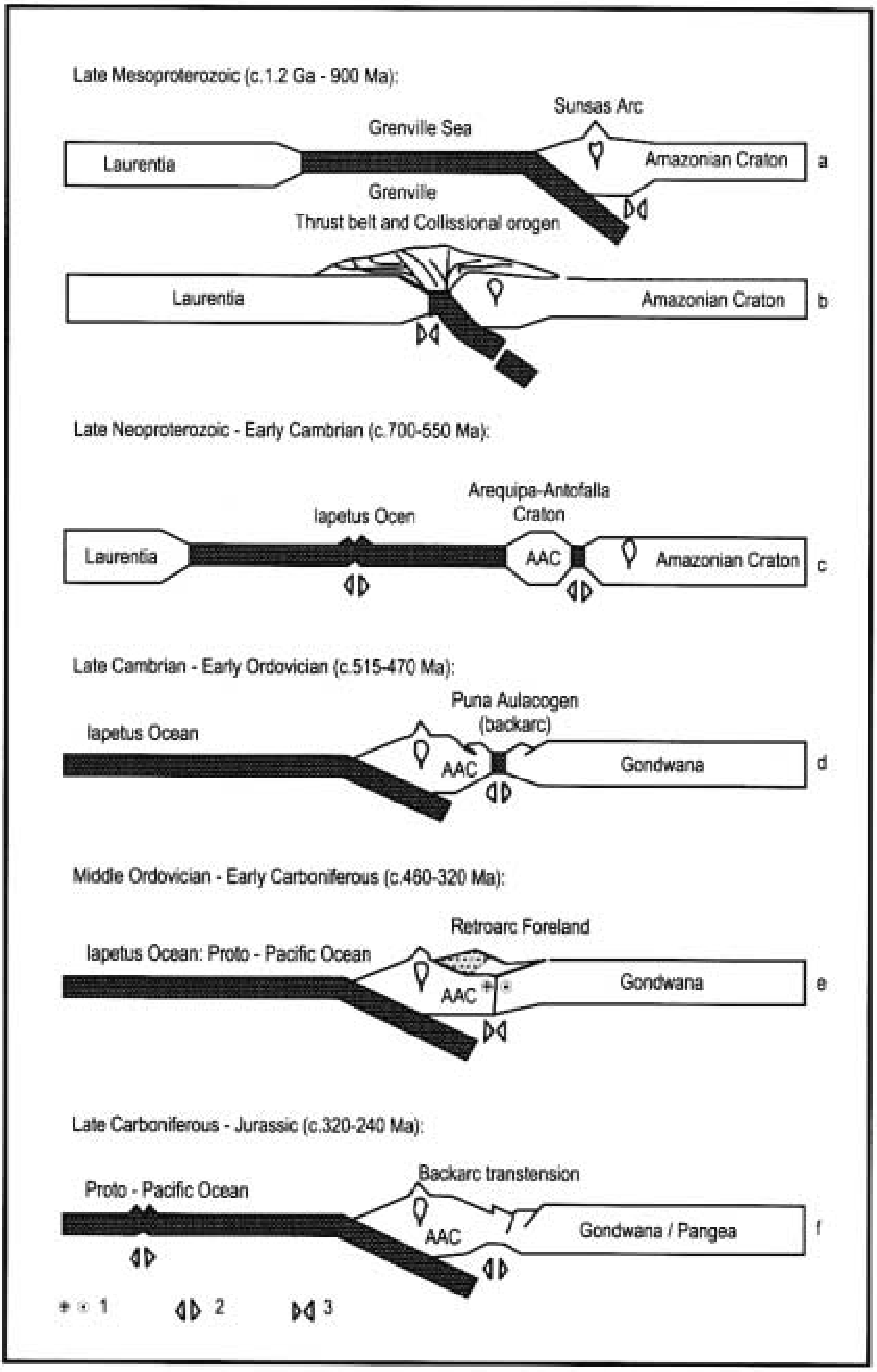


FIGURE 5 - Simplified conceptual model for the pre-Andean geodynamic evolution of the Central Andes (5° - 27°S) as proposed in the text. Approximate location of cross sections in Fig. 4. Overall regional stress field: 1, transcurrent; 2, extensional; 3, compressional.

(Inca Plateau?) subducted (Gutsher *et al.*, 1999). In addition, it is accompanied by important shallow earthquakes (< 20 depth) related to crustal shortening.

The contact zone between the Nazca and South American plates concentrates only a part of the convergence motion as shown by GPS measurements (Kendrick *et al.*, 1999). In the case of the Antofagasta 1995 earthquake, the last great Chilean event ($M_w = 8.1$), a locked portion of the plate interface was ruptured. The observed slip release (between 2 and 7 m) was less than the expected plate motion accumulation (over 8 m) and suggests that the resultant motion causes permanent deformation through uplift of the Altiplano and crustal shortening in eastern Andes (Norabuena *et al.*, 1998).

Low magnitude earthquakes recorded by local seismic arrays are revealing small-scale differences in seismicity of the Benioff-Wadati Zone and the continental crust (Comte *et al.*, 1999), and should be related with some specific tectonic processes. Body-wave velocity models at the scale of the South American Plate (Bijwaard *et al.*, 1998) and at smaller scale (Comte *et al.*, 1994; Dorbath *et al.*, 1996; Myers *et al.*, 1998) show a medium lateral heterogeneity where the descending oceanic Nazca Plate can be observed over 600 km depth as zones of faster seismic velocities due to the colder, denser material than the underlying asthenosphere. The crustal thickness underneath the Altiplano Plateau is of 60 - 75 km, twice as thick as normal, but its exact origin remains enigmatic.

Volcanic and seismic activity

Like all active margins, the Andes are submitted to intense volcanic activity, which presents, however, several peculiarities. Distribution of the active volcanoes indicates that the volcanic arc activity is not continuous along the Andean margin, defining a Northern, a Central and a Southern Volcanic Zone (NVZ, CVZ and SVZ; Fig. 3), that have been interpreted as related to the dip of the subduction plane. As a matter of fact, the Northern and Central Volcanic Zones overlie segments with relatively steep Wadatti-Benioff Zones, whereas the volcanic gap of southern Ecuador and northern-central Peru corresponds to a zone of flat subduction. In this latter case, the lack of an asthenosphere wedge between the subducting slab and the upper continental plate would prevent arc magma generation (Barazangi and Isacks, 1976). Moreover, the NVZ is characterized by basaltic andesite, whereas andesite, dacite lava and ignimbrite dominate the CVZ. The latter feature is interpreted as due to fractional crystallization during the ascent of the andesite magma through a thickened continental crust.

In addition, this part of the world has already generated great earthquakes of magnitude over 8.2 followed in general by destructive tsunamis. They occurred repeatedly with a certain periodicity of more than 100 years as the greatest instrumentally ever-recorded earthquake located in southern Chile (37°S - 46.5°S). This earthquake occurred on May, 22 1960 ($M_w = 9.4$, depth = 25 km) and was generated by a 20m slip upon an area of about 1000 x 200 km along the contact zone between the two plates. Over magnitude 8.7, the southern Peru on August 14, 1868 ($M = 9.0$), the northern Chile on May 9, 1877 ($M_w = 9.0$) and the Ecuador-Colombia on January 31, 1906 ($M_w = 8.8$) great earthquakes and their associated tsunamis caused a lot of damage. On the other hand, on June 9, 1994 the largest deep earthquake ($M_w = 8.2$, depth = 630 km) in recorded history occurred beneath the Andes providing more constraints on the dynamics and geometry of the Nazca Plate at that depth.

Pre-Andean Evolution of the North-Central Andes Domain

This section attempts to provide a coherent overview of the paleogeographic and geodynamic evolution of the Central Andes Domain during the Paleozoic, along with *some comments on its Proterozoic basement. The synthesis* has been compiled from different sources, and also includes some new concepts and interpretations. The area here considered as Central Andes is roughly 5°-27°S, covering part of Peru, Bolivia, northern Chile and northwestern Argentina (Fig. 4), and therefore excludes the Amotapes of northwestern Peru and southern Ecuador, and the Precordillera of western Argentina.

In contrast with the rather complex Phanerozoic accretionary history of the South American margin in the Northern and Southern Andes (Colombia-Ecuador and Chile-Argentina, respectively), the evolution of the Central Andes appears to be somewhat simpler, as there is no evidence for the accretion of allochthonous terranes during the Phanerozoic. Recent studies suggest that the crustal basement in most of the Central Andean area formed part of the Grenville Orogen, as a result of the collision between Laurentia and Amazonia in the Mesoproterozoic (Wasteneys *et al.*, 1995; Sadowsky and Bettencourt, 1996; Tosdal, 1996). Paleozoic rocks of the Central Andes record the break-up of the Neoproterozoic Proto-Pangea (Rodinia) in the latest Neoproterozoic-Cambrian to form a passive margin along western Gondwana (Bond *et al.*, 1984; Powell *et al.*, 1993), and its later evolution as an active margin during most of the Paleozoic and until present times (Sempere, 1995). The continuous superposition of magmatic, tectonic and sedimentary events has led to complex lateral variations, both in cross section and along strike, and has originated a wide range of settings for the development of mineral deposits (Fontboté *et al.*, 1990; Schneider, 1990; Fornari and Hérail, 1991) and hydrocarbon generation and accumulation (Gohrbandt, 1992; Moretti *et al.*, 1995; Tankard *et al.*, 1995) related with Paleozoic rocks.

Proterozoic basement

Due to the protracted superposition of orogenic events in the Central Andes, the interpretation of the Proterozoic evolution of the region is inevitably very fragmentary, and should be considered as a mere working hypothesis. The modern basement of the Central Andes consists of two crustal blocks with different origins: the Arequipa-Antofalla Craton

(Ramos *et al.*, 1986) and the Amazonian Craton (Teixeira *et al.*, 1989) or Central Brazil Shield (figs. 1 and 2). The Arequipa-Antofalla Craton is a Proterozoic terrane interpreted to have originated as the tip of a pre-Grenville Laurentian promontory (comprising Labrador, Greenland and Scotland) that was incorporated into the Grenville Orogen (Dalziel, 1994; Wasteneys *et al.*, 1995). Pb isotope composition seems to contradict this model, indicating instead closer ties with the Amazonian Craton (Tosdal, 1996). The reconstruction of the remnants of the Grenville Orogen in South America (Sadowski and Bettencourt, 1996) indicates that the Central Andes corresponds to an area intermediate between the magmatic arc (represented by the Sunsas igneous province, in eastern Bolivia and western Brazil) and the thrust belt (southeastern Canada) of the Grenville Orogen (Fig. 5b), and explains the similar trends identified between the Proterozoic outcrops along the Andes, and those of the Brazilian Shield (Litherland *et al.*, 1985, 1989).

Paleoproterozoic ages indicated by Rb/Sr whole-rock isochrons and bulk U/Pb zircon geochronology represent the pre-Grenville Laurentian-Amazonian protolith, and Mesoproterozoic ages of granulite-facies metamorphism indicated by U/Pb single-grain zircon geochronology represent the main collisional events of the Grenville Orogen (Wasteneys *et al.*, 1995; Sadowski and Bettencourt, 1996; Tosdal, 1996). Rifting during break-up of Rodinia in the Neoproterozoic-Cambrian led to separation of Laurentia from Amazonia (Fig. 5-c), leaving behind the parautochthonous Arequipa-Antofalla Craton attached to Amazonia (Central Brazil Shield). The boundary zone between the two crustal blocks, which is located beneath the Eastern Altiplano and Eastern Cordillera, thus constitutes a paleosuture and crustal weakness zone inherited from the Mesoproterozoic evolution of the Grenville Orogen (Figs. 1 and 2). This zone remained active during the Paleozoic, and ever since, with variable behaviour depending on the regional state of stresses (Ramos, 1988; Dorbath *et al.*, 1993; Forsythe *et al.*, 1993).

Tectonomagmatic episodes

Tectonic and magmatic events took place in the Central Andes in a rather continuous manner during the whole Paleozoic, shifting their foci and areal extent with time as a result of plate interactions, variable geometry of the plates involved, and the resulting stress regimes. These variable conditions led to apparently different styles of evolution depending on the local area under study. However, an overall tendency for the whole region may be discerned and simplified as follows. Apart from the aforementioned Mesoproterozoic (1.2 Ga - 900 Ma) ages resulting from the Grenville granulite-facies metamorphism, and the Paleoproterozoic (2.0 -1.9 Ga) ages obtained from its metagranitoid protolith, the crystalline basement underlying the Central Andes also presents Neoproterozoic to Middle Cambrian (600 - 520 Ma) ages of igneous and metamorphic events. These events are commonly assigned to the last phases of the Brasiliano Orogeny, which are referred to as Pampean Orogeny in the southern Central Andes (Rapela *et al.*, 1998).

Break-up of Rodinia and rifting of eastern Laurentia from Gondwana in the Neoproterozoic and Early Cambrian led to the development of passive margins on both sides of the Southern Iapetus Ocean (Fig. 5-c).

Along the pre-Andean margin of Gondwana S of 27°S, a westward shift of the spreading centre is interpreted to have left oceanic crust between a detached continental block (Pampean Terrane) and the Gondwana margin (Rapela *et al.*, 1998). Later eastward subduction and closure of this remnant sea during the Early Cambrian led to the collision of the Pampean Terrane in the Middle Cambrian. To the N of 27°S, a similar history is also probable, with the Arequipa-Antofalla Craton also partially rifting and then later colliding along its southeastern boundary with the Amazonian Craton. Meanwhile, the western margin ofboth the Pampean Terrane and the Arequipa-Antofalla Craton remained passive until the Late Cambrian (Fig. 5-c).

Beginning in the Late Cambrian or earliest Ordovician, this western passive margin became an active continental margin (Fig. 5-d). The San Nicolás Batholith in southwestern Peru is interpreted as the root of the magmatic arc resulting from eastward subduction of oceanic crust along the active margin of Gondwana during the Paleozoic (Mukasa and Henry, 1990). Ordovician-Devonian ages obtained for lower intercepts in U/Pb geochronology of basement rocks along the western Arequipa-Antofalla Craton reflect thermal overprinting and Pb-loss coinciding with peaks of this Paleozoic magmatic activity (Shackleton *et al.*, 1979; Damm *et al.*, 1990; 1994; Mukasa and Henry, 1990; Tosdal, 1996). Different rates of plate activity and sense of migration of the magmatic arc developed depending on regional stresses and inhomogeneities, and basin development also changed in accordance with these plate interactions.

The Paleozoic development of this active margin is characterized by back arc extensional conditions during the early phase (latest Cambrian-Early Ordovician) and late phase (Late Carboniferous-Permian), resulting in a strongly subsiding thinned crust, with partial rifting and syn-sedimentary basic volcanism (Fig. 5d). In contrast, a compressional regime (retro-arc foreland) characterized the intermediate phase (Middle Ordovician-Early Carboniferous; Fig. 5e). An apparent lack of evidence for Silurian and Devonian tectonomagmatic activity in the southwestern Central Andes has been interpreted as evidence for a passive margin resulting from rifting off of part of the Arequipa-Antofalla Craton (Bahlburg and Hervé, 1997). However, this interpretation is difficult to reconcile with the evidence for a coeval active plate margin in southern Peru (down to 17°S) and northern Chile and Argentina (up to 27°S), as well as with the Silurian age of igneous and metamorphic events in the same region (Damm *et al.*, 1990, 1991, 1994).

With regard to Late Paleozoic tectonomagmatic events in the Central Andes, these have been traditionally assigned to a "Eohercynian" Orogeny (Mégard *et al.*, 1971; Bard *et al.*, 1974; Dalmayrac *et al.*, 1980), including Late Devonian-Carboniferous K/Ar and Rb/Sr ages, and regional stratigraphic and petrographic evidence. However, only one U/Pb zircon age is reported (330 ± 10 Ma; Carlier *et al.*, 1982). Other U/Pb zircon dates on granitoid along the NW trending segment of the Eastern Cordillera of Peru and

Bolivia establish their time of emplacement as Permian or younger (McBride *et al.*, 1983; Heinrich *et al.*, 1988; Farrar *et al.*, 1990; Kontak *et al.*, 1990).

This more recent evidence questions the relation of the granitoid plutons with widespread Late Devonian-Early Carboniferous "Eohercynian" magmatism and orogenesis, as previously interpreted. Nevertheless, there is evidence for local transpressional uplift and deformation of the Eastern Cordillera in the latest Devonian and Early Carboniferous, which separated an Altiplano basin from a Subandean-Chaco basin (Mégard *et al.*, 1971; Dalmayrac *et al.*, 1980; Kley and Reinhardt, 1994; Sempere, 1995; Díaz-Martínez, 1996). Maximum burial depths (locally exceeding 10 km) were attained in different areas of the Central Andean Paleozoic basin at different times during the Late Paleozoic (Late Devonian-Permian interval). As a result of this deep burial, and probably in conjunction with transcurrent stresses along the aforementioned suture zone (Sempere, 1995), the stratigraphically lower units (Ordovician-Silurian) reached very low-grade to low-grade metamorphism, as indicated by vitrinite reflectance and illite crystallinity (Kley and Reinhardt, 1994). This thermal overprint resulted in the reset of K/Ar system, thus explaining the Carboniferous-Early Permian ages obtained by some authors (Dalmayrac *et al.*, 1980; Paton, 1990).

At the same time, erosion related with the mid-Carboniferous global regression resulted in a disconformity or low-angle unconformity throughout the Central Andes, with Upper Carboniferous and Permian units directly overlying Devonian, Silurian or Ordovician units (Kley and Reinhardt, 1994; Isaacson and Díaz-Martínez, 1995; Díaz-Martínez, 1996, 1998b). A similar unconformity is observed in the southern Central Andes (northern Chile and northwestern Argentina), where the development of an active margin with related fore-arc, intra-arc and back-arc basins took place during the Late Carboniferous and Permian (Breitkreuz *et al.*, 1988, 1989; Bahlburg and Breitkreuz, 1991; Breitkreuz and Zeil, 1994). Thus, the reinterpretation of the evidence indicates that the alleged "Eohercynian" Orogeny may be no more than the conjunction of different processes and events taking place during the Late Paleozoic throughout the Central Andean region, but not a single tectonomagmatic event or belt localized in space and time.

The sedimentary record in Peru and western Bolivia presents evidence of marginal arc volcanism beginning in the late Early Carboniferous (Ambo Group, 6°S - 17°S; Dalmayrac *et al.*, 1980; Díaz-Martínez, 1995, 1998b). This subduction-related magmatic activity propagated to the S (16°S - 24°S) along the active margin during the Late Carboniferous and into the Permian (Carlier *et al.*, 1982; Bell, 1987; Breitkreuz *et al.*, 1988, 1989; Bahlburg and Breitkreuz, 1991; Breitkreuz and Zeil, 1994; Sempere, 1995; Isaacson and Díaz-Martínez, 1995; Bahlburg and Hervé, 1997). Farther to the S (20°S - 42°S), thick volcano-sedimentary successions developed in the Permian and Triassic (Choiyoi Group), with associated high-level intrusions (Kay *et al.*, 1989).

Further evidence for Late Paleozoic magmatic activity is also found in the Eastern Cordillera of Peru and Bolivia. This activity began earlier in the Peruvian sector, with Late Permian-Triassic ages (Kontak *et al.*, 1990), and migrated towards the S, developing in Bolivia during the Middle Triassic to Early Jurassic (McBride *et al.*, 1983; Heinrich *et al.*, 1988). The activity was not related to subduction processes, but instead, consists of alkaline volcanism and plutonism which are interpreted as a result of partial rifting and transtension along the suture zone between the Arequipa-Antofalla Craton and the Amazonian Craton (Fig. 5f), due to regional stresses during Pangea break-up (Kontak *et al.*, 1990; Atherton and Petford, 1991).

Tectono-sedimentary cycles

Most of the Late Cambrian to Devonian sedimentation in the Central Andes took place along a wide and elongated epicontinental marine basin broadly parallel to the margin of Gondwana (Fig. 4). The main source area for this basin was situated to the W and SW, although this source area was not a rifted-off former "Pacific continent" (as considered by Bahlburg and Hervé, 1997). Instead, it is considered that the source area was the active margin of Gondwana itself (Isaacson and Díaz-Martínez, 1995; Sempere, 1995), influenced by its discontinuous activity, specially the uplift and forward migration of the fold-thrust belt. Deepening of the basin and increased subsidence rates broadly coincide in time with the aforementioned Ordovician-Devonian peaks of magmatic activity, and demonstrate the syn-tectonic character of deposition, in close relation with tectonic piling and uplift along this fold-thrust belt (Sempere, 1995; Díaz-Martínez *et al.*, 1996).

The Paleozoic marginal orogen which fed the basin is today completely dismantled, with its roots cropping out in the Arequipa Massif and other smaller outcrops scattered throughout the Arequipa-Antofalla Craton. Extreme burial depths locally exceeding 10 km were achieved during the Late Paleozoic, leading to the reset of isotope ages during their maxima, and previous to the widespread erosion of depocenter areas. Paleozoic basin fills were later variably eroded, and the resulting stratigraphic gaps are wider at the margins of the basin, especially towards the western uplifted areas. The original width of the Paleozoic orogen seems narrower today due to tectonic erosion along the western margin of the Arequipa-Antofalla Craton (Stern, 1991), as well as compressional shortening due to Cenozoic folding and thrust-piling. Both processes (tectonic erosion and dismantling of the orogen and basin fill) continued during Mesozoic and Cenozoic subduction, uplift and erosion in the Central Andes, and resulted in the scarcity of evidence for the more marginal (fore arc and intra arc) and older basins.

Paleozoic sedimentation in the Central Andes may be subdivided into three main tectono-sedimentary cycles (limited by major unconformities): latest Proterozoic-Middle Cambrian, Late Cambrian-Early Carboniferous, and Late Carboniferous-Early Triassic. The first tectono-sedimentary cycle equates to the Pampean Cycle of Aceñolaza *et al.* (1990). The second tectono-sedimentary cycle equates to the Tacsarian and Cordilleran cycles of Suárez-Soruco (1989), and to the Tacsara, Chuquisaca and Villamontes supersequences of Sempere (1995). The third tectono-sedimentary cycle equates to the Cuevo supersequence of Sempere (1995). Overall, Paleozoic sedimentation was rather continuous, although remarkable changes in subsidence rates took place during certain time intervals. Sedimentary evidence for increased tectonic instability and uplift occur near the Ordovician-Silurian and Devonian-Carboniferous boundaries (Díaz-Martínez *et al.*, 1996), which have been traditionally used for further subdivision of the sedimentary pile (Suárez-Soruco, 1989; Gohrbandt, 1992; Sempere, 1995). However, as mentioned above, tectonomagmatic processes along the margin have been rather continuous and synchronous with the development of basins. Hence, it does not seem appropriate to maintain the traditional subdivisions of the Paleozoic sedimentary record based on discrete orogenic events.

Paleogeography and paleoclimates

Deposition within a large marine basin (Peru-Bolivia Basin) located adjacent to a marginal orogen prevailed during most of the Paleozoic (Figs. 1 and 2). The overall and progressive increase of rigidity and thickness of the crust beneath the Central Andes in the Paleozoic, led to a gradual increase of terrestrial (subaerial and lacustrine) sedimentation in the Late Paleozoic. Fluvial deposits are very rare until the Devonian, and become frequent in the Carboniferous. Eolian deposits begin to be present in the Late Carboniferous and are frequent until the Jurassic. The areal extent of marine deposition in the Late Permian and Triassic was very limited.

The Central Andes underwent important latitudinal movements during its drift as part of Gondwana's western margin in the Paleozoic. Despite the scarcity of confident paleomagnetic data, several models have been proposed for the Paleozoic paleogeographic evolution of the region. The overall trend consists of a subtropical (30°) latitudinal position during the Early Cambrian, and a shift towards higher latitudes in the Middle and Late Cambrian (Courjault-Radé *et al.*, 1992). During the Ordovician, Silurian and Devonian, the Central Andes remained at mid to high latitudes, with variable shifts. A gradual shift towards lower latitudes took place during the Early Carboniferous, and the area has remained at tropical latitudes ever since the Late Carboniferous (Díaz-Martínez *et al.*, 1993; Isaacson and Díaz-Martínez, 1995; Sempere, 1995; López-Gamundí and Breitkreuz, 1997). Paleozoic tropical carbonates and evaporites in the Central Andes are present only within Early-Middle Cambrian deposits and Late Carboniferous-Permian deposits. Thin carbonate-bearing units are present in the Ordovician, Silurian and Devonian systems containing cool-water faunal associations. Plant remains are frequent in the sedimentary record beginning in the Middle Silurian, and coal development is observed in the late Early Carboniferous (Ambo Group) and in the Late Carboniferous-Early Permian (Copacabana Formation). Glacially influenced deposits are preserved within latest Ordovician-Early Silurian units, and within Late Devonian-Early Carboniferous units. However, true tillites and glacially striated pavements are only very rarely found (Starck, 1995), and most of the evidence consists of glacially faceted and striated clasts, both as dropstones and within resedimented units (Díaz-Martínez *et al.*, 1999). These glacially-related deposits are interpreted as a result of glaciation of tectonically uplifted highlands, and therefore unrelated with the large continental icecaps in Gondwana.

Inherited pre-Andean structures

Structures inherited from the pre-Andean geodynamic evolution of the Central Andes have exerted an important influence on its later development. The crust beneath the Central Andes originally formed as part of Proterozoic orogens. The trend of the structures developed during the formation of these orogens greatly conditioned later Paleozoic basin formation and crustal weakness zones, and therefore influenced the geometry and distribution of tectonism and deposition. The boundary zone between the Arequipa-Antofalla Craton and the Amazonian Craton is the principal of these features (Fig. 4), probably inherited as a suture zone from the collision between Laurentia and the Amazonian Craton as part of the Grenville Orogen (Fig. 5). This crustal weakness zone was the location of (a) rifting in the Neoproterozoic and Cambrian, (b) back arc and foreland successor basin formation from the Ordovician to Carboniferous, (c) syn-sedimentary magmatism in the Ordovician and Silurian, (d) transpressive stresses originating local uplifts in the Devonian and Early Carboniferous, and (e) syn-sedimentary magmatism, and transtensional stresses originating semigrabens and grabens in the Late Carboniferous to Jurassic. In turn, the resulting Paleozoic features also influenced later events. For instance, (a) Paleozoic basin geometry and facies distribution greatly conditioned the location of decollement levels and lateral variations in the propagation of thrusts during Cenozoic tectonism (Baby *et al.*, 1989, 1992; Sempere *et al.*, 1991), and (b) listric faults originated during Late Paleozoic-Early Mesozoic extensional conditions were reversed during the Cenozoic. The same boundary zone was active with magmatism and strong tectonism during the Cenozoic, resulting in the formation of the Eastern Cordillera and Subandean fold-thrust belts.

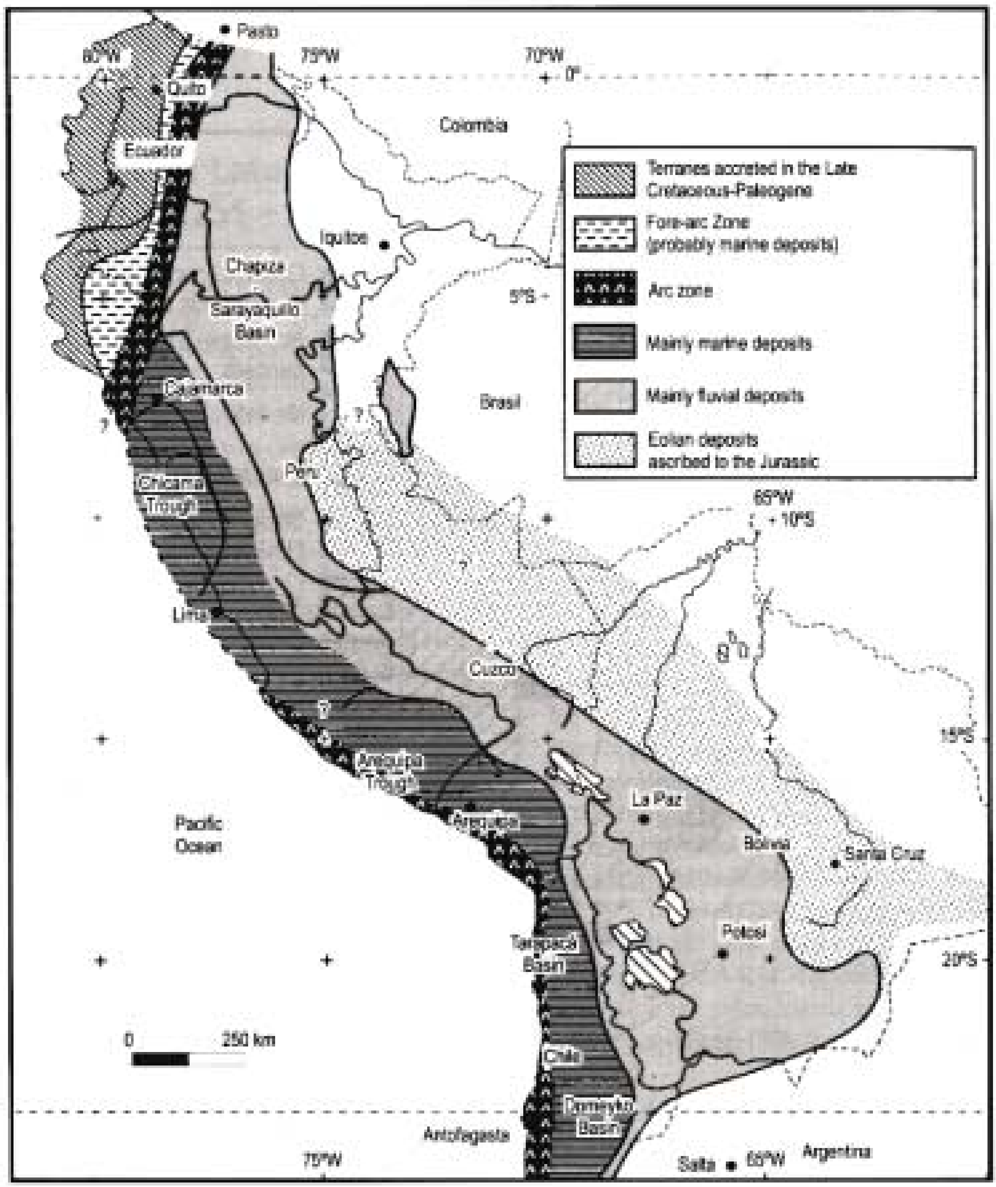


FIGURE 6 - Paleogeographic framework of the Andes of Ecuador, Peru, Bolivia and northern Chile for the Jurassic. (not all features showed are of the same age).

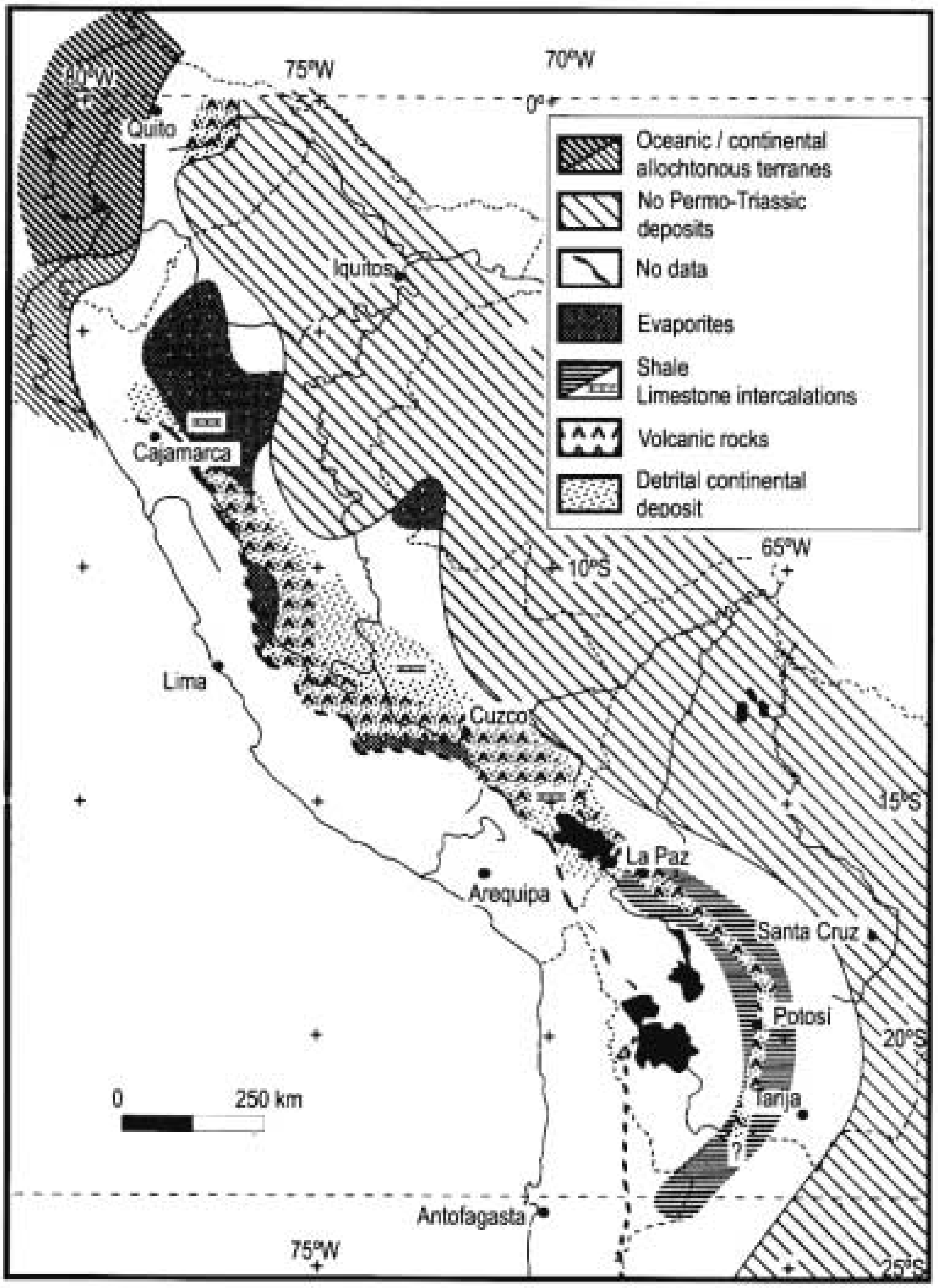


FIGURE 7 - Paleogeographic sketch of Ecuador, Peru, Bolivia and northern Chile for the Late Permian-Triassic period (after Dalmayrac et al., 1980).

Pre-Orogenic (Late Permian-middle Oligocene) Evolution of the North-Central Andes

This period may be divided into two main sub-periods, of Late Permian-Late Jurassic and latest Jurassic-Paleogene age, corresponding to different geodynamic, paleogeographic and tectonic settings.

Late Permian to Late Kimmeridgian (255 - 140 Ma)

Prior to the Sinemurian, the magmatic arc is not well-defined. During Jurassic, the paleogeographic framework exhibits the classical three-fold division of an active margin (Fig. 6). The fore-arc realm, is nearly unknown, because of subsequent tectonic erosion and/or deformation. It was located in the present-day Eastern Cordillera of Ecuador (Cordillera Real), whereas it probably lies offshore Peru and Chile. The magmatic arc controlled the paleogeographic pattern. It trends NNE in the northern segment; runs along the present-day Subandean Zone of Ecuador and crosscut obliquely the northern Peruvian margin. Farther S, it appears locally on the SE trending coast of southern Peru, and extends into northern Chile, where its present-day orientation is roughly N-S. The back-arc domain covers what are now the Oriente Basin of Ecuador, the Western Cordillera of Peru and most of the coast of northernmost Chile. Distal back-arc areas covering eastern Peru and Bolivia (Fig. 6) bordered it to the E.

Late Permian - Middle Sinemurian (255 - 195 Ma)

This period is characterized by Late Permian-Triassic extensional tectonics along NW-SE trending grabens, and by Late Triassic-Early Liassic marine transgressions. Coeval alkaline volcanism and intrusions due to partial melting of the lower crust were locally associated with deep-seated metamorphism.

Late Permian - Early Norian (255 - 215 Ma)

Very little is known about the westernmost areas of the margin. Coastal areas are, however, affected by significant although poorly understood deformation and metamorphism. In the Eastern Cordillera of Ecuador, type I granitoid (Tres Lagunas) yielded ages ranging from 257 (Sm/Nd) to 200 - 189 Ma (Rb/Sr, Aspden and Litherland, 1992; Litherland *et al.*, 1994), and in southwestern Ecuador orthogneiss yielded metamorphic ages of 234 to 198 Ma (K/Ar, Feininger and Silberman, 1982; Rb/Sr, Aspden *et al.*, 1995). Intrusion and metamorphism are interpreted as due to a strong thermal event of Late Triassic age, resulting in partial crustal melting and high temperature metamorphism. They would be associated with significant dextral movements related to the Tethyan oblique rifting (Litherland *et al.*,, 1994).

In northern and central Peru, no pre-Late Jurassic arc volcanic rocks are known. In the coast of southern Peru (Arequipa Massif), metamorphic and intrusive rocks yielded K/Ar ages ranging from 213 to 187 Ma (Stewart *et al.*, 1974; Romeuf *et al.*, 1993). In southern Peru, a Triassic-Liassic age is assumed by Boily *et al.* (1984) for the arc volcanism, since it is post-dated by, and locally intercalated with, Sinemurian marine deposits (Vargas, 1970; Jaillard *et al.*, 1990). However, one of the plutons intruding these volcanic rocks yielded a 211 Ma age, thus indicating a pre-Liassic minimum age (Romeuf *et al.*, 1993).

In northern Chile (26 - 31°S), basic to granitic, I-type intrusions yielded ages between 231 and 201 Ma (Berg and Baumann, 1985; Irwin *et al.*, 1988; Gana, 1991), and are roughly coeval with syn-metamorphic deformation dated at 220 - 201 Ma (Irwin *et al.*, 1987). Farther to the E, in Peru, thick sequences of fine- to coarse-grained, red sandstone and conglomerate beds (Mitu Group) were deposited in a subaerial environment, and filled fault controlled grabens (Mégard, 1978; Kontak *et al.*, 1985; Carlotto, 1998). The associated volcanic rocks are alkaline basalt with subordinate tholeiitic basalt, and acidic, dacitic to rhyolitic pyroclastites (Fig. 7). Shoshonitic and peralkaline suites (Kontak *et al.*, 1985) as well as comendite (Noble *et al.*, 1978) are present, indicating an intraplate extensional tectonic regime. Because they overlie fusulinid-bearing limestone of Early Permian age, the volcanic and sedimentary rocks of the Mitu Group have been ascribed to the Late Permian-Triassic (Newell *et al.*, 1953; Laubacher, 1978). Farther to the N (Ecuador), this period corresponds to non-marine clastics and subordinate volcanics (Sacha Formation, Rivadeneira and Sánchez, 1991). Deposits similar to those of the Mitu Group have been recently identified in Bolivia, where they are found in major paleogeographic structures, which governed the Andean tectonics (Sempere *et al.*, 1999; Fig. 7). In northern Chile, deep transtensional rift basins were filled by thick sequences of subaerial basic, intermediate and acid volcanic rocks and subaerial siliciclastic and volcaniclastic rocks of Late Permian-Triassic age (Chong, 1976; Suarez and Bell, 1991; Flint *et al.*, 1993).

The Late Permian-Early Triassic age has been confirmed in southern Peru by 270 - 200 Ma ages obtained from intrusions and shoshonite, which, however, extent into the Early Jurassic (200 - 180 Ma, Kontak *et al.*, 1985; Clark *et al.*, 1990). In southern Peru, coeval biotite granodiorite and monzogranite derived from partial melting of the lower crust (Granitoid province) are crosscut by mafic dykes showing chemistry and mineralogy similar to that of the alkali basalt of the Mitu Group (Kontak *et al.*, 1985). Farther the SE, most of the granitic to granodioritic, peraluminous intrusions of the Bolivian Cordillera Real yielded ages from 225 to 195 Ma (Ávila-Salinas, 1990). These sedimentary, tectonic and magmatic phenomena were interpreted as resulting from a NW-SE trending rifting of Late Permian-Triassic age, related to the break-up of Gondwana and/or to the Tethyan rifting (Vivier *et al.*, 1976; Kontak *et al.*, 1985; Jaillard *et al.*, 1990; Flint *et al.*, 1993; Sempere *et al.*, 1998). No clear evidences of subduction-related volcanism have been found as yet in these areas.

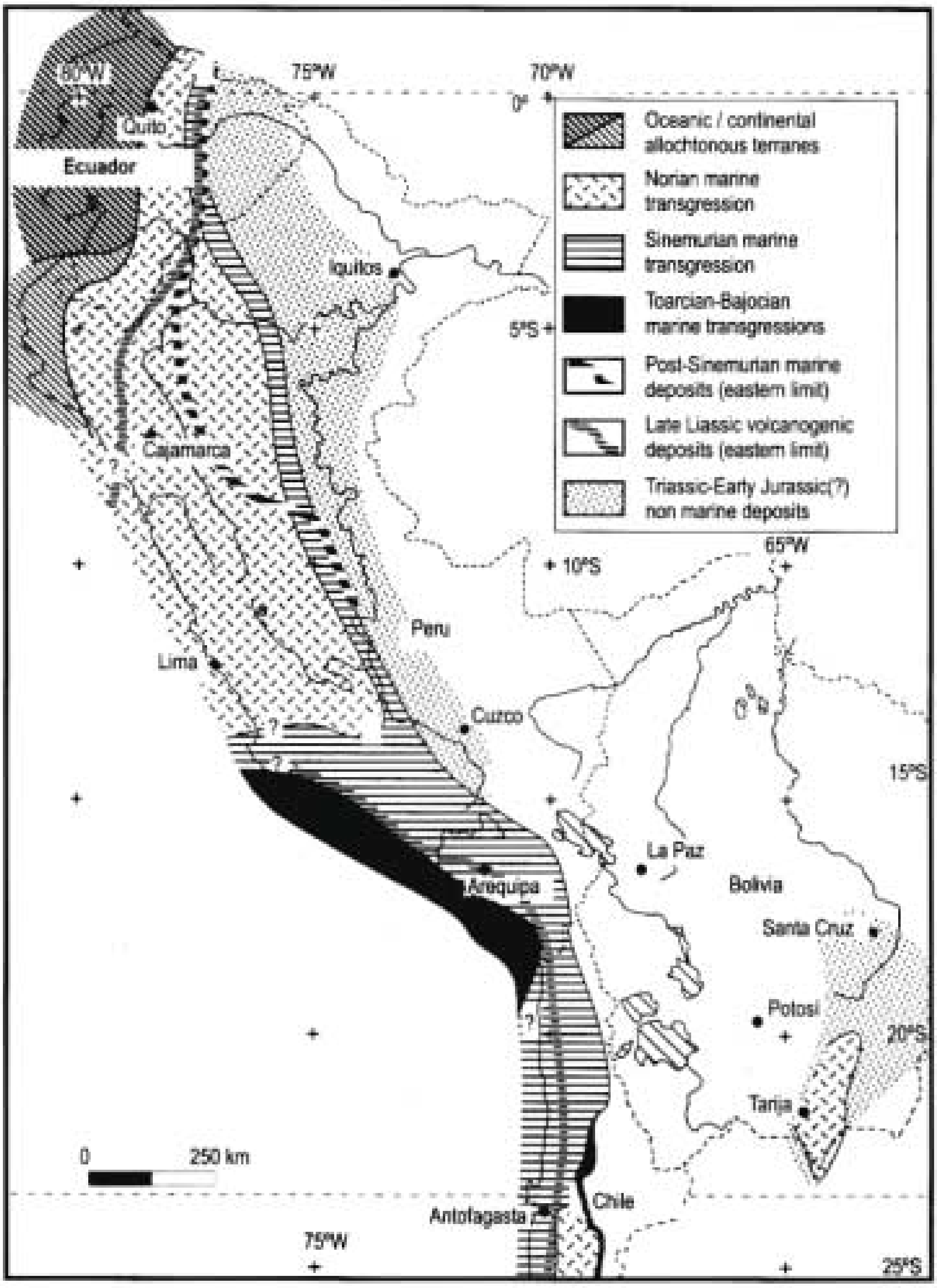
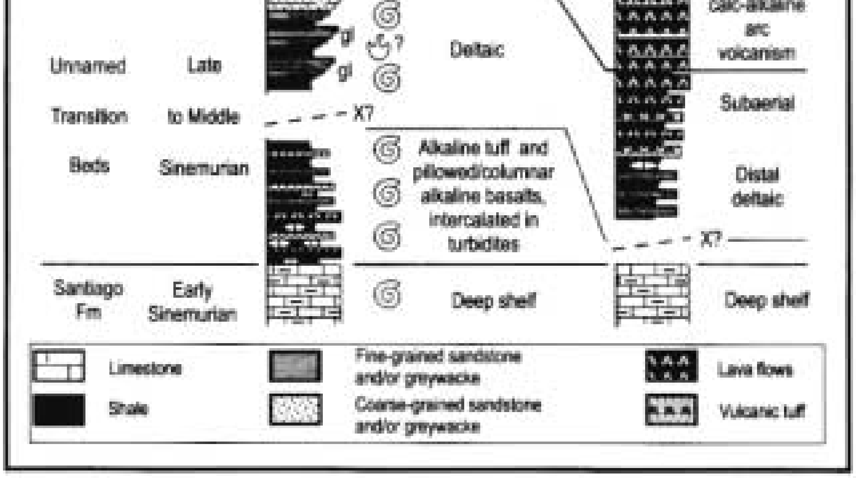


FIGURE 8 - Paleogeographic sketch of the Late Triassic and Early Jurassic marine transgressions.

FIGURE 9 - Transition between the marine shelf limestones (Santiago Fm) intercalated with alkaline lava flows and the subaerial continental arc volcanism (Misahualli Fm) in the southern Subandean zone of Ecuador (Cutucú Cordillera), during the Sinemurian (after Romeuf et al., 1997).



Late(?) Norian - Middle Sinemurian (215 - 195 Ma)

Two main marine transgressions and a relative tectonic quiescence characterize this time-span. The Late Norian marine transgression reached the western part of the Andean margin from southern Colombia to northern Chile (Fig. 8). Norian deposits usually overlie Permian-Triassic subaerial volcanic and sedimentary rocks, locally in angular unconformity, or may overlie the Paleozoic basement. In the central and northern parts of the basin (Ecuador, Peru), Norian deposits consist of shallow-marine shelf limestone and dolomite (Mégard, 1978; Pardo and Sanz, 1979; Loughman and Hallam, 1982; Prinz, 1985; Rivadeneira and Sánchez, 1991; Rosas *et al.*, 1997), whereas in the eastern (eastern Peru, Mégard, 1978) and southern areas (southern Bolivia, northern Chile, Gröschke *et al.*, 1988; Suárez and Dalenz, 1993; Ardill *et al.*, 1998), they begin with a transgressive clastic, locally evaporitic sequence, evolving toward marine deposits.

A stratigraphic hiatus of latest Norian-Rhetian age occurred locally in Peru (Mégard, 1978; Loughman and Hallam, 1982; Prinz, 1985), which may explain local disconformities (Pardo and Sanz, 1979). The second major marine transgression is of Late Hettangian to Early Sinemurian age (Fig. 8). In Peru, these transgressive deposits onlap onto the Permian-Triassic volcanogenic rocks (Mégard, 1978; Loughman and Hallam, 1982). In the central part of the basin, they consist of dark limestone units rich in organic material to bituminous shale, the upper part of which is rich in phosphate (Mégard, 1978; Loughman and Hallam, 1982; Prinz, 1985; Rosas *et al.*, 1997). In many southern areas, the Sinemurian transgressive beds are the first marine deposits to be recorded. They may locally onlap onto the Paleozoic basement (Antofagasta, Muñoz *et al.*, 1988; Baeza and Quinzio, 1991), or overlie undated (Triassic?) arc volcanic rocks (southern Peru, Vargas, 1970; Vicente *et al.*, 1982; Iquique and Arica, Muñoz and Charrier, 1993).

In the back-arc areas of central Peru, scattered lava flows interbedded in Early Liassic marine sediments display alkaline chemistry, indicating that extensional tectonic regime still prevailed (Rosas and Fontboté, 1995; Romeuf *et al.*, 1997; Figs. 9 and 12). In summary, the Late Triassic-Early Jurassic period is marked by a mild extensional regime, interpreted as related to the break-up of Gondwana and to the Tethyan rifting. Large-scale marine transgressions occurred in the Late Norian and Early Sinemurian, respectively. Subduction of a paleo-Pacific Plate beneath this part of the South American Plate has not been proved as yet.

Late Sinemurian - Kimmeridgian (195 - 150 Ma)

This interval is marked by active subduction-related magmatism, sinistral wrenching movements along NW-SE faults, and tectonically controlled sedimentation. Three main domains are distinguished: (1) the NNE trending Ecuadorian-northern Peruvian segment is characterized by a NNE-trending magmatic arc bordered to the E by a back-arc subaerial basin; (2) in northern and central Peru, the carbonate shelf sedimentation continued and was interrupted by a Middle Jurassic tectonic event; (3) in southern Peru and northern Chile, the western magmatic arc was flanked to the E by a marine, subsiding back-arc basin, bordered by emergent eastern areas (Fig. 8).

In Colombia (Mojica and Dorado, 1987), southern Ecuador (Romeuf *et al.*, 1997) and northwestern Peru (Pardo and Sanz, 1979), Middle to Late Sinemurian marine deposits are intercalated with alkaline basaltic flows, whereas subduction-related volcanic arc rocks appear in the immediately overlying deposits, which evolve toward subaerial environments. The Late Sinemurian age of the beginning of the magmatic arc activity is confirmed by dates obtained from associated arc-related intrusions (195 - 190 Ma, Aspden *et al.*, 1987; Litherland *et al.*, 1994). The activity of this NNE trending continental magmatic arc (Misahualli and Colán formations) continued on during the Middle Jurassic, and eventually ceased by latest Jurassic times (150 - 140 Ma, Aspden *et al.*, 1987; Litherland *et al.*, 1994, Romeuf *et al.*, 1995). To the W of the magmatic arc, marine volcaniclastic sediments, dated as late Middle to early Late Jurassic in southern Ecuador and northern Peru (Mourier, 1988), are interpreted as fore-arc deposits.

In southwestern Peru, the disconformable marine transgression is dated as Early Sinemurian, Early Toarcian, Late Toarcian (commonly), or even Late Bajocian-Bathonian, according to the regions, suggesting either a contrasted and complex paleogeography, or the juxtaposition of distinct units due to subsequent displacements (Kurth *et al.*, 1996). Most of the successions display a sedimentary hiatus of Late Bajocian-Bathonian age (Vicente, 1981, 1989). Volcanic arc rocks formerly ascribed to the Triassic-Liassic (Chocolate Formation) yielded radiometric dates of 177 - 157 Ma and Aalenian-Bathonian paleontological ages (Roperch and Carlier, 1992; Romeuf *et al.*, 1995), thus suggesting that they are coeval with the magmatic arc of Ecuador, northern Peru and northern Chile. These volcanic rocks are overlain by volcaniclastic and siliciclastic marine shelf deposits of Aalenian-Callovian age (Upper Río Grande and Guaneros formations, Rüegg, 1956; Aguirre and Offler, 1985; Romeuf *et al.*, 1995), thus indicating a progressive and significant decrease of the arc activity during the Middle Jurassic. No volcanic rocks of Late Jurassic age known.

In the coast of northern Chile, outpouring of a thick accumulation of subduction-related dacite, andesite, basalt and tuff (La Negra Formation, Fig. 10) began also during the Late Sinemurian-Pliensbachian (Muñoz *et al.*, 1988; Baeza and Quinzio, 1991; Muñoz and Charrier, 1993). They follow basic intrusions related to probably transtensional sinistral movements (Pichowiak *et al.*, 1990). In northernmost Chile, this volcanic series is overlain by marine sediments of early Middle Jurassic to Late Jurassic age, according to the areas (Muñoz and Charrier, 1993; Kossler, 1997), but volcanic activity is assumed to have lasted until the Late Jurassic (Charrier and Muñoz, 1994) or even the Early Cretaceous (Mpodozis and Ramos, 1989; Scheuber *et al.*, 1994), farther S. Extensional or transtensional regime prevailed during the Jurassic, favouring the formation and play of the Atacama wrench fault system (Brown *et al.*, 1993).

East of the magmatic arc, the Jurassic back-arc basin received poorly dated, mainly argillaceous and volcaniclastic subaerial sedimentation (Chapiza and Sarayaquillo formations; Tschopp, 1953; Seminario and Guizado, 1976; Rivadeneira and Sánchez, 1991; Fig. 6). These sedimentary rocks rest on the Liassic carbonate beds to the W, or on the Paleozoic basement to the E. In the latter case, they may include unfossiliferous Late Triassic-Liassic beds. They comprise a lower sequence of shale, fine-grained sandstone, dolomite and evaporite of sabkha environment, and an upper sequence consisting of coarse-grained sandstone and conglomerate of fluvial environment (Tschopp, 1953; Mégard, 1978). However, Jurassic marine platform limestone units are locally known in easternmost Ecuador (Petroproducción, unpublished data).

In the back-arc areas of Peru, platform carbonate deposition, started in the Late Triassic, went on until the early Middle Jurassic (Figs. 11 and 12). This carbonate shelf was bordered to the E by continental back-arc deposits consisting in fluvial red beds to the N, and locally significant accumulations of eolian sandstone to the S (Sempere, 1995). No information is available on the western part of central Peru. On the Peruvian Platform, the bituminous facies and phosphate-rich carbonate beds of the Sinemurian transgression (Aramachay Formation) are overlain by stratified oolitic and skeletal limestone presenting local cross-stratified calcarenite beds of Late Sinemurian to Late Toarcian, maybe Aalenian age (Condorsinga Formation; Mégard, 1978; Westermann *et al.*, 1980; Loughman and Hallam, 1982; Prinz, 1985). These Liassic deposits are overlain by sandy-argillaceous limestone of Late Aalenian-Late Bajocian age (Chunumayo Formation, Mégard, 1978; Westermann *et al.*, 1980). The carbonate series ends locally (Huancayo) with unconformable fluvial sandstone (Cercapuquio Formation) overlain by laminated carbonate, shale and evaporite beds of tidal flat environment, ascribed to the Bajocian (Chaucha Formation; Mégard, 1978; Moulin, 1989; Fig. 11). However, in many areas, the upper part of the platform carbonate sequence is eroded (Prinz, 1985), due to the Bathonian tectonic event.

In the back-arc areas of northern Chile, the Sinemurian transgressive beds are overlain by open marine deposits (Chong, 1976; Vicente *et al.*, 1982; Muñoz *et al.*, 1988; Ardill *et al.*, 1998). In southern Peru (Arequipa Trough), this succession comprises a Pliensbachian to Bajocian diachronic transgression, Bathonian turbidite beds, Callovian black shale and Oxfordian-Kimmeridgian shelf sandstone (Vicente, 1981, 1989; Vicente *et al.*, 1982; Fig. 13). In northernmost Chile, the turbidite beds progressively disappear and carbonate interbeds appear in the succession (Muñoz *et al.*, 1988). Farther S (Domeyko Cordillera), the succession is dominated by offshore shale deposits locally rich in organic matter (Chong, 1976; Ardill *et al.*, 1998). These marine back-arc basins are bordered to the E (southern Peru, Bolivia) by emergent areas which received locally thick, coarse-grained, quartz-rich eolian sandstone ascribed to the Early to Middle Jurassic (Ravelo and Ichoa formations; Oller and Sempere, 1990; Carlotto, 1998; Sempere *et al.*, 1998).

Scattered alkaline lava flows are known during this interval in the back-arc areas. Alkaline basalt flows are interbedded in the Sinemurian limestone of central Peru (Rosas *et al.*, 1997), scarce sills and intracontinental tholeiitic flows dated at 185 - 170 Ma occur in Bolivia (late Early to early Middle Jurassic, Soler and Sempere, 1993; Sempere *et al.*, 1998), and volcanic flows interbedded in the Bajocian to Callovian deposits of the Domeyko Basin are regarded as related to transtensional movements (Ardill *et al.*, 1998; Fig. 14). On the other hand, the mid-Jurassic Arequipa Trough has been interpreted as a pull-apart basin due to sinistral movements (Vicente *et al.*, 1982). Therefore, this period seems to be dominated by a mild extension, probably due to a sinistral transtensional regime (see also Scheuber *et al.*, 1994).

However, the southern part of the area (10°S - 25°S) is characterized by a sedimentary hiatus of Late Bajocian-Bathonian age (Mégard, 1978; Vicente, 1989; Ardill *et al.*, 1998). This "Bathonian phase" is classically regarded as responsible for the disconformity that separates the lower and upper red beds of the Peruvian Eastern Basin (Mégard, 1978). It is coeval with a rapid exhumation (3000 m) of the Eastern Cordillera of central Peru (Laubacher and Naeser, 1994), with metamorphism of the arc zone of northern Chile (170 - 150 Ma, Lucassen and Thirlwall, 1998) and with intense folding and reverse faulting of the arc zone of north central Chile (163 - 140 Ma, Irwin *et al.*, 1997). Therefore, a poorly documented compressional deformation seems to have occurred in the Late Bajocian-Bathonian.

In summary, between 195 and 140 Ma, the NNE trending Colombian-Ecuadorian and the N-S trending Chilean segments are characterized by abundant arc magmatism, unstable tectonic regime, subaerial back-arc sedimentation and locally marine fore-arc deposits. At the same time, transtensional, probably sinistral tectonic regime prevailed along the NW-SE trending Peruvian segment, which resulted in localized pull-apart basins and in the alternation of marine sedimentation and emergent periods. This situation has been interpreted as the result of a southeastward convergence of the paleo-Pacific Plate beneath South America (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995).

Kimmeridgian to late Paleocene (140 - 57 Ma)

During the Kimmeridgian-Paleocene interval, the paleogeographic framework was still controlled by the fore-arc, arc and back-arc zones. However, the trend of the arc zone became NE in Peru and N-S in Chile, and the back-arc area presented a three-fold subdivision inherited from the Jurassic structures. No magmatic arc is known at this time in Ecuador, which constituted a back-arc domain.

The Kimmeridgian-early Late Cretaceous fore-arc zone is virtually unknown. During the latest Cretaceous, fore-arc zones appear in northernmost Peru-southwesternmost Ecuador and southern Peru-northern Chile, due to the eastward shift of the magmatic arc from 80 Ma. The magmatic arc crops out widely in coastal Peru and northern Chile. The northern extension of the NW trending Peruvian arc crosscuts obliquely the southernmost part of Ecuador, since the subduction zone probably extended northwestwards into the oceanic realm. However, insights into the geodynamic evolution of the East-Pacific oceanic realm are provided by the analysis of the stratigraphic content and geochemical

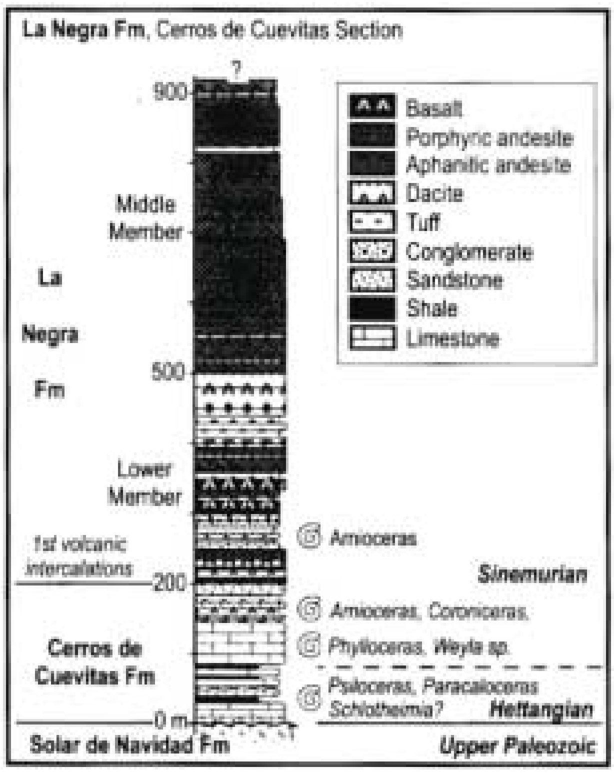
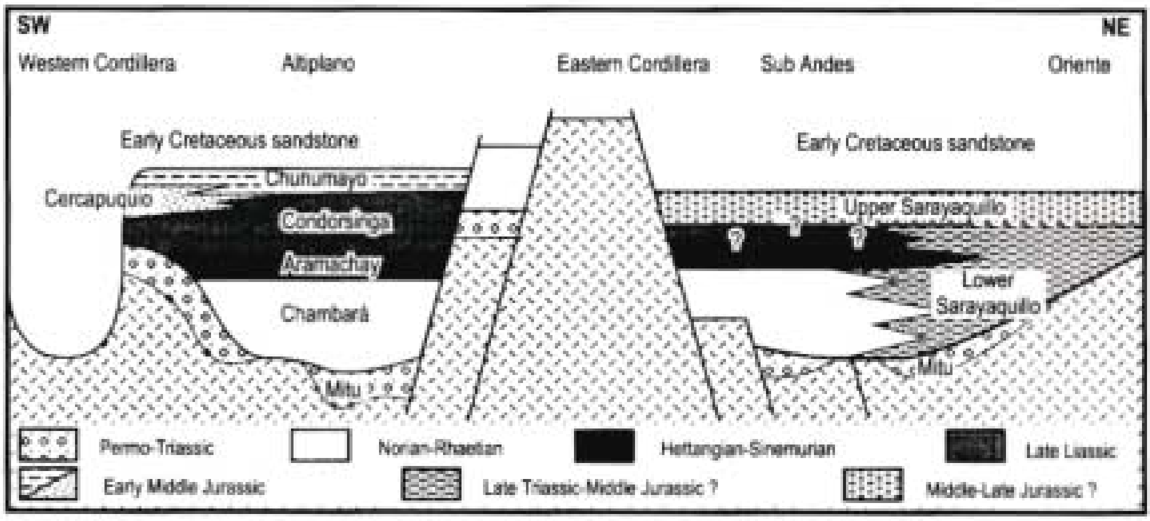
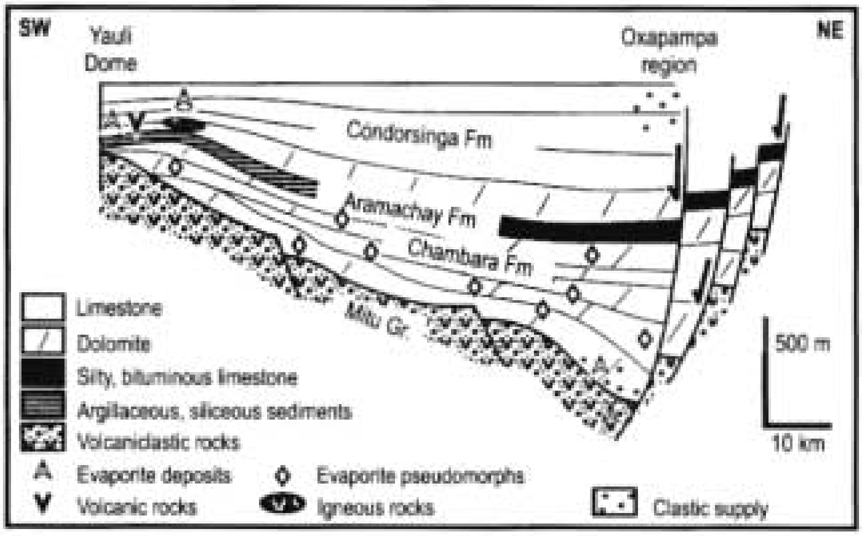


FIGURE 10 - Section of the La Negra Formation in the Jurassic arc zone of northern Chile (after Muñoz et al., 1988).

FIGURE 11 - Paleogeographic sketch of the Pucara Group (Late Triassic-Early Middle Jurassic) of Central Peru (after Loughmann and Hallam, 1982). No scale.

FIGURE 12 - Paleogeographic sketch of the Pucara Group (Late Triassic-Early Middle Jurassic) of Central Peru (after Rosas and Fontboté, 1995). Compare with Fig. 11.





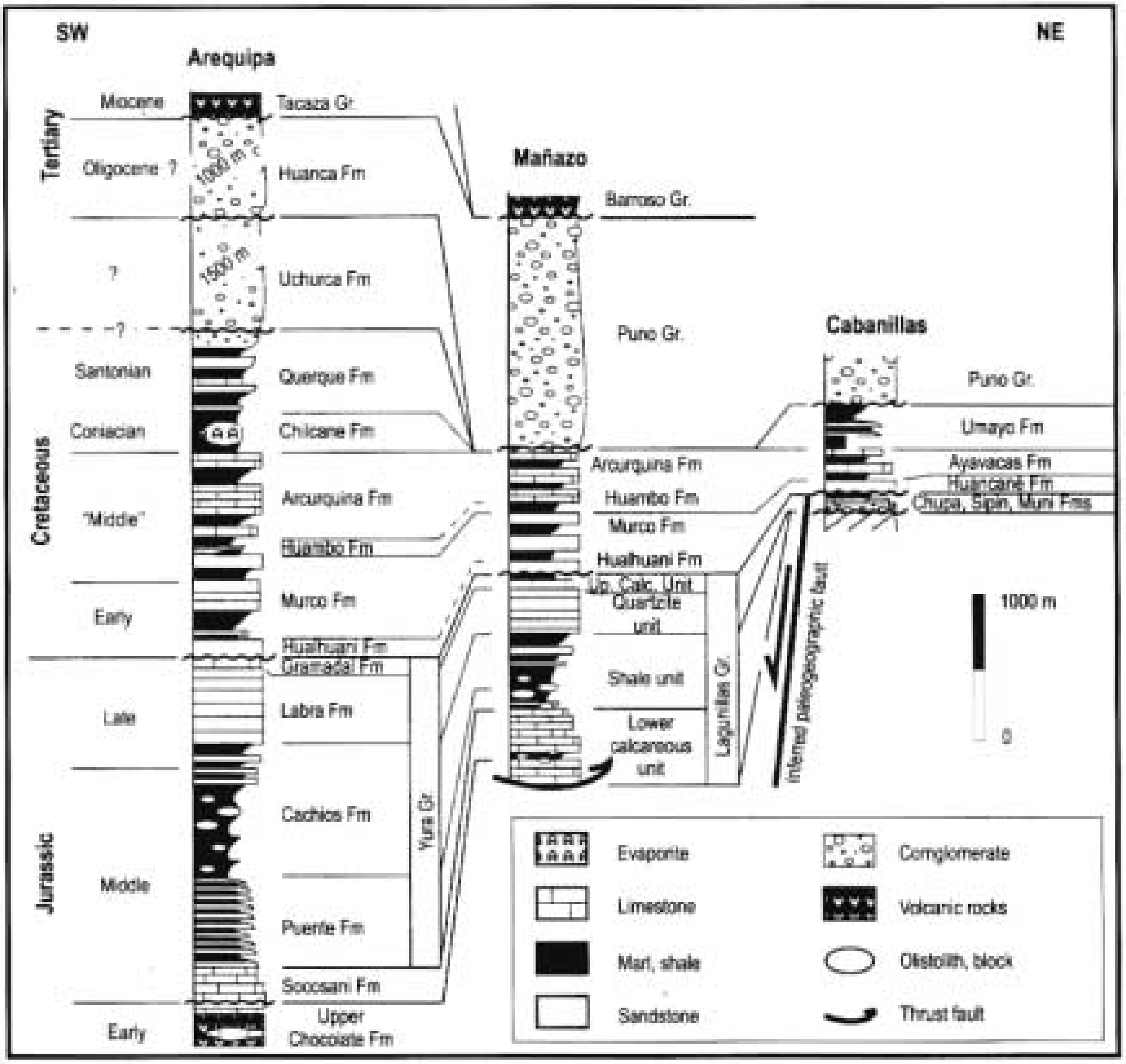


FIGURE 13 - Sections of the Jurassic-Tertiary ssuccessions accross the Arequipa Basin of southwestern Peru (after Jaillard and Santander, 1992).

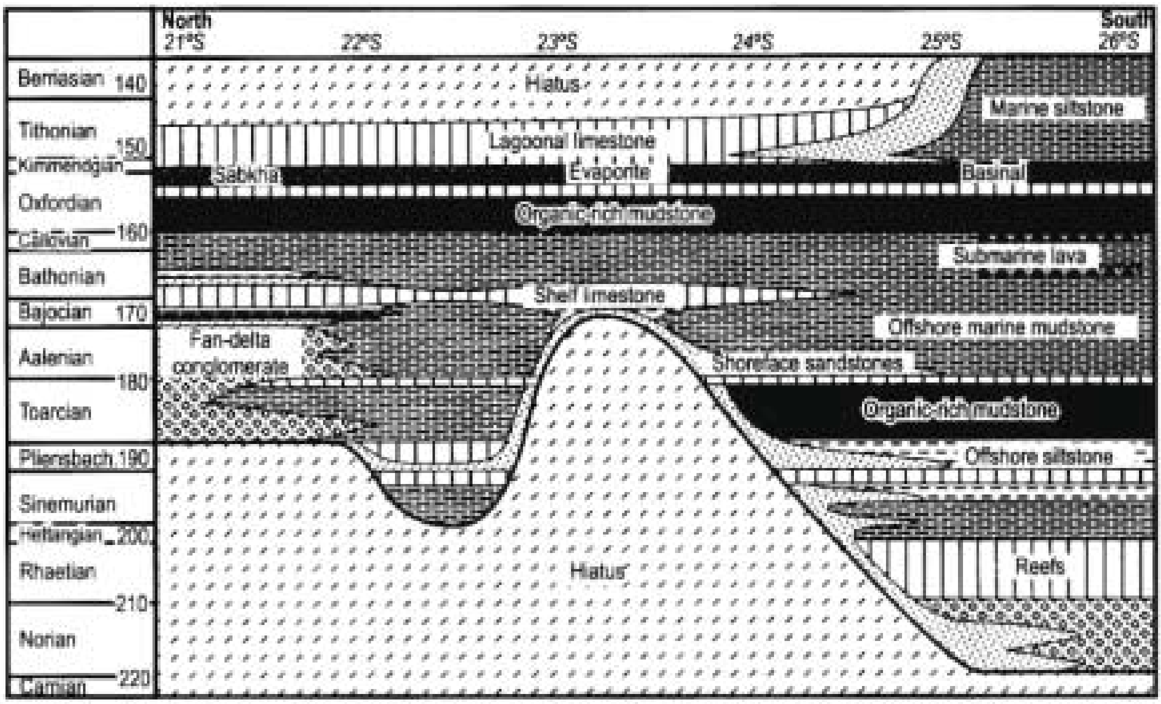


FIGURE 14 - Chronostratigraphic chart of the Jurassic back-arc Domeyko Basin (northern Chile) (after Ardill et al., 1998).

signatures of the Cretaceous oceanic terranes subsequently accreted to the Ecuadorian margin.

In Peru, the back-arc area comprised a western, mobile and subsiding basin (West-Peruvian Trough of Benavídes, 1956, present-day Western Cordillera), an axial positive threshold (Marañon Geanticline, Cuzco-Puno Swell, and present-day Eastern Cordillera) and an eastern, stable and moderately to little subsiding basin (present-day Subandean Zone and eastern basin, Fig. 16). The latter extends northwards into Ecuador (present-day Oriente Basin), and southwards into Bolivia (Potosí Basin). The Kimmeridgian-late Paleocene time-span (150 - 55 Ma) can be divided into three periods, characterized by distinct paleogeographic and tectonic settings, controlled by the plate kinematics evolution.

Kimmeridgian? - Aptian (150 - 110 Ma)

Latest Jurassic to earliest Cretaceous times are marked by a complete reorganization of the paleogeographic pattern and tectonic evolution of the Andean margin, interpreted as the result of a drastic change in the convergence direction, which triggered a magmatic arc re-organization and significant tectonic deformation (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995). These events were followed in the Early Cretaceous by an important diachronous marine transgression in the whole area.

Kimmeridgian? - Berriasian (150 - 135 Ma)

In Ecuador and northern Peru (and Colombia), the activity of the continental volcanic arc (Misahualli and Colán formations) ceased by the end of Kimmeridgian times (150 to 140 Ma ago, Aspden *et al.*, 1987; Mourier, 1988; Litherland *et al.*, 1994; Fig. 17). The deformed and eroded magmatic arc is then overlain by unconformable fluvio-marine sandstone, the diachronous base of which is dated as Valanginian (?) to Albian, from SW to NE (Villagómez *et al.*, 1996; Robert *et al.*, 1998). This major hiatus and unconformity suggests the occurrence of a significant latest Jurassic-Early Cretaceous tectonic event (Litherland *et al.*, 1994). No Kimmeridgian deposits are known in the fore-arc areas of Ecuador and northern Peru.

In southwestern Ecuador, the Raspas high pressure metamorphic complex yielded a 132 Ma K/Ar age (Feininger and Silberman, 1982) and 130 - 115 Ma Ar/Ar and Sm/Nd ages (Malfere, 1999). These are interpreted as cooling ages, subsequent to the accretion and HP metamorphism of an oceanic plateau and accretionary prism in the latest Jurassic-earliest Cretaceous (Gabriele *et al.*, 1999; Malfere *et al.*, 1999). This suture extends northwards along the western edge of the Eastern Cordillera of Ecuador (Peltetec suture, Litherland *et al.*, 1994; Aspden *et al.*, 1995) and to the W of the Central Cordillera of Colombia (Amaime Terrane, Aspden and McCourt, 1986; Toussaint and Restrepo, 1994). According to Litherland *et al.* (1994), this event corresponds to the accretion of a continental microplate (Chaucha Terrane) to the Ecuadorian margin.

In northwestern Peru, Early (?) Tithonian lagoonal deposits are abruptly overlain by deep shelf shale, and then by a thick series of coarse-grained volcaniclastic turbidite beds of Late Tithonian age, which reworked the Jurassic volcanic arc. This evolution points to the creation of a deep, N-S trending sedimentary basin interpreted as a pull-apart basin (Jaillard and Jacay, 1989), which extends southwards (Fig. 6). This succession follows (Jaillard and Jacay, 1989) with Late Tithonian black shale of shelf environment, Tithonian-Berriasian massive sandstone of nearshore environment (Tinajones Formation; Wilson, 1984), and disconformable massive clean sandstone of presumably Berriasian age (Chimú Formation, Benavídes, 1956; Jaillard and Jacay, 1989).

In the back-arc areas of Ecuador and northern Peru, the Late Jurassic back-arc red beds seem to grade upwards into coarser-grained red beds interbedded with basaltic to rhyolitic flows (Yaupi Mb, Jaillard, 1997), locally dated as earliest Cretaceous (Hall and Calle, 1982; Jaillard, 1997). This magmatic activity would be coeval with small-sized stocks, which intrude the Jurassic red beds and are disconformably capped by Early Cretaceous sandstone (Tschopp, 1953; Tafur, 1991).

In central and southern Peru, Kimmeridgian times were marked by the arrival of clastic deposits. Siliciclastic shelf sediments abruptly overlie the Callovian black shale of the Arequipa Trough (Vicente *et al.*, 1982; Fig. 13). These sediments are correlated with undated unconformable conglomerate of the Peruvian (Chupa Formation, Klink *et al.*, 1986; Jaillard, 1995) and Bolivian Altiplano (Condo Formation), which, however, might be younger. The Early Tithonian marine transgression is recorded in the western part of central and southern Peru by shallow marine limestone (Jaguay Formation, Rüegg, 1961; Gramadal Formation, Chávez, 1982; Batty and Jaillard, 1989), which appear to be overlain by Tithonian-Berriasian (?) black shale and sandstone (Oyón Formation, Mégard, 1978; Tiabaya outcrops, Geyer, 1982).

Volcanic arc activity is known in the Lima area (Atherton *et al.*, 1985; Alemán, 1996). It seems to have begun by Late Tithonian times, since the upper part of the arc series yielded ammonites of Late Tithonian age (Bulot, personal communication, 1998; formerly ascribed to the Berriasian; Wiedmann, 1981), and have continued during part of Berriasian times (Alemán, 1996). Disconformable clean massive sandstone is ascribed to the Berriasian in the western parts of central and southern Peru (Chimú, Goyllarisquizga and Hualhuani formations, Benavides, 1956; Batty and Jaillard, 1989). However, in spite of limited paleontological evidence, the base of these deposits seems to be diachronous, being much younger toward the E (Wilson, 1963; Jaillard, 1995; Robert *et al.*, 1998). In the eastern basins of Peru and Bolivia, no earliest Cretaceous deposits have been recognized so far, below the unconformable Early Cretaceous sandstone units.

In northernmost Chile, the Kimmeridgian phase is marked by a marine regression which a culminated with a local hiatus and an angular unconformity, by sinistral wrench movements along the N-S trending Atacama Fault System, and by subsidence of the back-arc areas (Bogdanic and Espinosa, 1994). In the back-arc basin of northernmost Chile, Oxfordian limestone and shale grade upwards into limestone and shale with interbeds of evaporite in the lower part, and sandstone intercalations in the upper part (Muñoz *et al.*, 1988). Farther S (Domeyko Basin), the Callovian marine black shale beds are overlain by evaporite beds deposited in basin to sabkha environments, and then by inner shelf limestone of Late Kimmeridgian-Early Tithonian age, which laterally grade southwards into marine siltstone (Chong, 1976; Ardill *et al.*, 1998; Fig. 14). Farther S (31°S), the Coastal Cordillera was deformed by W-verging open to tight folds between 140 and 126 Ma (Irwin *et al.*, 1987). The deformation is coeval with significant sinistral lateral displacements of fore-arc and arc slivers along the Atacama Fault System (26°S), dated between 145 and 125 Ma (Kurth *et al.*, 1996).

During the latest Jurassic-earliest Cretaceous, magmatic arc activity seems to have continued without changes of location. However, a magmatic gap seems to have occurred between 150 and 140 Ma (Hammerschmidt *et al.*, 1992; Fig. 15) and the magmatic activity appears to decrease significantly near the Jurassic-Cretaceous boundary (Mpodozis and Ramos, 1989; Scheuber *et al.*, 1994; Charrier and Muñoz, 1994).

The end of the NNE trending Ecuadorian-north-Peruvian magmatic arc by Tithonian times, and the beginning of the *activity of the NW trending Peruvian magmatic arc in the Late* Tithonian, expresses a drastic change in the convergence direction, which passed from nearly southwards to nearly northeastwards (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995). This plate kinematics reorganization correlates with geodynamic events in the southeastern Pacific and the Tethys Ocean. During the Late Jurassic, the southeastern Pacific accretion ridge would have been oriented roughly NE-SW and connected with the Tethyan accretionary system (Caribbean and Central Atlantic oceans, Duncan and Hargraves, 1984). The outpouring of a major plume along the *Pacific accretionary ridge in Tithonian times (Chatzky ridge)* would have disturbed the accretion direction; brought about the break-up of the Eastern paleo-Pacific Plate, and created a triple junction (Nakanishi *et al.*, 1989). Meanwhile, the Tethyan realm was marked by a significant slowdown in the spreading rates, allowing the newly created NW-SE trending Southeastern Pacific Ridge to impose a northeastward drift direction for the Eastern paleo-Pacific Plate. These events would have provoked, in the Tithonian (140 Ma), a change of convergence direction, according to the process proposed by Duncan and Hargraves (1984) for Early Cretaceous times.

This major geodynamic change may account for the accretion of the oceanic terrane of Ecuador and Colombia, the creation of the Chicama Basin of northern Peru, the widespread emergence and subsequent gap of Late Tithonian-Berriasian deposits, the unconformity of the Early Cretaceous deposits and the possible compressional deformation recorded in northern Peru. These events may be correlated with the Araucan phase of northern Chile (Stipanicic and Rodrigo, 1969; Scheuber *et al.*, 1994).

Berriasian-Aptian (140 - 110 Ma)

This period is marked by the widespread deposition of disconformable, diachronous quartz-sandstone units, by a relative tectonic quiescence; and along the Peruvian and Chilean margins, by the ongoing, although mild, volcanic arc activity. In Ecuador and northern Peru, very little is known about this period. In the back-arc areas of Ecuador, one K/Ar date and one palynological age suggest that subaerial red bed deposition continued during part of this period (Bristow and Hoffstetter, 1977; Baldock, 1982). The overlying transgressive disconformable sandstone beds are of Albian age (Jaillard, 1997). In the Cordillera Real of Ecuador, Litherland *et al.* (1994) mentioned numerous K/Ar resets in the Jurassic granite, interpreted as due to significant dextral movements related to the collision of displaced terranes. In the neighbouring Marañon Basin of northeastern Peru, a sedimentary hiatus seems to separate the Jurassic red beds and the disconformable, diachronous transgressive sandstone beds of Early Cretaceous age (Laurent, 1985).

Meanwhile, somewhere in the southeastern paleo-Pacific domain, a mantle plume was responsible for the formation of a large and thick Oceanic Plateau that was accreted to the Andean margin in Late Cretaceous times (Cosma *et al.*, 1998; Reynaud *et al.*, 1999), and crops out presently in the Western Cordillera of Ecuador where it has been dated as Barremian or Hauterivian (123 Ma, Lapierre *et al.*, 1999).

In Central and Southern Peru, scarce outcrops of volcanic rocks in the coastal and western areas suggest that volcanic arc activity continued at least locally until Aptian times (Bellido, 1956; Vidal *et al.*, 1990; Alemán, 1996). However, the occurrence of thick intercalations of quartz-rich sandstone and unusual volcanic-free shelf limestone in the Lima area (Rivera *et al.*, 1975; Alemán, 1996) suggests that, either volcanic activity was local and sporadic, or part of the coastal terranes have been subsequently displaced (May and Butler, 1985).

Farther E, in the back-arc areas, well-sorted, clean, quartz-sandstone beds were deposited from Late Berriasian times onwards (Benavides, 1956; Wilson, 1963). In northern Peru and Ecuador, these were dated as Late Berriasian-Early Valanginian in the eastern region, (Benavides, 1956; Rivera *et al.*, 1975; Bulot, personal communication, 1998), Berriasian-Barremian in the western part of the Eastern Basin of Peru (Tarazona, 1985), Aptian in the centre of the basin, and early Late Albian in the eastern parts of the Eastern Basin (Villagómez *et al.*, 1996; Robert *et al.*, 1998, Fig. 18). In central Peru, such a diachronism was suggested by Wilson (1963), and, although paleontological evidence is poor, a comparable diachronism may occur in southern Peru (Jaillard, 1995).

Paleocurrents indicate that the Guiana and Brazilian shields were the sources of the clastic supply. Paleoenvironments evolve from subaerial/fluvial to nearshore/shallow marine from E to W, and deposition is mainly controlled by eustatic variations (Moulin, 1989). The isopach map clearly indicates an eastern depocenter situated in northern Peru (present Marañon River), and western depocenters N of Lima and around Arequipa (Jaillard, 1994; Fig. 19). Scarce syn-sedimentary tectonic features suggest a mild extensional regime (Moulin, 1989).

In the back-arc area of northernmost Chile, Late Jurassic strata are overlain by fine-grained sandstone, siltstone and shale with occasional tuff and andesitic lava of

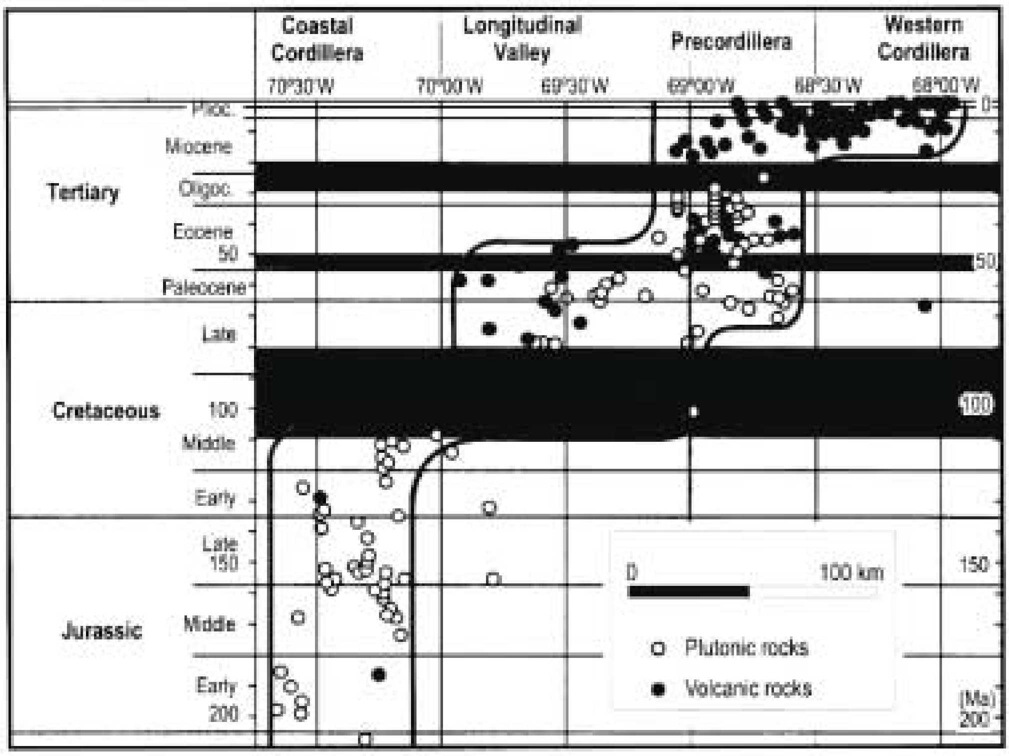


FIGURE 15 - Age, location and nature of the magmatic arc rocks in northern Chile (after Hammerschmidt et al., 1992).

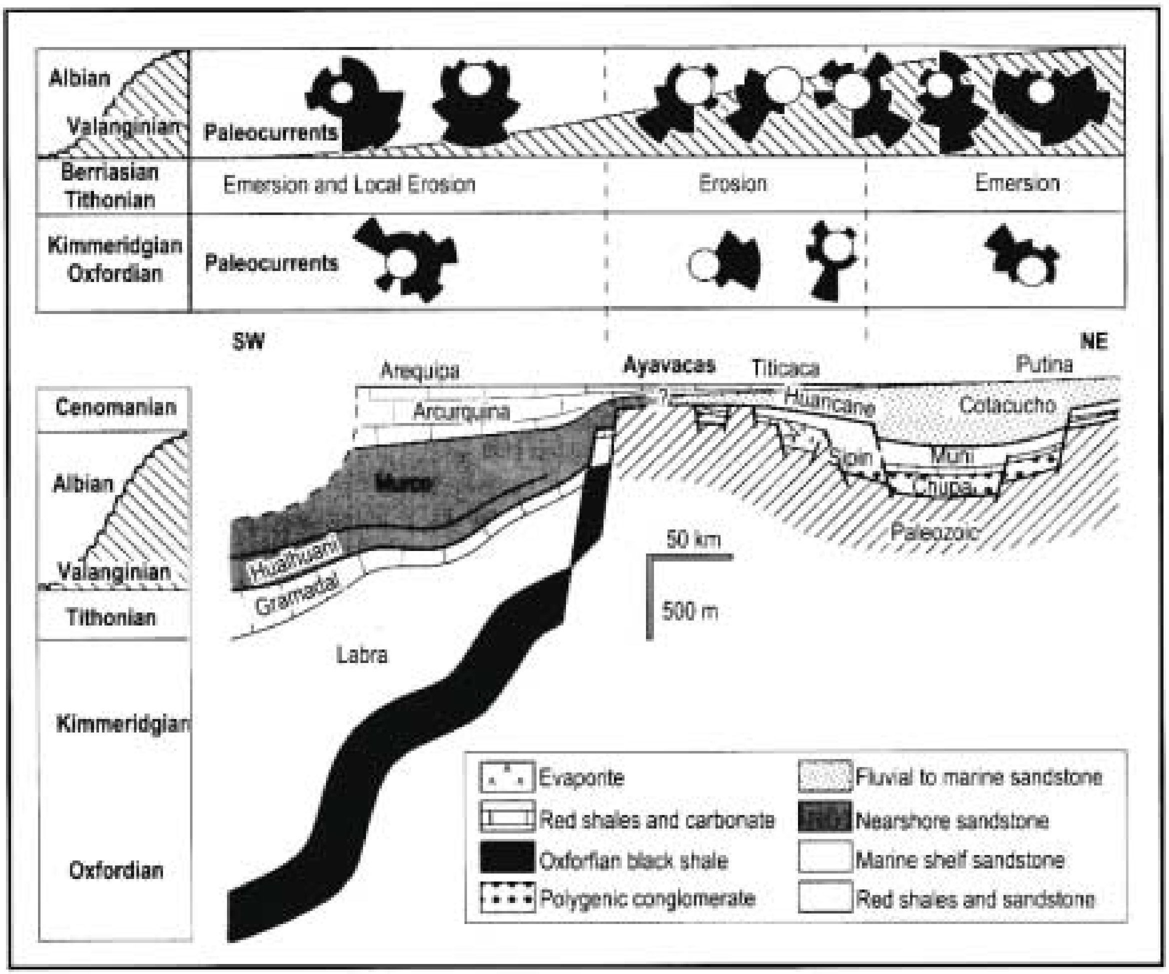


FIGURE 16 - Paleogeographic profile of southern Peru at the end of the Cenomanian (after Jaillard, 1994). Note the change in paleocurrent directions between Late Jurassic and Early to Middle Cretaceous times.

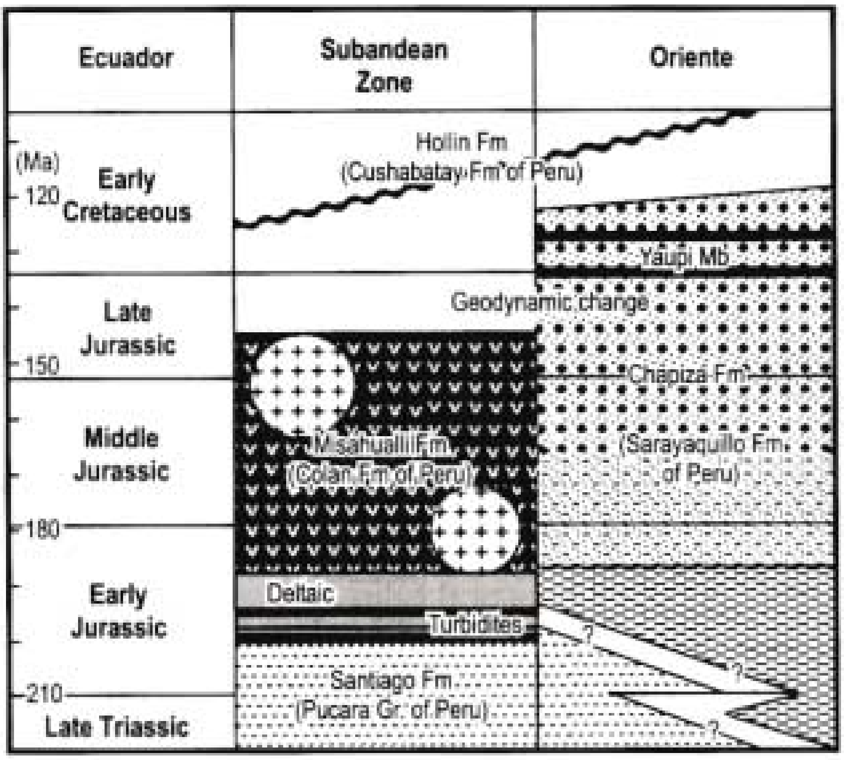


FIGURE 17 - Chronostratigraphic sketch of the Jurassic-Early Cretaceous series of eastern Ecuador (after Jaillard, 1997).

Kimmeridgian-Barremian age (Bogdanic and Espinosa, 1994; Scheuber *et al.*, 1994). Farther S, a limestone unit of Hauterivian-Barremian (and Aptian?) age crops out in the Coastal Cordillera (Bogdanic and Espinosa, 1994). The Early Cretaceous volcanic arc crops out farther S in the Coastal Cordillera. There, magmatic arc activity went on without significant changes in the location of the magmatic front; a maximum activity is recorded in the Aptian (120 - 110 Ma, Hammerschmidt *et al.*, 1992; Fig. 15). In Bolivia, undated continental conglomerate and sandstone are ascribed to the Early Cretaceous, since they conceal the "Araucan" angular unconformity (Sempere, 1994). They crop out in the centre of the Potosí Basin and are associated with alkaline basalt and basaltic andesite indicating extensional conditions (Soler and Sempere, 1993). Farther to the SE, in northern Argentina, sub-alkaline to alkaline granitic plutons (130 - 120 Ma) indicate that an extensional tectonic regime prevailed in the distal back-arc areas (Viramonte *et al.*, 1999).

The abrupt arrival in the Early Cretaceous of huge amounts of detrital quartz derived from the E in northern South America, may be interpreted as the result of both the large-scale westward tilt of the South American Plate due to the South Atlantic rifting, and a significant climatic change with increased precipitation, which allowed the detrital material to be transported for large distances (Jaillard, 1994).

Albian - Turonian (110 - 88 Ma)

This period was marked by a large-scale marine transgression; by important magmatic activity along the Chilean and Peruvian margins; and by the beginning of compressional deformation ("Late Albian Mochica phase", Mégard, 1984; 105 - 100 Ma). The only known outcrop corresponding to the fore-arc zone is represented by the eastern flank of the Amotape-Tahuin Massif of northern Peru-southern Ecuador. There, Paleozoic rocks are covered by undated transgressive siliciclastic conglomerate beds, overlain by massive shelf-limestone and anoxic laminated black limestone of Early to Middle Albian age (Olsson, 1944; Fischer, 1956; Reyes and Caldas, 1987; Jaillard *et al.*, 1999). This succession is comparable to that of the coeval series known from the back-arc areas of northern Peru.

The shelf carbonate sedimentation was rapidly overlain by basinal black shale interbedded with siliciclastic turbidite beds of Late Albian age, exhibiting slumping and bearing large-scale olistoliths (Copa Sombrero Group or Formation; Morris and Alemán, 1975; Reyes and Caldas, 1987; Jaillard *et al.*, 1999). These express an unstable tectonic setting, interpreted as the result of the creation of a pull-apart basin (Lancones-Celica Basin) related to the northward migration of the Amotape-Tahuin Paleozoic massif (Jaillard *et al.*, 1999). In the Cenomanian and Turonian, continuation of the turbiditic sedimentation (Morris and Alemán, 1975) suggests that the northward migration of the Amotape-Tahuin fore-arc sliver went on, with a possible maximum tectonic activity during the Cenomanian (Jaguay Negro Formation).

During Albian times, the arc zone of Peru and southernmost Ecuador was marked by the outpouring of huge volumes of subduction-related calc-alkaline lava (Casma Group, Celica, Copara and Matalaque formations; Atherton *et al.*, 1983; Soler 1991; Reynaud *et al.*, 1996; Fig. 20), which are locally interbedded with ammonite-bearing sediments indicating a Middle to Late Albian age (Myers 1975; Reyes and Caldas, 1987). These lava flows are associated with volcaniclastic turbidite beds deposited in strongly subsiding intra-arc basins (Atherton and Webb, 1989), interpreted as pull-apart basins related to dextral wrenching (Soler, 1991). Folds in the Albian volcanogenic pile are locally

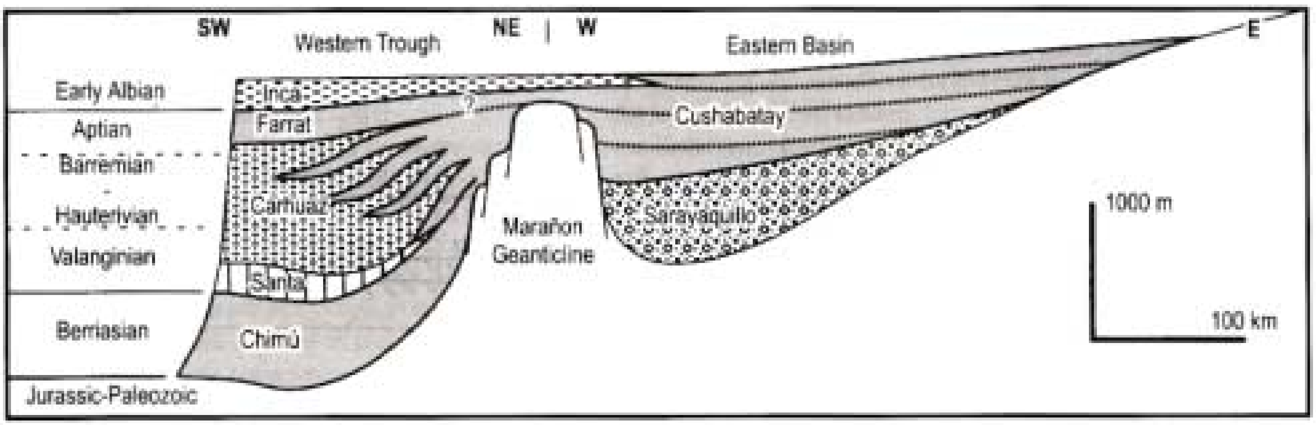


FIGURE 18 - Paleogeographic profile of southern Peru at the end of the Early Cretaceous.

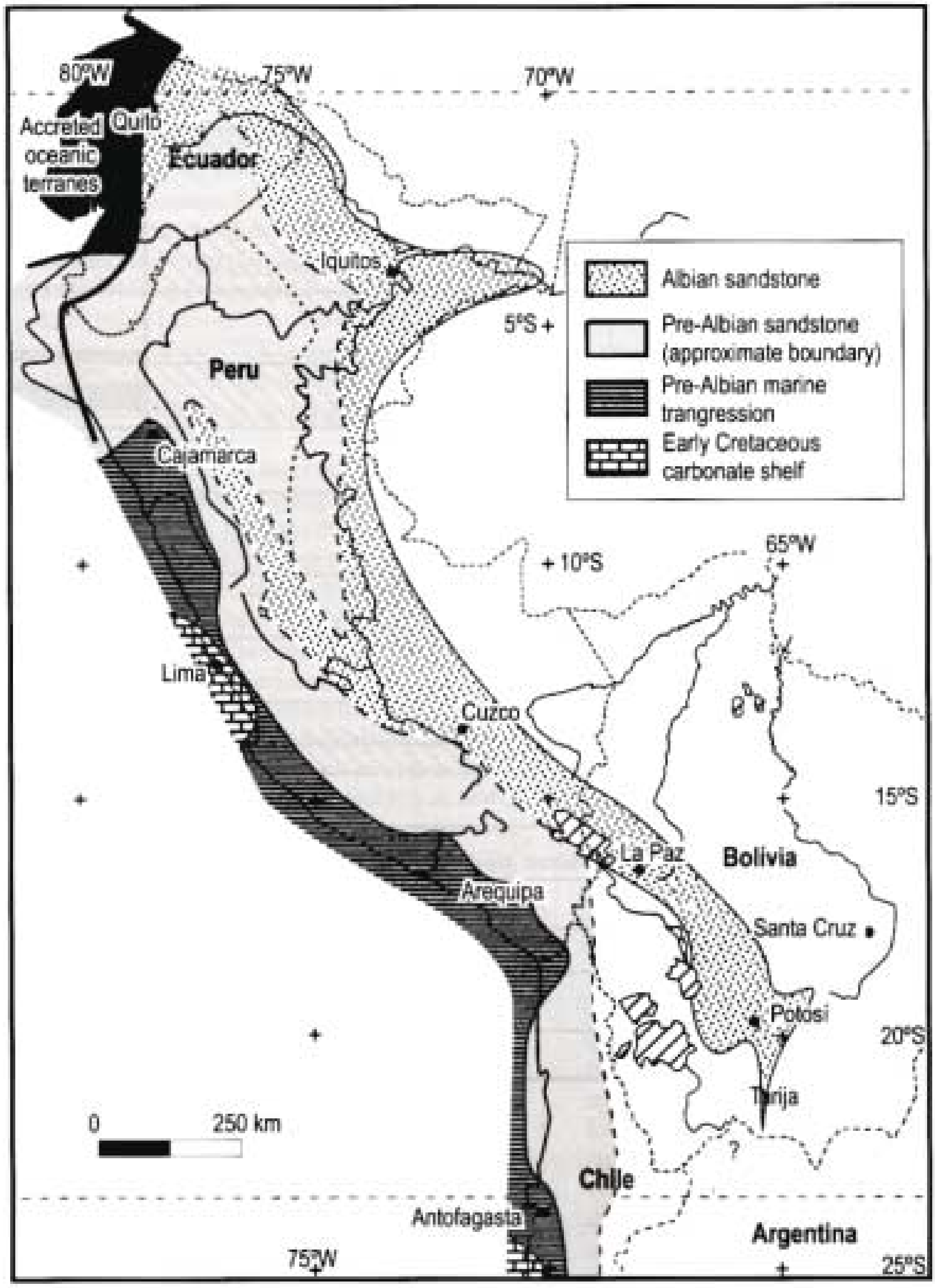


FIGURE 19 - Paleogeographic sketch of the Early Cretaceous transgression.

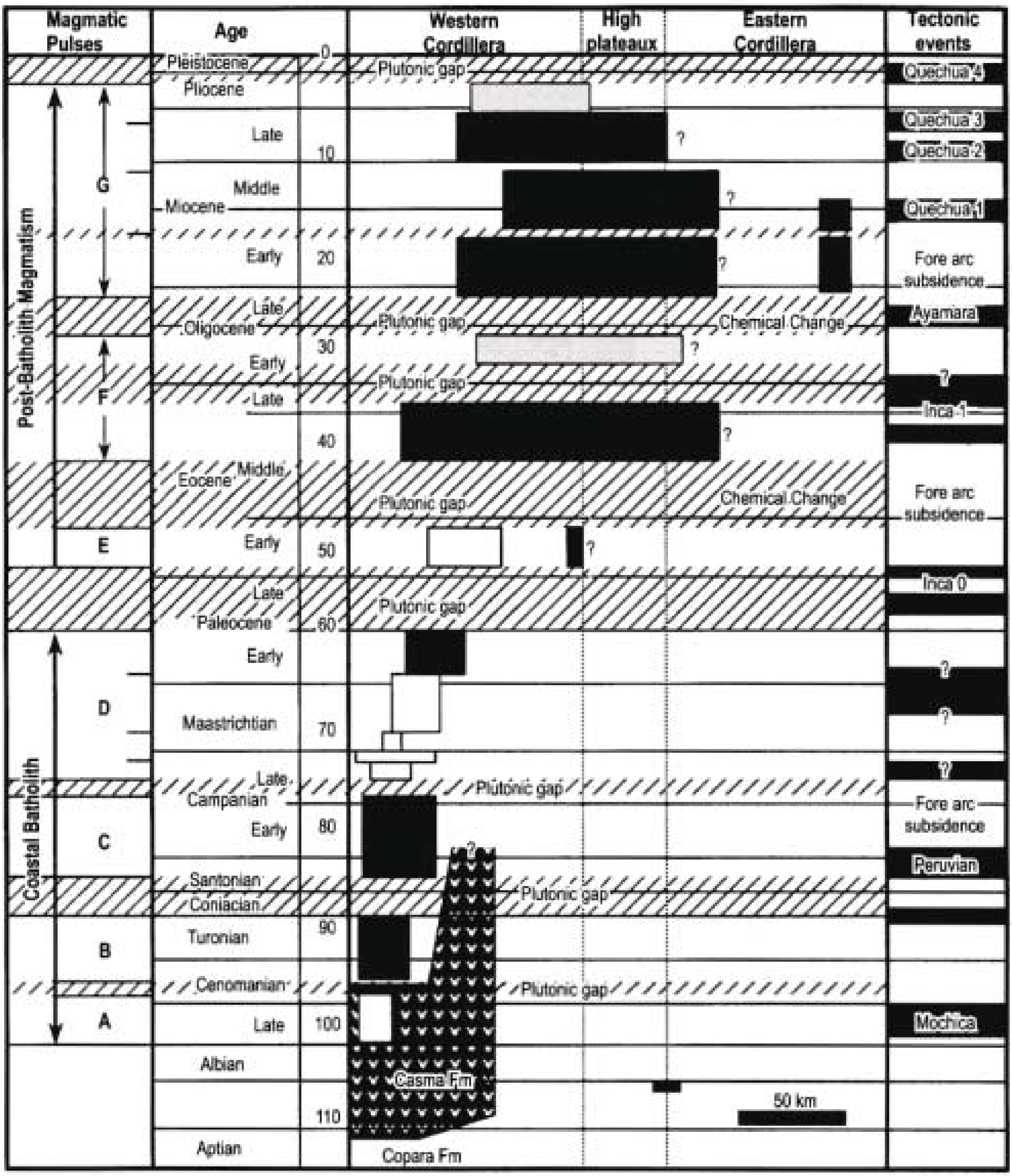


FIGURE 20 - Age, location and intensity of magmatic events in the magmatic arc of central Peru, and their relations to Andean tectonic events (after Soler, 1991). Light shaded areas: low magmatic activity, dark shaded areas: high magmatic activity.

cross-cut by basic to intermediate intrusions dated at 104 - 101 Ma (Wilson, 1975; Cobbing *et al.*, 1981; Bussel, 1983), thus indicating that compressional deformations began by Middle Albian times (Cobbing *et al.*, 1981).

Deformation was associated with significant dextral movements (Myers, 1975; Bussel and Pitcher, 1985). In the arc zones of Peru and Ecuador, the Late Albian tectonic compression is marked by local folding and faulting, by the end of marine sedimentation, by a general decrease of magmatic activity (Soler and Bonhomme, 1990), and by the replacement of volcanic effusions by plutonic intrusions, suggesting that the arc zones were significantly uplifted (Cobbing *et al.*, 1981; Soler, 1991). These calc-alkaline plutons, which intrude the volcanic arc, define the so-called Coastal Batholith of Peru (Pitcher, 1978; Cobbing *et al.*, 1981; Soler, 1991). Effusive magmatism ceased by earliest Cenomanian times, and incipient plutonic activity was rather low (Soler and Bonhomme, 1990), except in central-southern Peru where intrusions are dated at 101 - 94 Ma (Beckinsale *et al.*, 1985). The stability of the magmatic front in central Peru suggests that the Late Albian tectonic event did not change significantly the shape of the active margin (Soler and Bonhomme, 1990; Jaillard and Soler, 1996; Fig. 20). Most of the arc zone of Ecuador and Peru seems to have remained emergent during Cenomanian-Turonian times, since the Albian volcanic piles are usually unconformably capped by Santonian to Campanian transgressive sediments (Jaillard, 1995; Jaillard *et al.*, 1996). Plutonic activity was high during the Cenomanian (Beckinsale *et al.*, 1985; Mukasa, 1986, 94 - 90 Ma pulse of Soler and Bonhomme, 1990), but no intrusions of Turonian age are known (90 - 85 Ma plutonic gap, Soler, 1991).

In northern Chile, the locus of the magmatic arc significantly shifted eastward during Aptian-Albian times (Hammerschmidt *et al.*, 1992), and the Middle Cretaceous magmatic arc (115 - 90 Ma) is partly superimposed on the Jurassic arc (Charrier and Muñoz, 1994; Scheuber *et al.*, 1994; Fig. 15). In northernmost Chile, arc-related andesite, breccia, agglomerate, tuff, sandstone and conglomerate of Albian age overlie the Early Cretaceous deposits of the back-arc basin (Scheuber *et al.*, 1994). They yielded 115 - 104 Ma dates and are crosscut by 115 - 80 Ma intrusions (Bogdanic and Espinosa, 1994). An extensional or transtensional regime is assumed to have prevailed (Scheuber *et al.*, 1994). However, although no deformation of Albian age has been recognized, the significant eastward shift of the Middle Cretaceous arc (50 km) may result from a shortening event. Volcanic arc activity continued until early Late Cretaceous times (Scheuber *et al.*, 1994).

No arc-related magmatism is known on the continental margin N of 3°S (Ecuador). Therefore, the Peruvian subduction zone extended probably northwestwards into the oceanic domain by means of an intra-oceanic subduction zone, which gave way to the formation of island arcs of Albian to early Late Cretaceous age. This interpretation is supported by the occurrence, on the accreted oceanic terranes of Ecuador, of pre-Cenomanian island arc rocks (Las Orquídeas and Toachí units, Jaillard *et al.*, 1995; Benítez, 1995; Cosma *et al.*, 1998), overlain by volcaniclastic arc series of Cenomanian to Santonian age (Cayo and Pilatón formations, Faucher *et al.*, 1971; Kehrer and Van der Kaaden, 1979; Jaillard *et al.*, 1995). Note that, there too, volcanic activity seems to have ceased by Cenomanian times.

In the back-arc areas of Ecuador and Peru, the Aptian-Albian boundary is marked by scattered volcanic manifestations, varying from basaltic flows to rhyolithic tuff, intercalated within the first transgressive units. This bimodal volcanism as been locally determined as alkaline, indicating an intracontinental extension (Soler, 1989).

Earliest Albian times are marked by the beginning of a major large-scale marine transgression which reached its maximum extent by Turonian times (Figs. 13 and 21). The Lima area already received a marine sedimentation from Early Cretaceous times (Rivera *et al.*, 1975; Alemán, 1996). The Early Albian transgression first reached the western part of the back-arc areas (Benavides, 1956; Jaillard, 1995; Robert *et al.*, 1998), where it deposited red to yellow coloured silt and sandstone, with glauconitic and locally oolitic limestone in the upper part (Inca and Pariahuanca formations, Benavides, 1956; Wilson, 1963; Moulin, 1989).

The Albian transgression is then marked by three pulses of mid Early, early Middle and early Late Albian age (Robert *et al.*, 1998). The first pulse only reached the western part of the Eastern basins (Chulec Formation, Benavídes 1956; Wilson, 1963; Robert, in progress), whereas the third one reached locally the eastern border of the Eastern Basin of Ecuador (Basal Napo Shales) and northern Peru, where it may rest on Paleozoic rocks (Jaillard, 1997). This latter transgression probably reached also the axial swell of southern Peru and triggered the deposition of transgressive, partly marine sandstone (Huancané Formation, Carlotto *et al.*, 1995; Jaillard, 1995), which grade to the E into thicker deposits (lower part of Putina Group, Audebaud *et al.*, 1976; Jaillard, 1995). It could have reached also the Bolivian Potosí Basin where a few tens of metres of coarse-grained, locally conglomeratic sandstone are known (La Puerta Formation, Sempere, 1994). The early Middle and early Late Albian transgressive pulses are associated with widespread anoxic deposits in central and northern Peru and in Ecuador (Pariatambo, Basal Napo, Chonta formations, Villagómez *et al.*, 1996; Robert *et al.*, 1998). Late Albian times are then marked by the development of carbonate shelves in the western part (Yumagual Formation, part of Mujarrún, Jumasha and Arcurquina formations, Benavides, 1956, 1962; Wilson, 1963; Jaillard, 1987), and by the deposition of deltaic-fluvial sandstone in the eastern parts of the back-arc areas (Agua Caliente Formation, T sandstone, Putina Group), the progradation and retreat of which are mainly controlled by eustatic variations with a subordinate influence of tectonic events (Jaillard, 1994, 1997).

The effects of the Late Albian tectonic event are mild in the back-arc areas. The Late Albian shelf carbonate sedimentation exhibits slumps, syn-sedimentary faults and breccia, clastic dykes and differential subsidence, which together express an extensional regime (Audebaud, 1971; Jaillard, 1987, 1994). Farther to the E, progradation of deltaic systems may be regarded as the result of a slight uplift related to the Late Albian tectonic event (Jaillard, 1987, 1997). In northern Argentina, alkaline volcanic rocks (110 - 100 Ma) are interpreted as related to a rift episode (Viramonte *et al.*, 1999).

In the western part of the back-arc areas of Ecuador and Peru, the carbonate shelf sedimentation recorded significant eustatic transgression near the Albian-Cenomanian boundary, in Middle Cenomanian, early Late Cenomanian (widely characterized by *Neolobites vibrayeanus* (= *N. kummeli*), and Early Turonian times. Each transgression is marked by conspicuous marly levels which grade upwards into massive platform carbonate exhibiting frequently desiccation features, thus indicating a shallow marine environment (Jaillard, 1987, 1995, 1997; Fig. 22). In some southern parts of the Eastern Basin of Peru, the Albian-Cenomanian fluvio-marine sandstone beds (Oriente and Putina groups) are overlain by Early Turonian marine shale, illustrating the large extent of the Turonian transgression. In some areas, the upper part of the Turonian shelf limestone exhibits mild syn-sedimentary tectonic features announcing the tectonic event of the Turonian-Coniacian boundary (Jaillard, 1987, 1995, 1997).

Subsidence was intense in the western areas of northern Peru, and decreased drastically toward the NE and SE. The Albian-Turonian series reaches nearly 2000 m in the western part of northern Peru (Benavides, 1956; Wilson, 1963), about 300 m in the Oriente Basin of Ecuador (Jaillard, 1997), 600 m in southwestern Peru (Benavides, 1962; Jaillard, 1995), and 30 m in the Potosí Basin of Bolivia (Sempere, 1994). The axial swell continuously behaved as a positive area.

The Middle to Late Albian deformation was the first significant compressional deformation recorded in the Cretaceous evolution of the Andean margin, which affected mainly the fore-arc and arc zones. It coincided with a period of high convergence rate and with the opening of the South Atlantic Ocean at equatorial latitudes, which triggered the westward drift of the South American Plate and therefore, the trenchward motion of the overriding plate (Frutos, 1981; Jarrard, 1986; Soler and Bonhomme 1990; Jaillard and Soler, 1996). The strong dextral wrench component of this deformation (Bussel and Pitcher, 1985; Soler, 1991; Jaillard, 1994) resulted from the northeastward motion of the Farallón Plate, indicated also by the lack of any arc magmatism along the Colombian-Ecuadorian margin (Aspden *et al.*, 1987). Due to the oblique direction of the oceanic plate, convergence was accommodated by lateral displacement and wrenching deformation along the edge of the active margin, rather than by shortening of the whole margin. The resumption of volcanic activity along the Peruvian margin may be related to the beginning of the Middle Cretaceous period of high convergence rate (Soler, 1991).

Coniacian - late Paleocene (88 - 57 Ma)

This period is marked by a major change in the paleogeographic pattern, by the occurrence of compressional tectonic events, the intensity of which increased through time, and by the incipient eastward migration of the arc zone in Peru. A progressive but general marine regression, the arrival of fine-grained detrital deposits and the eastward shift of depocenters in the eastern basins marked sedimentation. Tectonic events are of Late Turonian-Early Coniacian (88 Ma), Santonian (85 Ma), Late Campanian? (80 - 75 Ma) and Late Maastrichtian age (70 - 65 Ma). Some of these events coincide with the accretion of oceanic terranes in Ecuador or northern Peru.

Turonian-Coniacian boundary event and Coniacian - Early Santonian evolution (88 - 85 Ma)

In the fore-arc Celica-Lancones Basin (northern Peru-southern Ecuador), the youngest fossil recovered from the turbidite series is of Early Conician age (Petersen, 1949; Jaillard *et al.*, 1999). A sedimentary gap occurred then during Late Coniacian and Santonian times. In the Talara fore-arc basin of northwestern Peru, no deposits are known between the Albian shelf carbonate and the Campanian transgressive marine deposits (González, 1976; Morales, 1993). No information is available on the other fore-arc zones. In the oceanic fragments accreted to Ecuador, the Turonian-Coniacian boundary roughly coincides with the beginning of the Cayo island-arc activity, as expressed by thick series of coarse-grained volcaniclastic turbidite beds (Cayo and Pilatón formations, Jaillard *et al.*, 1995; Benítez, 1995; Kehrer and Van der Kaaden, 1979).

The Albian volcanic arc series of southern Ecuador are unconformably capped by Early Campanian transgressive marine deposits (Naranjo Formation, Jaillard, 1997). In southern Peru, Albian volcanic rocks are unconformably capped by undated shelf limestone of Senonian age (Omoye Formation, Vicente, 1981; Jaillard, 1994). In the arc zones, the effects of the Late Albian and Turonian-Coniacian deformation are, therefore, indistinguishable. However, the deformational event seems to have occurred before the Early Campanian transgression of southern Ecuador. In the Coastal Batholith of Peru, significant wrench movements associated with a variable compressional regime has been recognized during Turonian-Coniacian times (Bussel and Pitcher, 1985). Plutonic intrusions are very scarce, and volcanic activity is unknown (90 - 85 Ma magmatic gap, Soler, 1991; Fig. 20). However, although the magmatic front appears to be nearly stable, a slight eastward shift of some kilometres can be detected (Soler and Bonhomme, 1990; Jaillard and Soler, 1996). In northern Chile, the magmatic arc is marked by a well-expressed magmatic gap between 90 and 80 Ma (Hammerschmidt *et al.*, 1992; Scheuber *et al.*, 1994), which followed (Hammerschmidt *et al.*, 1992) or was associated with the eastward shift of the magmatic front (Scheuber *et al.*, 1994).

In the western part of the back-arc areas of Ecuador, Peru and Bolivia, the end of the carbonate platform sedimentation marks the Turonian-Coniacian boundary. In the N, it is replaced by ammonite-rich marine shale with limestone interbeds (Celendín, Upper Napo, Upper Chonta formations, Tschopp, 1953; Benavides, 1956; Wilson, 1963; Jaillard, 1987, 1997; Figs. 21 and 22), whereas in the S, the Turonian limestone units are overlain by red shale and silt with abundant evaporite layers (Chilcane, Aroifilla formations, Vicente, 1981; Sempere *et al.*, 1997; Figs. 13 and 21). In Ecuador and northern Peru, two main marine transgressions are recognized, of Early Coniacian and Late Coniacian-Early Santonian age, respectively. They determine two thickening-upward progradational sequences, of Coniacian and Early Santonian age, respectively (Jaillard, 1997). The appearance of detrital quartz in the Coniacian sequence indicates the creation of new source areas. No Late Santonian fauna has been found so far in these sequences.

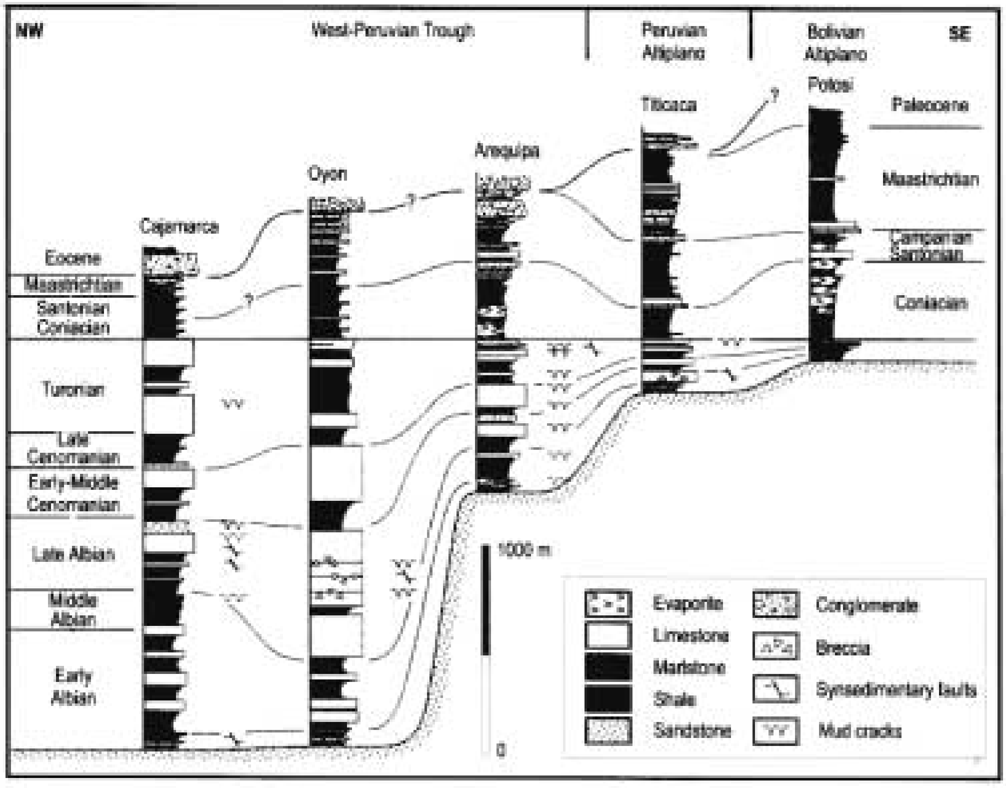
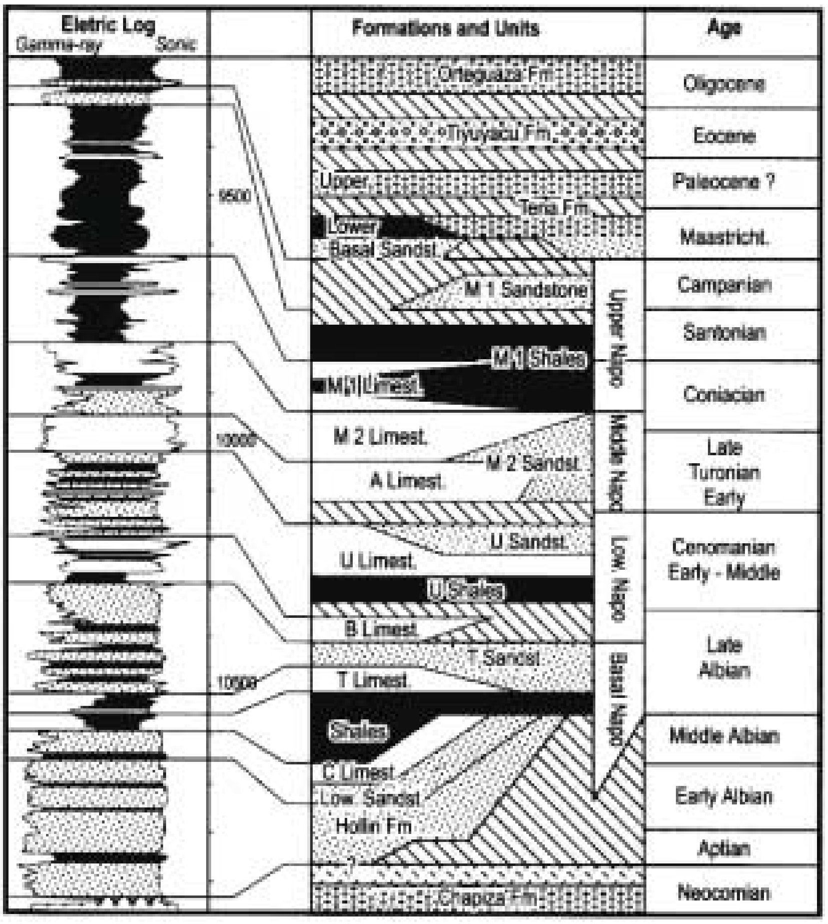
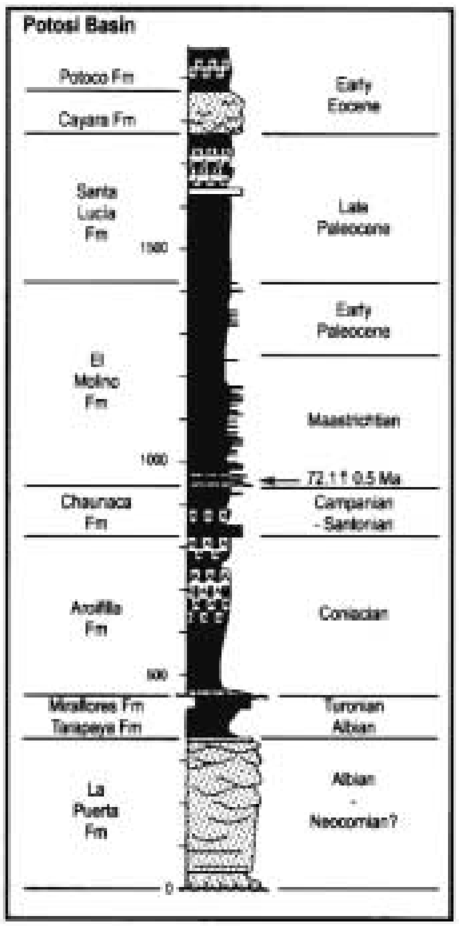


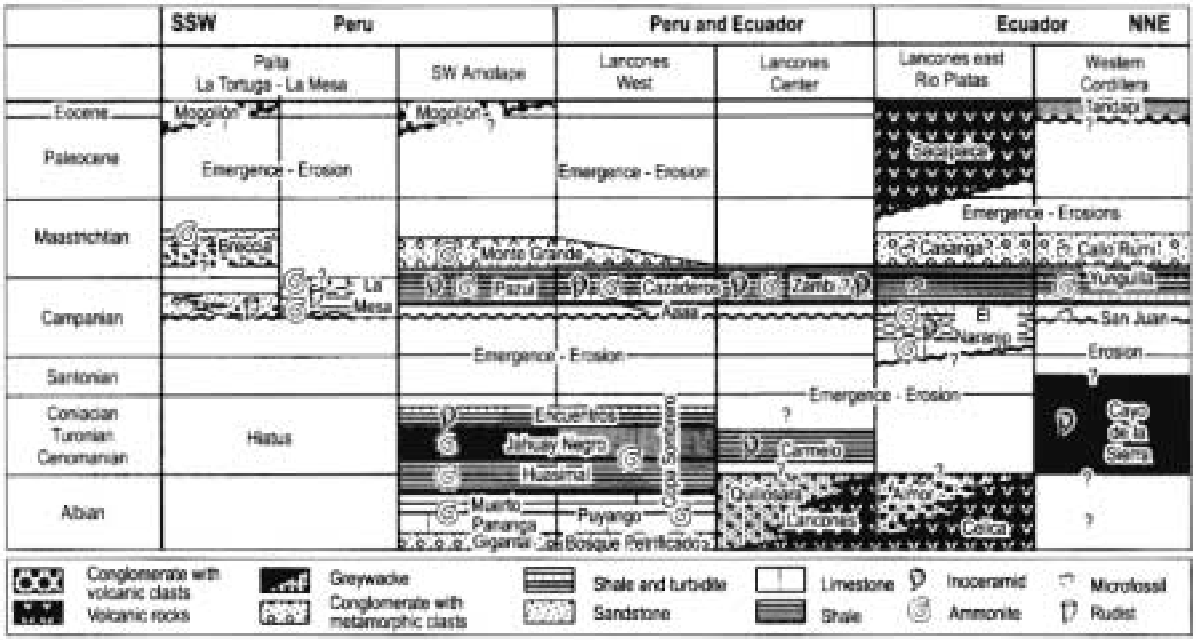
FIGURE 21 - Correlation of representative Albian-Eocene stratigraphic successions of Peru and Bolivia (after Jaillard and Soler, 1996).

FIGURE 22 - Representative well log and chronostratigraphic chart of the Cretaceous succession of the Oriente Basin (eastern Ecuador; after Jaillard, 1997).

FIGURE 23 - Representative section of the Cretaceous series of the Potosi Basin (central Bolivia; after Sempere et al., 1997).







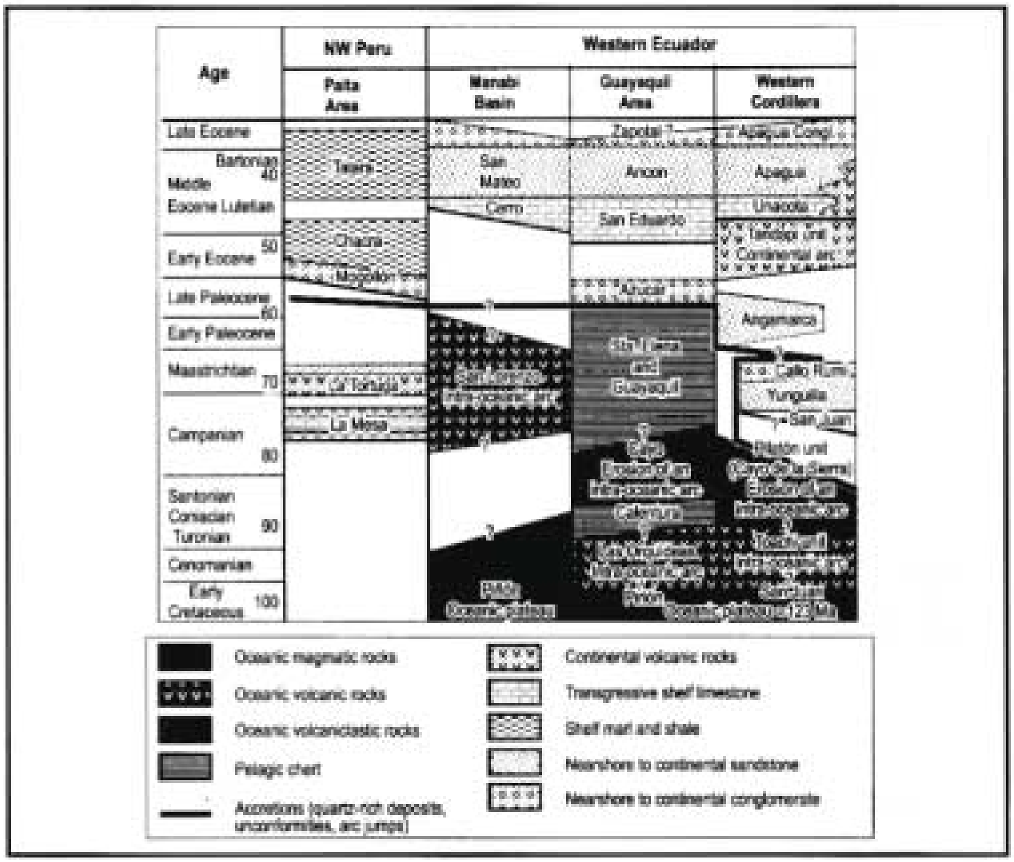


FIGURE 24 - Chronostratigraphic chart of the Middle Cretaceous to early Eocene sedimentary units of the fore-arc zones of Ecuador and northern Peru (after Jaillard et al., 1999).

FIGURE 25 - Chronostratigraphic chart of fore arc stratigraphic successions of northern Peru and western Ecuador. Probable accretion episodes are shown by the black line separating oceanic magmatic and sedimentary rocks (dark colours) from quartz-rich sediments and continental arc volcanism (light colours) (after Cosma et al., 1998).

Since the Coniacian to Early Santonian marine shale is generally overlain by transgressive sandstone of Campanian age, a sedimentary gap of Late Santonian-Early Campanian age is inferred. In southern Peru, and probably in Bolivia, the Late Coniacian-Early Santonian transgression is well marked and forms a thin marine layer (Middle Querque; Vicente, 1981; Jaillard, 1994), which is used as a correlation layer (Middle Vilquechico, Chaunaca formations, Jaillard *et al.*, 1993; Sempere *et al.*, 1997; Carlotto, 1998; Fig. 23).

In the back-arc areas, subsidence significantly increased during the Coniacian-Early Santonian time-span. In Bolivia, the spectacular increase in subsidence is regarded as the result of a foreland-type, flexural subsidence (Sempere, 1994; Sempere *et al.*, 1997; Fig. 23), due to significant tectonic shortening in the western areas (northern Chile). Extension prevailed, however, in these back-arc areas (Soler and Sempere, 1993). The occurrence of disoxigenated deposits (northern Peru, Ecuador) and of evaporites (southern Peru, Bolivia) suggests that the back-arc basin was separated from the open sea by an incipient morphological barrier, which was more pronounced to the S. In the whole area, Coniacian-Santonian marine deposits onlap eastwards onto the Guiana and Brazilian shields (Sempere, 1994). In eastern Ecuador, northeastern and southern Peru and in eastern Bolivia, the Coniacian-Santonian beds are the first Cretaceous marine shales to be deposited, and overlie the Albian (?)-Turonian fluvio-marine massive sandstone units (Jaillard, 1995, 1997; Sempere *et al.*, 1997). This significant eastward migration of the early Senonian depocenter is associated with a reorganization of the isopach maps, which become narrower and elongated parallel to the present-day chain, suggesting that the back-arc basins began to behave as distal foreland basins.

This Late Turonian-Early Coniacian paleogeographic reorganization is associated with local tectonic manifestations. In Bolivia, continental red beds unconformably overlie Middle Cretaceous marine strata (Vilcapujio event, Sempere, 1994) and in Ecuador (Jaillard, 1997), Coniacian silt or shale disconformably overlie the eroded Turonian limestone. In the Oriente Basin of Ecuador, the Late Turonian-Coniacian deposits exhibit significant thickness variations related to syn-sedimentary faulting of Late Turonian-Coniacian age (Christophoul *et al.*, 1999). In this area and in the Eastern Basin of northern Peru, mild compressional deformation has been recognized (Dashwood and Abbots, 1990; Gil *et al.*, 1996; Rivadeneira and Baby, 1999). This, together with the change in sedimentation and paleogeography, the slight retreat of the magmatic arc and the increase in subsidence, indicate that tectonic deformation and mild shortening affected the western areas.

Santonian-Early Campanian tectonic event and Campanian-Middle Maastrichtian evolution (85 - 68 Ma)

The Santonian (Early Campanian?) event is a major turning point in the Andean evolution, recognized a long time ago as the Peruvian phase (Steinmann, 1929). In the fore-arc Celica-Lancones-Basin (northern Peru-southern Ecuador), diachronous, latest Santonian to Middle Campanian transgressive shelf sediments unconformably overlie the deformed turbidite series of pre-Santonian age deposits (Jaillard *et al.*, 1997, 1999; Fig. 24). The compressional closure of the basin seems to be associated with the intrusion of syn-tectonic gabbro (Reyes and Caldas, 1987), locally dated at 82 Ma (Mourier, 1988). In the Talara fore-arc basin of northwestern Peru, the Albian shelf carbonate units are also covered disconformably by Campanian transgressive marine deposits (González, 1976; Macharé *et al.*, 1986; Séranne, 1987; Morales, 1993). In the fore-arc zone of Paita (northern Peru), Middle Campanian transgressive deposits rest unconformably on the Paleozoic basement (Bengtson and Jaillard, 1997; Jaillard *et al.*, 1999). Therefore, the main deformation of these fore-arc zones occurred during the latest Coniacian-Early Campanian time-span.

In the Celica-Lancones Basin, the middle Campanian transgressive beds are overlain by basinal dark shale interbedded with fine-grained turbidite beds of Late Campanian-Early Maastrichtian age (Jaillard *et al.*, 1999). South of Paita, the Middle Campanian transgressive sequence consists of transgressive marlstone and sandstone, rudist-bearing massive limestone, and transgressive marl and limestone grading upwards into sandstone and conglomerate, suggesting the occurrence of a Late Campanian tectonic event (La Mesa, Bengtson and Jaillard, 1997). Farther to the W (La Tortuga), the succession follows with a 3000 to 4000 m-thick series of alluvial to marine breccia, overlain by transgressive nearshore sandstone containing ammonites of Maastrichtian (probably Middle Maastrichtian) age (Bengtson and Jaillard, 1997; Fig. 24). These are unconformably overlain by latest Paleocene-early Eocene coarse-grained conglomerate, suggesting that a new tectonic event deformed this area in the Late Maastrichtian or Paleocene. No information is available about the other fore-arc zones.

In Santonian-Early Campanian times, the Ecuadorian margin underwent the accretion of an oceanic terrane constituted by an oceanic plateau dated at 123 ± 12 Ma (Lapierre *et al.*, 1999; Reynaud *et al.*, 1999) overlain by intra-oceanic island arc series (Fig. 25). This event is marked by a regional hiatus of Campanian age on the continental margin, by a significant thermal event which affected the Eastern Cordillera of Ecuador around 85 - 80 Ma (Cordillera Real, Litherland *et al.*, 1994) and by the abrupt arrival of disconformable quartz-rich turbidites of Late Campanian (?)-Maastrichtian age on the accreted oceanic series (Yunguilla Formation; Faucher *et al.*, 1971; Kehrer and Van der Kaaden, 1979; Cosma *et al.*, 1998). In the oceanic domain, the collision led to the end of the Middle Cretaceous island arc activity (Cayo Formation, Benítez, 1995), and to the onset, farther W, of a new island arc of Late Campanian-Maastrichtian age (San Lorenzo Formation, Lebrat *et al.*, 1987; Ordoñez, 1996). Since the accreted island arc series is locally dated as Coniacian in the Western Cordillera (Faucher *et al.*, 1971), the accretions occurred between the Coniacian and the Late Campanian. This arc jump expresses a significant reorganization of the intra-oceanic subduction zone geometry (Cosma *et al.*, 1998). The accreted oceanic terrane (Pallatanga unit, McCourt *et al.*, 1998), characterized by its association with the Yunguilla Formation, crops out presently along the eastern edge of the Western Cordillera of central and northern Ecuador (San Juan-Pujilí Suture, Juteau *et al.*, 1977; McCourt *et al.*, 1998).

The Early Maastrichtian Yunguilla Formation (Bristow and Hoffstetter, 1977) consists of alternations of basinal shale and medium-grained turbidite beds reworking volcaniclastic and siliciclastic material. These locally overlie units of transgressive limestone of Late Campanian-Maastrichtian age (Kehrer and Kehrer, 1969). This succession, comparable to that of the Celica-Lancones and Paita areas, indicates the creation of a wide fore-arc basin of Middle Campanian-Middle Maastrichtian age (Fig. 24), which extended at least from the Paita area (5°S) to N of Quito (0°).

The Albian volcanic arc series of southern Ecuador are unconformably capped by Late Santonian-Early Campanian transgressive marine deposits (Naranjo Formation, Jaillard *et al.*, 1997), which allow refining the age of the main tectonic event as pre-Campanian. In southern Peru, the Albian volcanics are unconformably capped by the undated Omoye Formation (Vicente, 1981), which has been ascribed to the Santonian (Jaillard, 1994), although it might be younger (Campanian?). In the arc zones, the effects of the Late Albian, Turonian-Coniacian and Late Santonian deformation are, therefore, indistinguishable. In both areas, the transgressive sequence grades into coarser-grained, locally conglomeratic, nearshore to continental deposits, dated in southern Ecuador as Maastrichtian (Cosanga Formation, Baudino, 1995; Jaillard, 1997), which indicate new tectonic movements in the Maastrichtian.

The Santonian-Early Campanian event coincided with the beginning of a significant retreat of the Coastal Batholith of Peru (Soler and Bonhomme, 1990; Fig. 20). This, together with the subsidence of the Late Campanian-Maastrichtian fore-arc basin of northern Peru-southern Ecuador, suggests that tectonic erosion began to act as a significant mass transfer process in the fore-arc and arc zones at that time (Jaillard and Soler, 1996; Jaillard, 1997). The Late Santonian-Early Campanian event is followed by a major plutonic pulse in the Coastal Batholith of central Peru, during which mainly granodiorite bodies were emplaced (85 - 77 Ma episode of Soler, 1991). A probable magmatic gap (77 - 74 Ma) might correspond to the Late Campanian event, and is followed by a new magmatic episode (74 - 69 Ma), which began with dyke swarm emplacement (Soler, 1991). In southern Peru, a plutonic gap (84 - 70 Ma) may coincide with the Santonian and Late Campanian events. The latter are responsible for the major NE-vergent Lluta Thrust, near Arequipa, which resulted in the thrust of Precambrian rocks onto Cretaceous sediments (Vicente *et al.*, 1982). Since it involves Coniacian-Early Santonian beds and is concealed by latest Cretaceous-early Paleogene unconformable conglomerate beds, it is of Late Cretaceous age (Vicente, 1989).

In northern Chile, magmatic activity resumed around 80 Ma ago (Early Campanian, Hammerschmidt *et al.*, 1992), and the location of the magmatic arc significantly shifted eastward, thus indicating that the Middle and Late Cretaceous tectonic events resulted in significant crustal shortening and/or crustal erosion (Scheuber *et al.*, 1994). This significant contractional event dated as 90 - 78 Ma, resulted in the folding, emergence and erosion of the arc zone, the tectonic inversion of the Domeyko Cordillera and creation of the retro-arc Purilactis Basin (Mpodozis and Ramos 1989; Scheuber *et al.*, 1994).

In the back-arc areas of Ecuador and northern Peru, Late Santonian-Early Campanian times are marked by a regional sedimentary gap (Tschopp, 1953; Benavides 1956; Seminario and Guizado 1976; Jaillard 1987, 1997; Mathalone and Montoya, 1995), which coincides with the accretion and related deformations recorded in the westerly zones. In northern Peru and eastern Ecuador, the Santonian marine deposits exhibit a thickening-upward evolution expressing the arrival of sandy detrital material regarded as related to the incoming Late Santonian tectonic movements. In southern Peru and Bolivia, stratigraphic data are insufficient to demonstrate the occurrence and duration of this hiatus in the mostly continental deposits (Middle Vilquechico, Middle Yuncaypata, Chaunaca formations; Sempere *et al.*, 1997; Jaillard *et al.*, 1993; Carlotto, 1998). The Early Santonian age of the last marine deposits in northern Peru and Ecuador, however, supports a Late Santonian age for the main deformational event.

Campanian times are then marked by a short-lived, regional marine transgression, locally dated as Middle Campanian (northern Peru, Mourier *et al.*, 1988), and therefore, probably coeval with the main transgression in the fore-arc zone (Fig. 24). In Ecuador and northeastern and central Peru, this transgression is associated with conspicuous disconformable transgressive sandstone (M-1 Sandstones, Lower Vivian Formation) overlain by a thin layer of marine shale (Augusto *et al.*, 1990; Salas, 1991; Mathalone and Montoya, 1995; Jaillard, 1997). In southern Peru and Bolivia, the Middle Campanian transgression is correlated with a thin layer of charophyte-bearing shale overlain by fine-grained red beds of presumed Late Campanian age (Middle Vilquechico, Middle Yuncaypata, Upper Chaunaca formations, Jaillard *et al.*, 1993; Sempere *et al.*, 1997; Figs. 23 and 27). The hiatus between Campanian and Maastrichtian deposits suggests the occurrence of a tectonic event in the Late Campanian, but an eustatic origin for this sedimentary gap cannot be ruled out. Mid-Campanian alkaline volcanic rocks (80 - 75 Ma) point to an extensional strain in northern Argentina (Viramonte *et al.*, 1999).

A new regional marine transgression occurred in the Early Maastrichtian, which deposited transgressive sandstone units grading upwards into marine shale, which rest disconformably on the Campanian beds. In Ecuador and northern and central Peru, these Early Maastrichtian marine layers are dated by marine microfossils and very scarce ammonites (Lower Tena, Upper Vivian, Areniscas de Azúcar formations, Koch and Blissenbach, 1962; Rodríguez and Chalco, 1975; Vargas, 1988; Mourier *et al.*, 1988; Jaillard, 1997). They are generally overlain by charophyte-bearing fine-grained continental red beds of Maastrichtian age. A disconformity separates these deposits from the overlying fine-grained, continental Paleocene red beds (Upper Tena, Yahuarango, Sol formations), suggesting the occurrence of tectonic movements near the Maastrichtian-Paleocene boundary (Mathalone and Montoya 1995; Jaillard, 1997; Christophoul, in progress).

In southern Peru and Bolivia, the Early Maastrichtian maximum flooding is marked by ephemeral marine

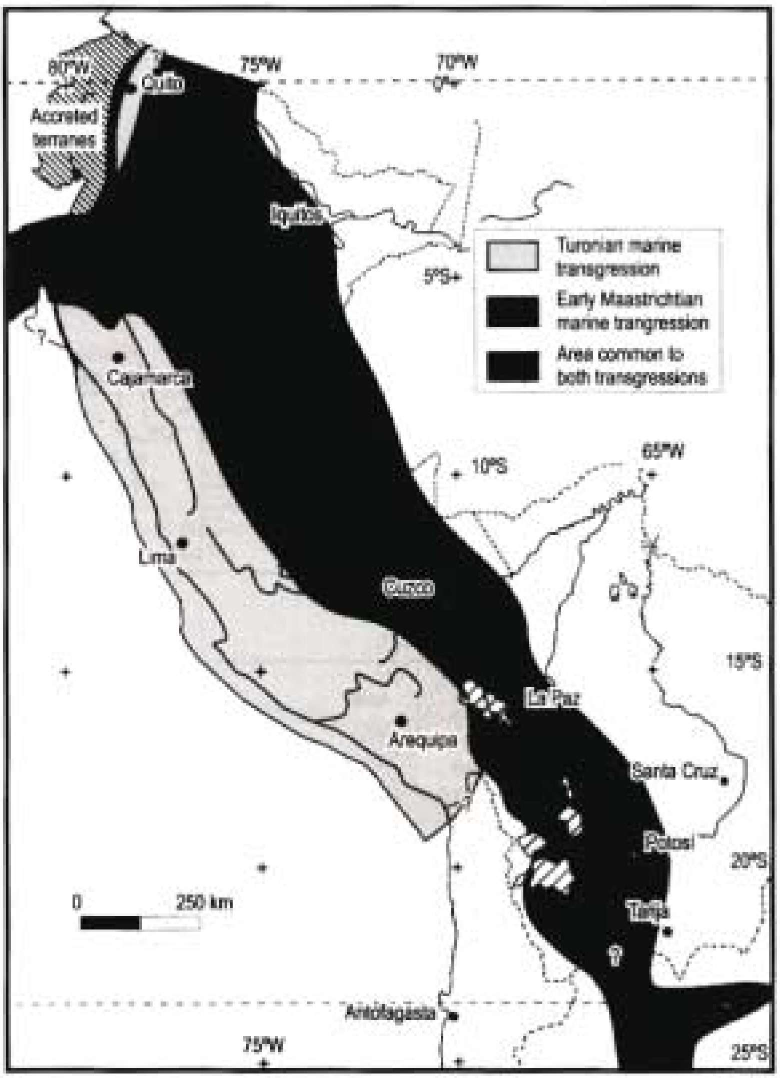
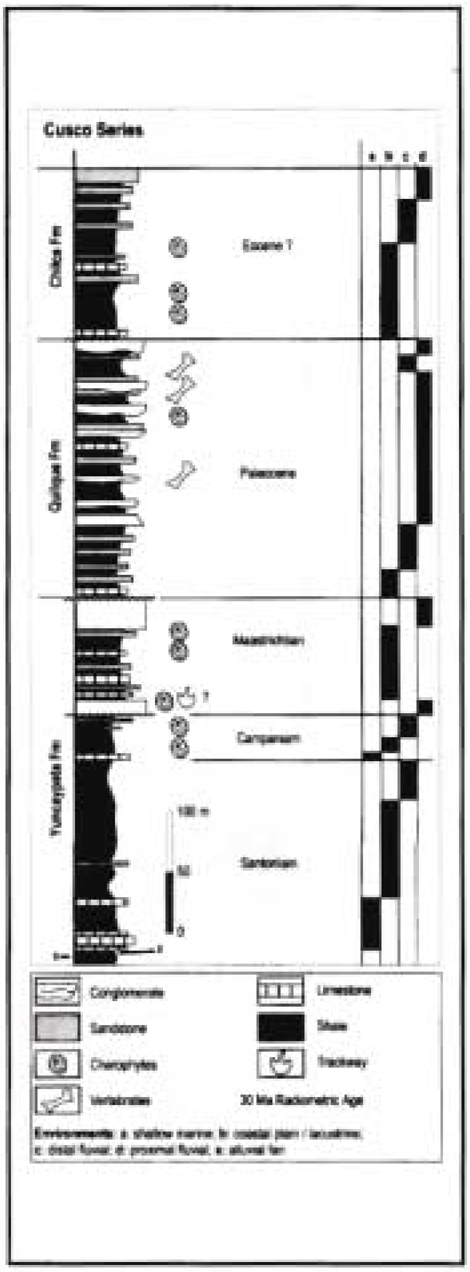


FIGURE 26 - Paleogeographic sketch showing the relative extensions of the Turonian and Maastrichtian marine transgressions.

FIGURE 27 - Late Cretaceous - Eocene stratigraphic succession in the Cusco area (Andes of southern Peru) (after Carlotto, 1998).



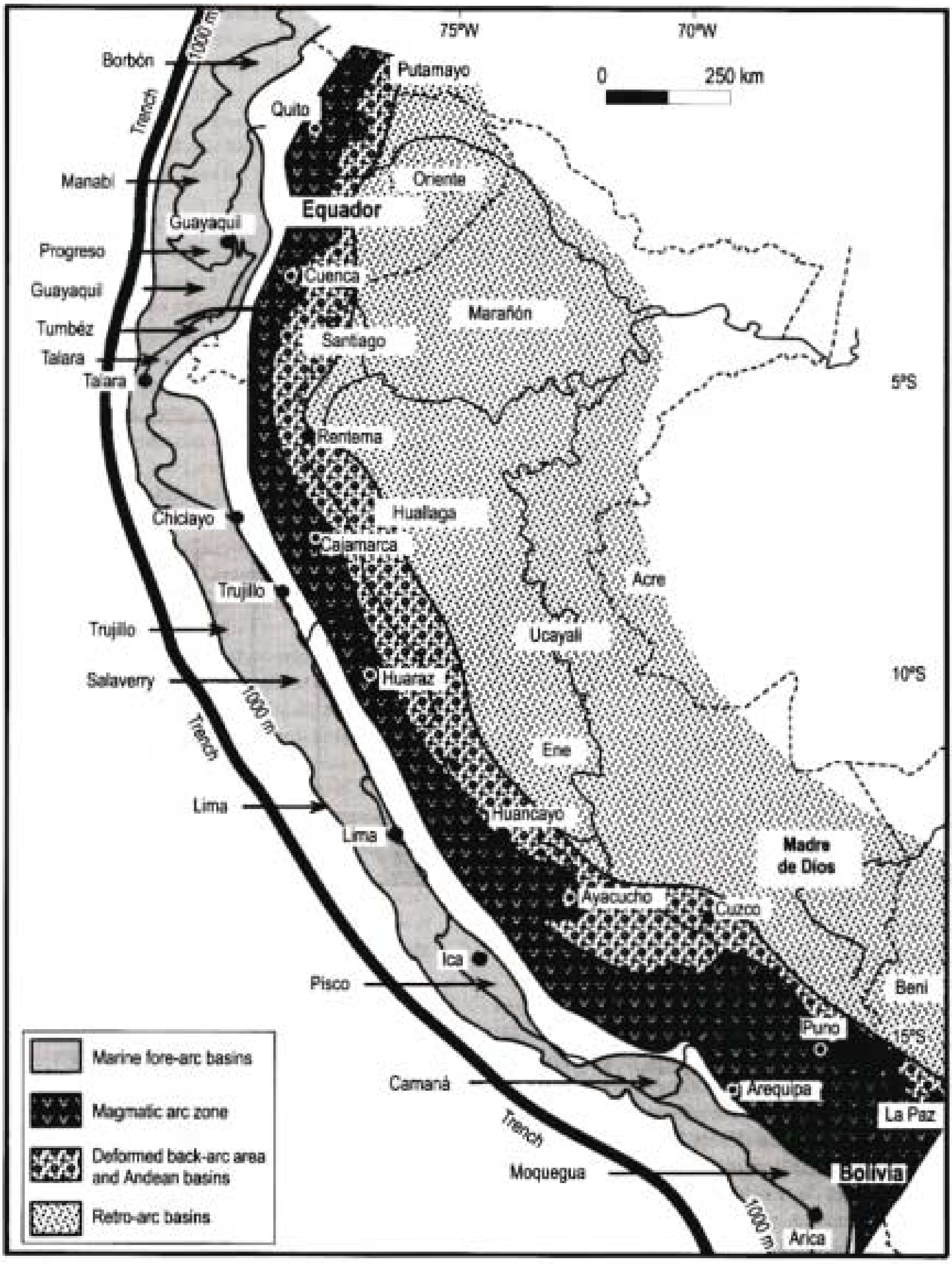


FIGURE 28 - Paleogeographic sketch of Ecuador and Peru for the Paleogene.

conditions (Jaillard *et al.*, 1993; Gayet *et al.*, 1993) and the Maastrichtian deposits significantly onlap onto the eastern border of the Cretaceous basin (Sempere *et al.*, 1997), thus expressing the gradual eastward shift of the depocenter during the Late Cretaceous (Fig. 26). Lacustrine conditions prevailed in part of the Bolivian-Argentine part of the basin (Rouchy *et al.*, 1993; Salfity and Marquillas, 1994). In southern Peru, an erosional disconformity separates the Maastrichtian and Tertiary beds (Jaillard *et al.*, 1993; Carlotto, 1998), but sedimentation is assumed to be continuous in Bolivia (Sempere *et al.*, 1997). The tectonic regime in the back-arc areas of northern Chile and Bolivia is assumed to have been extensional during the Maastrichtian, thus allowing sporadic shallow marine incursion and the outpouring of alkaline basic volcanics (Reyes *et al.*, 1976; Scheuber *et al.*, 1994; Viramonte *et al.*, 1999).

Late Maastrichtian - early late Paleocene (68 - 57 Ma)

The early to early late Paleocene sequence disconformably overlies latest Cretaceous strata. It consists of volcanic rocks (arc zone), and of fine-grained deposits, either marine (fore-arc zone) or continental (back-arc zone). In the arc and fore-arc zones, beside the regional hiatus of Late Maatrichtian-early Paleocene age and the frequent disconformities between Cretaceous and Tertiary beds, a Late Maastrichtian tectonic event is suggested by numerous intrusions (Mukasa, 1986), the emplacement of a centred complex (Mukasa and Tilton, 1985) and high strike-slip rates between 68 and 64 Ma in the Coastal Batholith of Peru (Bussel and Pitcher, 1985), and by numerous K/Ar resets indicating a thermal event at 70 - 60 Ma in the Cordillera Real of Ecuador (Litherland *et al.*, 1994).

Paleocene marine deposits are only known in the fore-arc Talara Basin of northwesternmost Peru (Iddings and Olsson, 1928; González, 1976; Zuñiga and Cruzado, 1979; Séranne, 1987; Morales, 1993; Fig. 29) and maybe in the offshore fore-arc zone of northern Peru, where the presence of pre-late Eocene sediments has been assumed locally (9°S, Kulm *et al.*, 1982). In the Talara Basin, the Late Cretaceous deposits are disconformably overlain by early Paleocene transgressive sandstone and conglomerate intercalated with marine grey shale, which grade upward into marine dark shale of middle Paleocene age (Weiss, 1955; Paredes, 1958; González, 1976). The upper unit overlaps to the E (Séranne, 1987), and is strongly eroded toward the S by pre-Eocene erosion (Paredes, 1958). To the E, thin continental to nearshore clastic facies grade westward into thick fine-grained deposits of relatively deep marine environment (Séranne, 1987). Farther to the S (Paita), the Cretaceous sedimentation ends up with the 3500 m thick breccia of the Maastrichtian La Tortuga Formation (Olsson, 1944; Fig. 24).

In the presently accreted oceanic terranes of Ecuador, since the San Lorenzo island-arc did not yield ages younger than Maastrichtian (Lebrat *et al.*, 1987; Ordoñez, 1996), its activity may have ceased by Paleocene times. Moreover, McCourt *et al.* (1998) recently identified an early to middle Paleocene quartz-rich turbidite series resting on accreted oceanic island arc series in the Western Cordillera of Ecuador. Since these early Paleocene turbidite beds are found to the W of the outcropping belt of the Yunguilla Formation, they might indicate that an other fragment of oceanic terrane has been accreted during the Late Maastrichtian and/or the earliest Paleocene. However, in some parts, the Maastrichtian-Paleocene sedimentation continued without any noticeable changes (Guayaquil Formation, Benítez, 1995; Keller *et al.*, 1997; Fig. 25).

In the magmatic arc of southwestern Ecuador, subaerial andesitic lava, breccia, agglomerate and acid tuff (Sacapalca Formation) are of latest Maastrichtian (67 Ma) to early Eocene age (Jaillard *et al.*, 1996; Hungerbühler, 1997; Pratt *et al.*, 1998), indicating that volcanic arc activity resumed after a gap that lasted from Late Albian times. In central and northern Ecuador, volcanic arc activity did not start before the early Eocene. In central and northern Peru, ring-complexes (68 - 64 Ma, Cobbing *et al.*, 1981; Soler, 1991) and calc-alkaline intrusions were emplaced (64 - 59 Ma, Cobbing *et al.*, 1981; Beckinsale *et al.*, 1985) and were possibly associated with coeval volcanism (Pararín Formation, Bussel, 1983). These intrusions are volumetrically important, indicating the local resumption of magmatic arc activity, but no significant compositional changes is noted with respect to the Late Cretaceous magmatism (Soler, 1991; Fig. 20). The 68 - 64 Ma period is also marked by important dextral wrench movements (Bussel, 1983; Bussel and Pitcher, 1985), which may be the expression of a Late Maastrichtian tectonic event. A magmatic gap then occurred during the late Paleocene (59 - 54 Ma, Soler, 1991). In southern Peru, plutonic intrusions began during latest Cretaceous times (78 Ma) and exhibit a major pulse during the early to middle Paleocene (62 - 57 Ma, Beckinsale *et al.*, 1985; Mukasa, 1986; Clark *et al.*, 1990). Associated volcanism (Toquepala Formation) consists of 3000 m of dacitic to rhyolitic tuff with minor andesitic intercalations, the composition of which suggests that the Andean crust was not thickened (Boily *et al.*, 1990). It is crosscut by gabbroic and granitic intrusions dated mainly at 66 - 63 Ma (Laughlin *et al.*, 1968; Vatin-Pérignon *et al.*, 1982; Mukasa and Tilton, 1985; Clark *et al.*, 1990). The resumption of arc magmatism recognised in Ecuador and Peru is expressed in northern Chile by abundant Late Maastrichtian-early Paleocene ages in the magmatic arc rocks (Hammerschmidt *et al.*, 1992; Charrier and Reutter, 1994), and by coeval volcanic intercalations in red beds deposited in proximal back-arc basins (Purilactis Group, 64 Ma, Flint *et al.*, 1993).

In the proximal back-arc zones of Peru (present-day Andes), Paleocene deposits unconformably rest on Late Cretaceous rocks (Noble *et al.*, 1990; Jaillard *et al.*, 1993; Mégard *et al.*, 1996), whereas the contact is only locally disconformable in the Eastern Basin (Vargas, 1988; Augusto *et al.*, 1990; Salas, 1991; Mathalone and Montoya, 1995; Gil *et al.*, 1996; Figs. 27 and 31). In addition, compressional deformation due to tectonic inversions near the Cretaceous-Tertiary boundary are common and widespread in the back-arc areas of eastern Ecuador (Dashwood and Abbots, 1990; Rivadeneira and Baby, 1999; Christophoul, in progress), northeastern Peru (Contreras *et al.*, 1996; Gil *et al.*, 1996) and Colombia (Cheilletz *et al.*, 1997).

In the distal back-arc areas of Ecuador and Peru, the Late Campanian-Maastrichtian sequence is overlain by a thick series of Paleocene fine-grained red beds (Upper Tena, Yahuarango, Sol 1, Quilque, Chilca, Santa Lucia formations, Kummel, 1948; Koch and Blissenbach, 1962; Mathalone and Montoya, 1995; Jaillard, 1997; Sempere *et al.*, 1997; Christophoul, in progress) which wedges out toward the W, mainly because of pre-Eocene erosion (Naeser *et al.*, 1991; Jaillard *et al.*, 1993; Carlotto, 1998). In the western zones, Paleocene deposits are usually lacking. However, in the Andes of Central Peru, a series of fluviatile red beds has been assigned to the Paleocene (Casapalca Formation, Jacay, 1994), although it may be younger. These Paleocene fine-grained red beds were deposited in wide, distal alluvial plains or in coastal setting. Clastic material proceeded from the smooth relief of the Paleo-Andes. Microfossils are dominated by charophyte associations (Gutierrez, 1982; Jaillard, 1994), but scarce foraminifera indicate local and sporadic marine influences (Koch and Blissenbach, 1962). In northeastern Peru, the Paleocene beds are disconformably overlain by transgressive conglomerate and marine to brackish beds of Eocene age (Pozo Formation). In Bolivia and northern Argentina, extensional conditions are marked by early Paleocene K-rich lava flows (65 - 60 Ma, Viramonte *et al.*, 1999).

The widespread hiatus, unconformities and detrital sedimentation, as well as the deformation and thermal event suggest that a significant, although poorly known, tectonic event occurred near the Maastrichtian-Paleocene boundary. This event might correspond to the accretion of an oceanic terrane, since part of western Ecuador received early Paleocene quartz-rich sedimentation, and recorded the end of the activity of an island arc (arc jump).

Late Paleocene - late Oligocene

This period corresponds to a transition between the pre-orogenic and the syn-orogenic periods. Compressional deformations became significant and involved the western parts of the back-arc areas, where marine sedimentation no longer occurred, except locally in Ecuador. The subsidence of fore-arc zones, which follows the compressional events, created sedimentary basins. Detrital sedimentation in the eastern area shows evidence for tectonically-induced disconformities. Finally, activity of the volcanic arc resumed, including along the Ecuadorian margin where arc magmatism was unknown since Late Jurassic times.

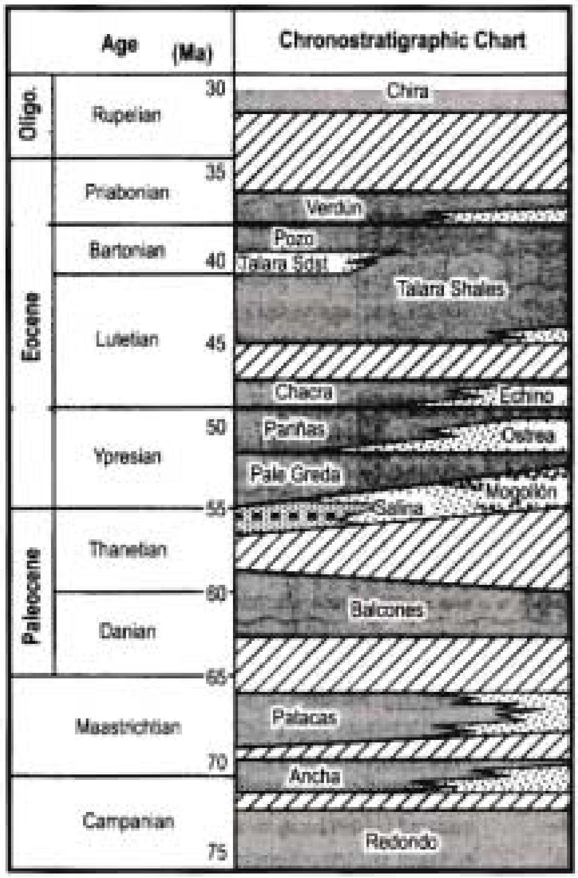
Because of the ongoing crustal shortening, eastward migration of the magmatic front, and uplift of the Andean domain, the paleogeographic pattern progressively changed during this period (Fig. 28). The fore-arc zones roughly correspond to the present-day coastal and offshore parts of the margin. Due to tectonic erosion, shortening, and/or flattening of the slab, the arc zone migrated and enlarged eastwards through time. It corresponded to the western part of the present-day Western Cordillera. The back-arc areas can be divided into a western, deformed and usually emergent area, also referred to as the "Paleo-Andes", and an eastern area, which still received sedimentation, and evolved through time toward a foreland retro-arc basin.

Late Paleocene event (58 - 55 Ma) and Eocene Sequence (55 - 40 Ma)

The late Paleocene event, first suspected by Cobbing *et al.* (1981) and Bussel and Pitcher (1985), is one of the major events in the Andean history (Marocco *et al.*, 1987; Noble *et al.*, 1990; Sempere *et al.*, 1997; Jaillard, 1997). It is coeval with an important plate kinematic reorganization in the Pacific realm, dated at 58 - 56 Ma, which resulted in a change in the convergence direction of the Farallón Plate. The latter changed direction from N or NNE to NE (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Atwater, 1989). The late Paleocene event is followed by the deposition of disconformable, well-identifiable sedimentary or volcanic sequences of early to early late Eocene age.

In the fore-arc zone of Ecuador, the collision of an oceanic terrane resulted in locally intense deformation of early late Paleocene chert (Santa Elena Formation) belonging to the accreted terrane (Benítez, 1995; Jaillard *et al.*, 1995). Since a thick series of quartz-rich coarse-grained turbidite beds of latest Paleocene age conceals the accretion, the latter occurred in the late Paleocene (Jaillard, 1997; Fig. 25). A further tectonic event of probable earliest Eocene age deformed the unconformable quartz-rich turbidite beds (Jaillard, 1997). In the Talara fore arc basin of northern Peru, sandstone and conglomerate (Basal Salinas) disconformably overlie Paleocene marine shale (Séranne, 1987) and grade southwards into diachronous, latest Paleocene to early Eocene, coarse-grained alluvial conglomerate (Mogollón Formation, Morales, 1993; Fig. 29). In the Paita area, the latter contain boulders of intra-oceanic origin, thus demonstrating that accretion occurred before the early Eocene in northern Peru. Since oceanic terranes are presently in Ecuador, they were subsequently displaced northwards along dextral faults, with a minimum estimate rate of 5 mm/y (Pecora *et al.*, 1999).

Due to the late Paleocene event, the fore-arc zones of Ecuador are marked by a widespread sedimentary hiatus encompassing most of the early Eocene (Benítez, 1995; Jaillard *et al.*, 1995). Sedimentation resumed diachronically since the end of the early Eocene. The Eocene sequence typically begins with breccia, slumped shale or transgressive peri-reefal limestone concealing fault-controlled relief. Diachronism and tectonic figures indicate an extensional tectonic regime, related to the subsidence that led to the deposition of the overlying thinning and shallowing-upwards sequence of marls interbedded with turbiditic sandstone (Jaillard *et al.*, 1995). The Eocene sequence ends up with disconformable, locally conglomeratic sandstone of nearshore to continental environment, dated as late middle to early late Eocene (Jaillard *et al.*, 1995). These indicate the beginning of the late Eocene tectonic movements (Jaillard, 1997). A comparable sequence is known in most of the coastal area (Benítez, 1995; Jaillard *et al.*, 1995) and the Western Cordillera (Santos *et al.*, 1986; Egüez, 1986; Bourgois *et al.*, 1990). This suggests that most of western Ecuador underwent a similar sedimentary evolution during the middle Eocene, and that most of oceanic terranes were probably already accreted to the continental margin by middle Eocene times (Jaillard, 1997; Cosma *et al.*, 1998). In the Western Cordillera, however, an island arc unit (Macuchi



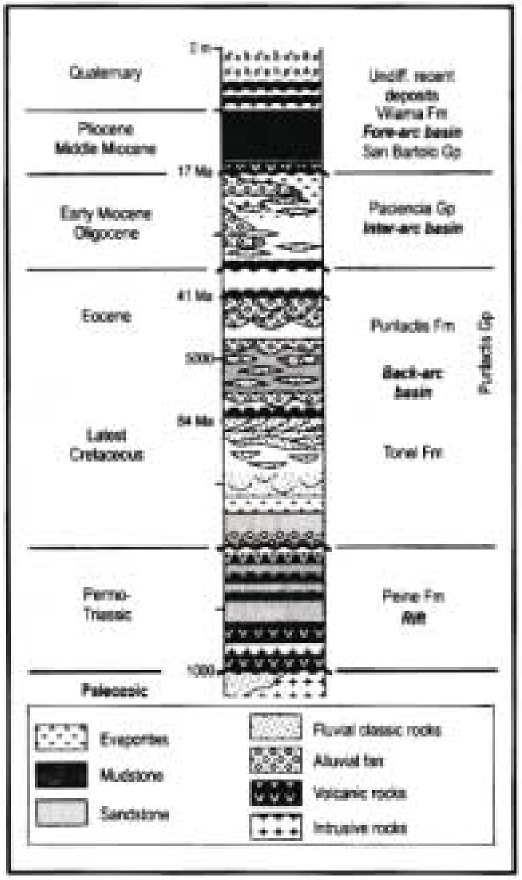
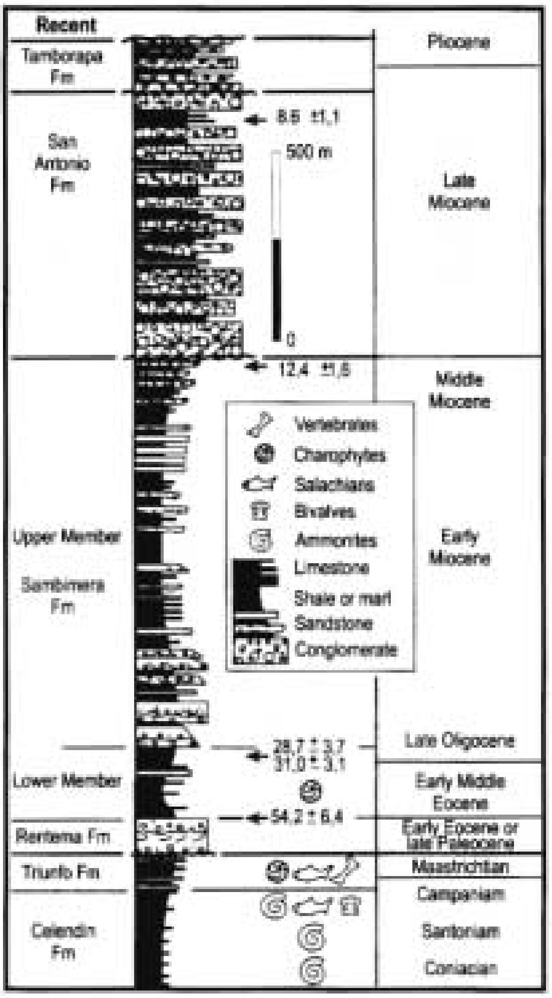


FIGURE 29 - *Chronostratigraphic chart of the Late Cretaceous-Paleogene Talara Basin (forearc zone of northern Peru) (after Morales, 1993).*

FIGURE 30 - *Mesozoic-Quaternary stratigraphic succession in the Atacama area (northern Chile) (after Flint et al., 1993). Due to the eastward migration of the arc zone, this area evolved from a back-arc to a fore-arc setting between latest Cretaceous and Pliocene times.*

FIGURE 31 - *Late Cretaceous-Pliocene stratigraphic succession in the Bagua area (northern Peru) (after Naeser et al., 1991).*



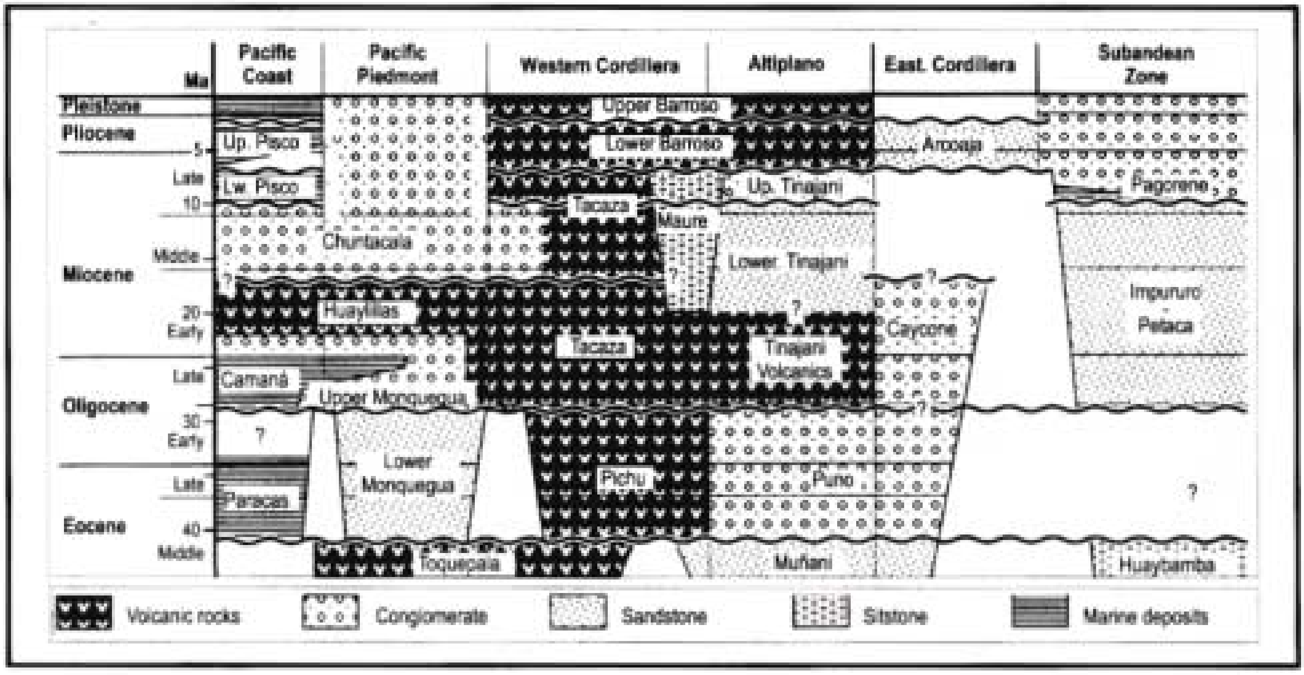


FIGURE 32 - Chronologic chart of the sedimentary and volcanic successions and deformational events in southern Peru (after Sébrier et al., 1988)

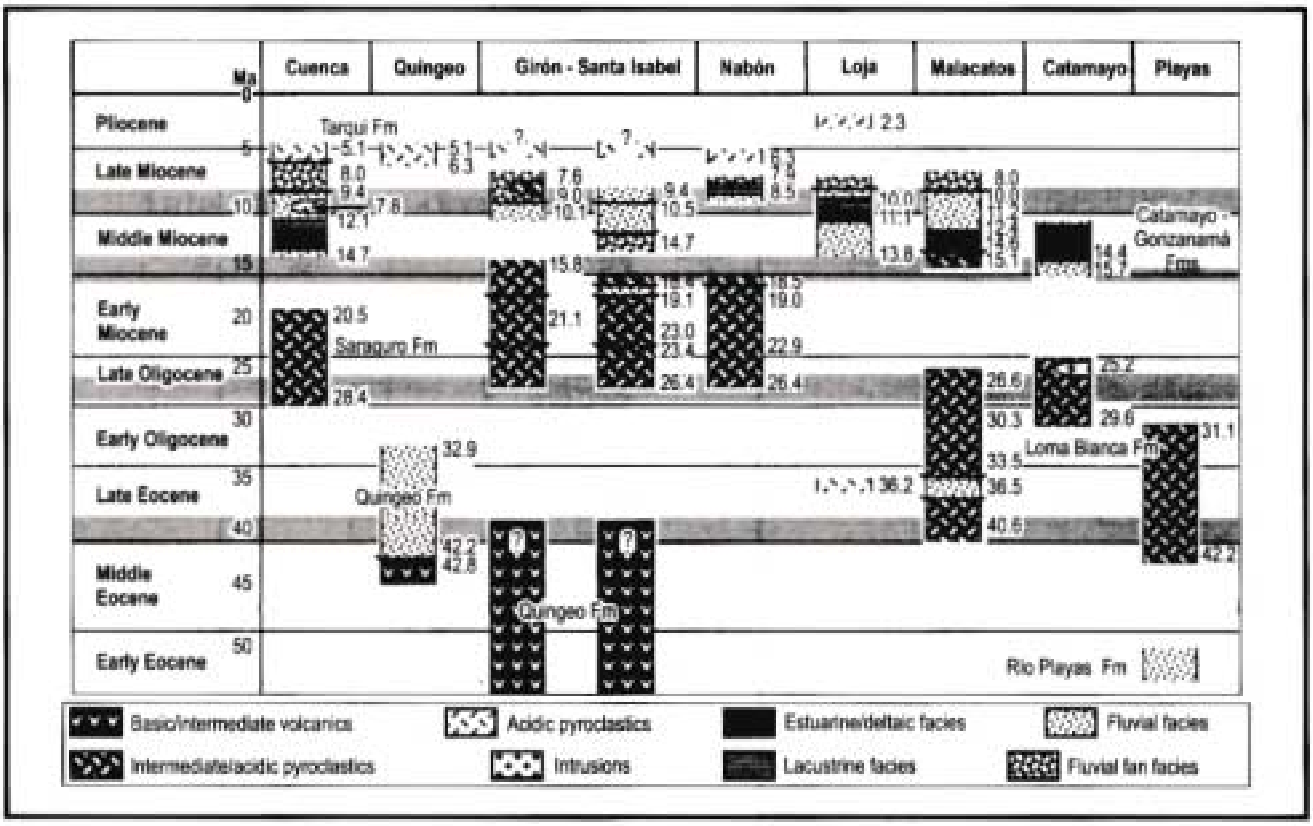


FIGURE 33 - Chronostratigraphic chart of the volcanic and sedimentary successions in the Andes of southern and Central Ecuador (after Hungerbühler, 1997, Steinmann et al., 1999).

Sandstone) seems to be associated with Eocene pelagic chert, and was therefore not yet in contact with the Andean margin (Hughes and Pilatasig, 1999).

In the fore-arc zone of Peru, although early Eocene deposits have been locally mentioned (Kulm *et al.*, 1982, Suess *et al.*, 1988), deposits of that age are well known only in the Talara Basin (Fig. 29). There, the very thick Eocene series comprises three main sedimentary sequences limited by disconformities of earliest Eocene, latest early to early middle Eocene, and late middle Eocene age, respectively (González, 1976; Séranne, 1987; Morales, 1993). The diachronous transgression begins with disconformable marine sandstone and conglomerate of latest Paleocene to early Eocene age, from W to E (Morales, 1993), the base of which contains boulders of oceanic origin, which post-date the accretion of oceanic terranes (Pécora *et al.*, 1999). The Eocene sequences generally consist of marine shale and fine to coarse-grained sandstone, which grade laterally E or NE into coarse-grained continental sandstone and conglomerate (González, 1976; Séranne, 1987). Pre-middle Eocene erosions removed part of the early Eocene sequence, and the second sequence locally rests on the basal conglomerate (Paredes, 1958). The upper part of the late middle Eocene sequence exhibits compressional syn-sedimentary deformations related to a transpressional regime (Séranne, 1987; Becerra *et al.*, 1990). The third sequence (late middle to early late Eocene) is made of unconformable massive coarse-grained sandstone of nearshore environment (González, 1976; Morales, 1993), announcing the late Eocene tectonic event. Overlying open marine shale beds are ascribed to the late Eocene (González, 1976) or the early Oligocene (Morales, 1993).

Near the early-middle Eocene boundary, subsidence of the fore-arc zone triggered the creation of a new generation of fore-arc basins, where thick middle to early late Eocene deposits unconformably rest upon Paleozoic to early Eocene units. In the offshore fore-arc basins of central and northern Peru, drillholes crosscut as much as 2000 m of middle Eocene shale, siltstone and sandstone of shallow marine, high-energy environment (Ballesteros *et al.*, 1988; Suess *et al.*, 1988), which usually unconformably overlies Cretaceous to Paleozoic rocks (Macharé *et al.*, 1986). It ends up locally with breccia, suggesting a tectonic instability of late middle Eocene to late Eocene age. Subsidence and basin tectonics is controlled by NNE trending normal faults (Macharé *et al.*, 1986; Azálgara *et al.*, 1991). The middle Eocene series is commonly directly overlain by Miocene marine sediments, thus providing evidence for a widespread sedimentary hiatus encompassing late Eocene, Oligocene, and often early Miocene times (Suess *et al.*, 1988; Von Huene *et al.*, 1988).

In the arc zone of southern Ecuador, subaerial arc volcanism and associated continental volcaniclastic sedimentation occurred during the late Paleocene and early Eocene (Sacapalca Formation, Jaillard, 1997; Hungerbühler, 1997). In the rest of Ecuador, resumption of arc volcanism is dated as early or middle Eocene (53 - 45 Ma, Egüez 1986; Wallrabe-Adams, 1990; Van Thournout *et al.*, 1990; Steinmann, 1997; Dunkley and Gaibor, 1998). It is commonly associated with subaerial volcaniclastic red beds of middle Eocene age deposited in proximal back-arc basins (Silante Formation, Wallrabe-Adams, 1990; Hughes and Pilatasig, 1999; Quingeo Formation, Steinmann, 1997; Playas Formation, Hungerbühler, 1997; Fig. 33). The renewal of arc activity along the Ecuadorian margin is due to the more easterly convergence direction, subsequent to the late Paleocene plate tectonic re-organization (Pilger, 1984; Pardo-Casas and Molnar, 1987).

In northern Peru, thick calc-alkaline subaerial volcanic series are dated between 55 and 40 Ma (Laughlin *et al.*, 1968; Cobbing *et al.*, 1981; Noble *et al.*, 1990; Soler, 1991), and post-date compressional deformation (Cobbing *et al.*, 1981; Bussel, 1983; Bussel and Pitcher, 1985). Although the chemical signature did not change with respect to the Paleocene magmatic arc (Soler, 1991), the Eocene magmatic activity is marked by a slight eastward shift of the western magmatic front, a decrease of plutonic intrusions in the Coastal Batholith and the beginning of the enlargement of the magmatic arc, which reached the present-day Western Cordillera (Noble *et al.*, 1990; Soler, 1991; Fig. 20). In contrast, a magmatic gap occurred in southern Peru during the early Eocene (Mukasa, 1986; Boily *et al.*, 1990; Clark *et al.*, 1990; Soler, 1991). However, subvolcanic stocks and associated porphyry copper deposits were emplaced in the Toquepala prospect between 57 and 52 Ma (Sébrier *et al.*, 1988; Clark *et al.*, 1990). Farther to the E (Inner Arc), the enlargement of the magmatic arc is marked by the emplacement of several calc-alkaline plutons, among which the large is the Andahuaylas-Yauri Batholith (48 - 34 Ma, Carlier *et al.*, 1996).

In northern Chile, the magmatic arc drastically shifted eastward between 55 and 48 Ma (Hammerschmidt *et al.*, 1992; Scheuber *et al.*, 1994), thus suggesting that significant crustal shortening and/or tectonic erosion occurred in the early Eocene (Fig. 15). This event has been ascribed to the late Eocene tectonic phase (Scheuber *et al.*, 1994), but is better correlated with the late Paleocene event. The latter is followed by a significant resumption of the magmatic activity along the whole margin. In northern Chile, the resumption of arc magmatism is expressed by numerous volcanic outcrops dated as middle Eocene (48 - 38 Ma, Hammerschmidt *et al.*, 1992). Products of this arc were deposited in a proximal, extensional back-arc basin, which received a thick, coarsening-upward pile of mainly volcaniclastic rocks, which may rest conformable on Late Cretaceous sediments or unconformably on older rocks (Hartley *et al.*, 1992; Flint *et al.*, 1993; Purilactis Formation of Charrier and Reutter, 1994; Fig. 30). Tectonic regime in the back-arc zone is no longer extensional (Scheuber *et al.*, 1994).

In the western back-arc areas of Ecuador (Cordillera Real), K/Ar age resets near 65 - 50 Ma indicate the occurrence of a noticeable thermal event (Aspden and Litherland, 1992; Litherland *et al.*, 1994), related to the late Paleocene tectonic event. In the Andes of northern Peru, early Eocene volcanic rocks dated at 55 to 50 Ma unconformably rest on deformed Cretaceous sediments (Cobbing *et al.*, 1981; Bussel, 1983; Noble *et al.*, 1990; Soler, 1991). In the Andes of central Peru, continental red beds bearing Paleocene-Eocene charophytes unconformably rest on Cretaceous sediments (Mégard, 1978; Mégard *et al.*, 1996).

In the present-day Andes of central Peru, unconformable continental red beds are locally dated by latest Maastrichtian charophytes, whereas in other parts, apparently similar, but conformable, red beds yielded late

Eocene-early Oligocene charophyte oogons and 40 to 37 Ma K/Ar dates (Mégard *et al.*, 1996). Therefore, some red beds may be of late Paleocene-middle Eocene age, but their characteristics and extension are unknown so far. Undated fluviatile red beds (Casapalca Formation) overlying the latest Cretaceous strata have been ascribed to the Paleocene (Jacay, 1994), although they may be younger. In southern Peru, Paleocene fluviatile red beds (Quilque Formation) are disconformably overlain by lacustrine deposits (Chilca Formation, Jaillard *et al.*, 1993; Carlotto, 1998), possibly of Eocene age.

The eastern back-arc areas are marked by a regional unconformity below massive coarse-grained sandstone and conglomerate of early Eocene age (Tiyuyacu Formation of Ecuador, Dashwood and Abbots, 1990; Benítez *et al.*, 1993; Jaillard, 1997; Rivadeneira and Baby, 1999; Rentema and Basal Pozo formations of Peru, Naeser *et al.*, 1991; Robertson Research, 1990; Cayara Formation of Bolivia, Sempere, 1994; Sempere *et al.*, 1997; Figs. 13, 23, and 31). They often post-date a sedimentary gap, which encompasses a large part of the Paleocene. Moreover, no early Eocene deposits have been accurately dated so far in the back-arc areas of Ecuador and Peru. Early to middle Eocene times are then marked by a regional marine transgression.

In Ecuador, the unconformable Lower Tiyuyacu Formation is overlain by marine to brackish beds of Eocene age (Benítez *et al.*, 1993). In eastern Peru, the early Eocene basal transgressive lag is overlain by a marine to brackish fine-grained layer of Eocene age and by coarsening-upward lacustrine to fluviatile red beds of middle to late Eocene age (Pozo Formation, Kummel, 1948; Müller and Aliaga, 1981; Robertson Research, 1990; upper part of Sol 3 Formation, Koch and Blissenbach, 1962; Gutierrez, 1982). The Ucayali Basin seems to be marked by a sedimentary hiatus of early Eocene age (Koch and Blissenbach, 1962). On the western border of the basin (Rentema), conglomerate beds dated at 54 Ma are conformably overlain by early to middle Eocene lacustrine deposits, equivalent to the Pozo Formation (Naeser *et al.*, 1991; Jaillard, 1994). Comparable lacustrine deposits of Eocene age are known in the Altiplano of southern Peru (Chilca Formation, Carlotto, 1998) and locally in Bolivia (Cayara Formation, Sempere *et al.*, 1997). Probably due to subsequent erosions, the overlying late Eocene succession is frequently lacking on the borders of the basin (Ecuador - Subandean Zone; Peru - Rentema, Ucayali Basin).

Middle-late Eocene event (40 - 35 Ma) and Oligocene evolution (35 - 28 Ma)

The late Eocene event has long been recognized in the Andes of Peru (Incaic phase, Steinmann, 1929), and has been further documented on the basis of radiometric data. It is followed by the deposition of unconformable beds of latest Eocene-middle Oligocene age. Sedimentation is chiefly tectonically driven in the eastern intermontane and foreland continental basins. This period ended with the late Oligocene Aymara tectonic event (28 - 26 Ma, Sébrier *et al.*, 1988; Sempere *et al.*, 1990; Fig. 32).

In the fore-arc basins, the late Eocene tectonic event was announced by the deposition of late middle Eocene disconformable coarse-grained deposits (42 - 40 Ma). It culminated in the late Eocene (37 - 35 Ma) with the deformation and emergence of many external fore-arc basins (Macharé *et al.*, 1986; Séranne, 1987; Ballesteros *et al.*, 1988; Jaillard *et al.*, 1995; figs. 25 and 29). This event, together with the late Oligocene crisis, is responsible for a widespread Oligocene hiatus in the fore-arc zone. However, sedimentation occured in a few basins (Talara, locally), and some internal (eastern) fore-arc basins were affected by significant subsidence, which allowed the deposition of late Eocene to early Oligocene marine (Pisco) or continental sequences (Moquegua) (Macharé *et al.*, 1988; De Vries, 1998). This subsidence pulse announced the accelerated subsidence, related to tectonic erosion processes, which affected the Andean fore-arc zones from the Eocene (Suess *et al.*, 1988; Bourgois *et al.*, 1990; Von Huene and Scholl, 1991).

In Ecuador, undated coarse-grained conglomerate beds of fan-delta environment, which unconformably overlie the Eocene sequence may be ascribed either to the latest Eocene-early Oligocene (Jaillard *et al.*, 1995; Fig. 25), or to the late Oligocene (Benítez, 1995). Marine shale, siltstone and fine-grained sandstone are dated as middle Oligocene (Playa Rica Formation, Benítez, 1995). They rest disconformably on the Eocene sequence and are separated from the Miocene deposits by a sedimentary hiatus (Benítez, 1995). In northern Peru (Talara Basin), the late middle Eocene sandstone beds are overlain by pelagic shale of debated, possibly early Oligocene, age (Chira Formation, Morales, 1993). In southern Central Peru (Pisco Basin), 600 m of transgressive shale, siltstone and subordinate sandstone of intertidal to nearshore environments are regarded as of late Eocene, maybe early Oligocene (?), age (Paracas Formation, Newell, 1956; Marocco and De Muizon, 1988; Macharé *et al.*, 1988). A middle Oligocene marine sequence has been recently described (De Vries, 1998), which probably correlates with the Oligocene beds of Ecuador (and northern Peru?). In Southern Peru (Moquegua), transgressive fan conglomerate, fluvial sequences and evaporite-bearing lacustrine silt and shale are ascribed to the Eocene, and infill an extensional, fault-controlled basin, probably created after the late Eocene event (Marocco *et al.*, 1985). As for many fore-arc basins of Peru, these beds unconformably overlie Precambrian to Mesozoic rocks deformed by the Late Cretaceous to late Eocene tectonic phases.

In the arc zone of central Ecuador (Cuenca), volcanic rocks of early middle Eocene age (43 Ma, Steinmann, 1997) are overlain by a 1000 m thick series of fluvial conglomerate and sandstone beds of late middle to late Eocene age (Quingeo Formation, 42 - 35 Ma, Steinmann, 1997; Fig. 33). In central and northern Peru, intrusions in the Coastal Batholith ceased by latest Eocene times (35 Ma, Beckinsale *et al.*, 1985; Mukasa and Tilton, 1985; Soler, 1991). With respect to the Cretaceous-Paleocene intrusions, the late Eocene-recent arc magmatism exhibits significant geochemical changes, regarded as resulting from the late Eocene tectonic event (Soler, 1991; Fig. 20). In the present-day Andes of northern and central Peru, the late Eocene event is materialised by a widespread unconformity, the age of which is bracketed between 44 and 40 Ma (Noble *et al.*, 1974, 1979, 1990; Mégard *et al.*, 1996). In southern Peru, the middle to late Eocene batholiths of the inner arc are intruded by acid, calc-alkaline subvolcanic stocks of earliest Oligocene age (34 - 32 Ma), thus indicating a strong uplift event during the late Eocene (Cárlier *et al.*, 1996; Carlotto, 1998).

In the arc zone of northern Chile, a significant angular unconformity is dated at 39 - 38 Ma (Hammerschmidt *et al.*, 1992). Late Eocene upright anticlines, which account for 25% shortening in the arc zone, were associated with arc-parallel dextral strike-slip movements and with E-vergent folds and reverse faults in the back-arc zone (Scheuber *et al.*, 1994). In the proximal back-arc area, the middle-late Eocene tectonic event is recorded by the post 42 Ma unconformity which separates the Purilactis and Paciencia groups (Flint *et al.*, 1993; Fig. 30).

In the arc zone of central-southern Ecuador, the late Eocene event is followed by an important pulse of arc volcanism (andesite, dacite and subordinate rhyolite) dated as latest Eocene-Oligocene (39 - 23 Ma, Saraguro Group; Egüez *et al.*, 1992; Dunkley and Gaibor, 1998). Within this pile, Dunkley and Gaibor (1998) identified erosional periods of latest Eocene-earliest Oligocene (36 - 34 Ma) and middle Oligocene age (30 - 27 Ma). In northern Ecuador, volcanic activity seems to have decreased in the Oligocene, but chronological data are scarce (Egüez, 1986; Wallrabe-Adams, 1990).

In the arc zone and the paleo-Andes of central Peru, a plutonic pulse of late middle and late Eocene age (42 - 36 Ma) is followed by a minor pulse of middle Oligocene age (31 - 30 Ma), the latter being restricted to the Paleo-Andes (Soler, 1991). Volcanic activity displays a correlative evolution, since the volcanic Calipuy Formation yielded ages of 41 - 35 Ma, and 31 - 29 Ma (McKee and Noble, 1982; Noble *et al.*, 1979; Soler, 1991). The middle to early late Oligocene magmatic quiescence is correlated with a low convergence period (31 - 26 Ma, Sébrier and Soler, 1991), and is marked by a subtle change in the geochemical composition of the arc magmatism (Soler, 1991). In the Altiplano and Eastern Cordillera of southern Peru, a significant episode of high-K alkaline magmatism occurred between 30 and 27 Ma (Bonhomme *et al.*, 1985; Bonhomme and Carlier, 1990), which express a local extensional regime (Carlier *et al.*, 1996) and is coeval with the emplacement of monzogabbro at the southern edge of the Altiplano (30 Ma, Clark *et al.*, 1990). These late Oligocene-earliest Miocene alkaline, shoshonitic and high-K calc-alkaline effusions and intrusions are interpreted as the result of partial melting of an enriched mantle wedge (Sébrier and Soler, 1991).

In the arc zone of northern Chile, Oligocene times are marked by the deposition of mainly sedimentary, fluvial beds, which indicate a period of magmatic quiescence (40 - 28 Ma, Azapa Formation and Paciencia Group, Coira *et al.*, 1982; Flint *et al.*, 1993; Garcia, 1997; Fig. 30). In the Paleo-Andes, the middle-late Eocene event is well-expressed. In the Western Cordillera of Ecuador, late Eocene times are marked by the deposition of subaerial conglomerate on the Eocene marine sequence, interpreted by some authors as the result of the accretion of the Western Cordillera terrane (Bourgois *et al.*, 1990; Litherland *et al.*, 1994; McCourt *et al.*, 1998).

Deformation is maximum in the Western Cordillera where E-verging fold and thrust belts developed (Mégard, 1978; Ángeles, 1987; Mourier, 1988). In southern Peru, late Eocene times (42 - 38 Ma) are also marked by thrusting to the NE along the southern border of the Altiplano (Laubacher, 1978; Farrar *et al.*, 1988; Carlotto, 1998), and also by SW-verging thrust faults NE of the Altiplano (Huancané Fault Zone, Laubacher, 1978). Farther to the NE, the middle-late Eocene event is responsible for widespread unconformities in the arc zone and the Altiplano, and for disconformities in the eastern areas (Sébrier *et al.*, 1988; Farrar *et al.*, 1988; Noble *et al.*, 1999).

In the present-day Western Cordillera of northern and central Peru, Late Cretaceous strata are folded and faulted, and in the eastern part, compressional deformations result in a 50 km-large, NE-verging fold and thrust belt (Marañon FTB, Mégard 1984, 1987), which occurs on the western border of the Mesozoic positive zone (Marañon Geanticline), and interpreted as the result of the tectonic inversion of normal paleo-faults (Mourier, 1988). This belt expresses a significant shortening of the continental crust and its overlying cover. They are associated with coarse-grained deposits exhibiting internal unconformities (Pacobamba Formation, Ángeles, 1999). In southern Peru, the Incaic Deformation resulted in a comparable NE-verging fold and thrust belt to the S of the Cusco-Puno Swell (Mañazo FTB, Jaillard and Santander, 1992; Carlotto, 1998), and in the SW-verging Huancané Fault Zone (Audebaud *et al.*, 1976; Laubacher, 1978). The subsequent erosion period is concealed by the deposition of widespread, unconformable coarse-grained conglomerate (Chanove *et al.*, 1969), and documented locally by Oligocene terrestrial faunas preserved in karst excavations (Hartenberger *et al.*, 1984). Farther to the NE, the middle-late Eocene event is responsible for widespread unconformities on the Altiplano, and by disconformities in the eastern areas (Laubacher, 1978; Sébrier *et al.*, 1988; Farrar *et al.*, 1988).

In the Cusco area (southern Peru), 5000 to 6000 m of alluvial red beds (San Jerónimo Group), formerly ascribed to the Late Cretaceous (Gregory, 1916; Jaillard *et al.*, 1993; Noblet *et al.*, 1995), are presently dated as late Eocene(?)-middle Oligocene age (Carlotto, 1998; Fig. 34). They comprise two thick coarsening-upward sequences affected by large-scale progressive unconformities showing evidence for syn-sedimentary compressional or transpressional deformation (Córdova, 1986; Noblet *et al.*, 1987; Carlotto, 1998). Farther to the E, as much as 2000 m of very coarse-grained fanglomerate and sandstone of late Eocene to middle Oligocene age (Anta Formation) unconformably overly Cretaceous to middle Eocene rocks, thus providing evidence for strong magmatic and tectonic activity (Carlotto, 1998). Coeval deposits are known farther to the SE from isolated basins exhibiting changing sedimentary and paleogeographic evolutions, and yielding scarce early Oligocene ages (30 - 27 Ma, Carlotto, 1998).

In the Altiplano Basin of Bolivia, the Paleozoic basement is unconformably overlain by a 3000 m thick series of red shale and sandstone beds, with evaporite units in the lower part, dated at 30 - 29 Ma (Tiwanacu Formation, Rochat *et al.*, 1998; Fig. 35). This succession exhibits eastward paleocurrents and is interpreted as the foreland sequence of the Western Cordillera deformed during the late Eocene event (Sempere *et al.*, 1990; Sempere, 1995, Rochat *et al.*, 1998). However, Lamb *et al.* (1997) determined westward paleocurrents and proposed that the Eastern Cordillera was also uplifted along an E-verging thrust fault during the middle-late Eocene deformation, and thus, separated the Altiplano Basin from the incipient eastern foreland basin (Lamb and Hoke, 1997).

In the back-arc basins of Ecuador late Eocene-Oligocene times are represented by disconformable quartz-rich conglomerate (Upper Tiyuyacu Formation), overlain by fine-grained red beds (Orteguaza Formation) exhibiting a conspicuous transgressive layer of partly marine glauconitic sandstone (Benítez *et al.*, 1993; Rivadeneira and Baby, 1999). In eastern Peru, the late Eocene event accounts for a widespread sedimentary hiatus encompassing the late Eocene-middle Oligocene time-span in the western and southern zones (Koch and Blissenbach, 1962; Naeser *et al.*, 1991; Figs. 31 and 36) and for a slight unconformity farther to the E and NE. There, Robertson Research (1990) identified a thin lacustrine unit of probable Oligocene age. Although available stratigraphic data are scarce and sometimes conflicting, they suggest a noticeable decrease of the tectonic subsidence during the late middle to late Eocene interval (40 - 35 Ma, Thomas *et al.*, 1995; Berrones and Cotrina, 1996; Contreras *et al.*, 1996).

The frequent lack of late Eocene-middle Oligocene deposits in the Oriente Basin contrasts with the thick accumulations of coeval deposits in the Paleo-Andes, which seem to have been marked, however, by an E to NE drainage system. This suggests that, either these deposits have been eroded due to a significant late Oligocene uplift of the Eastern Basin, or the entire detrital sediments have been trapped within the Andean basins, which acted therefore as the proximal foreland basins of the Western Cordillera FTB (Sempere, 1995; Carlotto, 1998), the Eastern Basin constituting a by-pass zone for low discharge rivers.

Orogenic Evolution of the North-Central Andes (Late Oligocene - Present)

Late Oligocene - middle Miocene evolution (28 - 10 Ma)

The late Oligocene "Aymara" tectonic event (28-26 Ma)

A major tectonic and geodynamic event occurred in the late Oligocene (28 - 26 Ma). It has been described by Sébrier *et al.* (1988) and Sempere *et al.* (1990; Fig. 32). The late Oligocene event is related to a major plate dynamics reorganization that occurred at 26 Ma. This consisted of the break up of the Farallón Plate into the Cocos and Nazca plates, accompanied by a change in the direction of convergence (Pilger, 1984; Pardo-Casas and Molnar, 1987). Convergence became approximately E-W, which triggered a progressive rotation of the strain, from NNE-SSW during the late Oligocene, to E-W at the end of the Miocene (Noblet *et al.*, 1988).

This tectonic event was marked by regional unconformities in the Andes (Sébrier *et al.*, 1988; Sempere *et al.*, 1990), by the deposition of disconformable coarse-grained conglomerate in the Eastern Basin, by the inception of eastward thrusting in the Sub-Andean Zone (Sempere *et al.*, 1990), and by a sharp increase of the subsidence rates in the Eastern Basin (Thomas *et al.*, 1995; Berrones and Cotrina, 1996). It also triggered or accelerated the subsidence related to subduction-related tectonic erosion in the fore-arc zones, since in most areas, pelagic Miocene deposits disconformably overlie Eocene shelf deposits (Macharé *et al.*, 1986; Suess *et al.*, 1988). This event is also marked by pre-23 Ma disconformities in the Andes of northern Peru (Mourier, 1988), by some resets of K/Ar ages in the Eastern Cordillera of Ecuador (35 - 25 Ma, Litherland *et al.*, 1994), by unconformities at the base of the late Oligocene volcanics of Ecuador (base of the Saraguro Formation, 29 - 26 Ma, Dunkley and Gaibor, 1998; Steinmann *et al.*, 1999), and by unconformities and syn-tectonic sediments in Bolivia (Sempere *et al.*, 1990; Rochat *et al.*, 1998).

Latest Oligocene-early Miocene evolution (26 - 17 Ma)

The major plate dynamics reorganization of late Oligocene age provoked a renewal of tectonic activity, which accelerated the shortening and uplift of the Andes, and induced thick continental sedimentation in the intermontane and retro-arc foreland basins. The late Oligocene tectonic event is post-dated by the creation of a nearly continuous belt of fore-arc (Macharé *et al.*, 1986) and by a sharp increase of the subsidence rates in the Eastern Basin (Thomas *et al.*, 1995). Offshore northern Peru, tectonic subsidence is marked by the unconformable rest of Middle Miocene pelagic deposits upon Eocene shelf deposits (Macharé *et al.*, 1986; Suess *et al.*, 1988; Von Huene *et al.*, 1988; Bourgois *et al.*, 1990; Fig. 37).

In coastal Ecuador, the early Miocene sequence begins locally with transgressive conglomerate overlain by marine shale and siltstone rich in planktic foraminifera and radiolaria (Dos Bocas and Villingota formations, Evans and Whittaker, 1982; Benítez, 1995). The upper part, of early middle Miocene age, locally grades laterally into coarser-grained subaerial deposits (Benítez, 1995). Rapid subsidence of the northern Talara and Tumbes basins is expressed by the deposition of a 250 to 1000 m thick transgressive series of locally conglomeratic sandstone beds, with marly and carbonate-rich intercalations of paralic environment (Máncora Formation), which unconformably rest on Paleozoic rocks (León, 1983). Further subsidence allowed the deposition of as much as 1000 m of shale, marlstone and sporadic turbidite units rich in planktonic foraminifera, which indicate a significant deepening of the basin during the early to middle Miocene (Heath Formation, León, 1983). Offshore northern Peru, early Miocene marine deposits are mentioned only locally. They overlie directly middle Eocene strata (Ballesteros *et al.*, 1988).

In southern central Peru (Pisco Basin), the latest Oligocene-early Miocene deposits consist of a 60 to 300 m thick series of transgressive shale, siltstone and fine-grained sandstone, which unconformably rests on Paleozoic to early

Oligocene rocks (Caballas Formation, Marocco and De Muizon, 1988; Macharé *et al.*, 1988). Farther to the S (outer fore-arc), the late Oligocene series is represented by sandstone, conglomerate and shale of nearshore environment (Camaná Formation, Rüegg, 1956; Macharé *et al.*, 1988). In the inner fore-arc zone of southern Peru (Moquegua Basin; Fig. 32), as much as 700 m of coarse-grained fluviatile deposits dated by volcanic intercalations (25 - 23 Ma, Upper Moquegua Formation, Noble *et al.*, 1985; Marocco *et al.*, 1985) recorded local and short-lived marine transgressions. On its eastern border, superimposed alluvial fans indicate a coeval tectonic uplift of the paleo-Andes. The coarse-grained fore-arc deposits indicate that the paleo-Andes of southern Peru were more actively uplifted than those of Central Peru during late Oligocene-early Miocene times. A similar situation is recorded in the Azapa Basin of northern Chile. The Chilean margin shows in its northern part a well-expressed extensional, asymmetric basin (Muñoz and Fuenzalida, 1997), similar to the Neogene basins occurring farther N (Von Huene and Scholl, 1991), and a horst and graben topography (Fig. 2C).

In the magmatic arc of southern Ecuador, a significant pulse of mainly acid to intermediate arc magmatism of late early Oligocene-early Miocene age has been recognized (33 - 16 Ma, Saraguro Formation, Aspden *et al.*, 1992; Lavenu *et al.*, 1992; Hungerbühler, 1997; Dunkley and Gaibor, 1998). Detailed analysis discloses ignimbritic events at 28 - 26 Ma, 24 - 22 Ma and 20 - 18 Ma, suggesting an extensional tectonic setting in central and southern Ecuador (Steinmann, 1997; Fig. 33).

In northern Peru, only continental greywacke and conglomerate yielded a 23 Ma K/Ar age (Noble *et al.*, 1990). In the magmatic arc zone and Paleo-Andes of Central Peru, after the 30 - 26 Ma plutonic gap, effusion of calc-alkaline basaltic andesite, andesite and dacite resumed near the Oligocene-Miocene boundary (26 - 19 Ma), and occupied a 150 km wide area (Sébrier and Soler, 1991; Soler, 1991; Fig. 20) They originated in the mantle wedge modified by fluids or melts issued from the subducting slab (Sébrier and Soler, 1991).

In the arc zone of southern Peru, after a long volcanic gap (55 - 27 Ma), huge volumes of basaltic to dacitic flows and tuff were outpoured (Tacaza and Sillapaca formations, Barroso Group, 27 - 15 Ma, Lefèvre, 1979; Tosdal *et al.*, 1984; Klinck *et al.*, 1986; Sébrier *et al.*, 1988; Clark *et al.*, 1990; Fig. 32). Their chemical signature indicates a significant crustal contamination, indicative of the beginning of crustal shortening and thickening (Boily *et al.*, 1990). Granodiorite plutons (25 Ma, Bonhomme *et al.*, 1985) and abundant calc-alkaline volcanism occurred along the western Cordillera-Altiplano boundary (Carlier *et al.*, 1996). This period is also marked by a significant peraluminous magmatism in the Eastern Cordillera of southern Peru (28 - 23 Ma, Kontak *et al.*, 1986; Laubacher *et al.*, 1988; Clark *et al.*, 1990; Carlier *et al.*, 1996), which would result from melting of crustal continental material (Sébrier and Soler, 1991). It is associated with and/or followed by emplacement of shoshonitic bodies and K-rich to high-K minettes exhibiting lamproitic affinities (25 - 20 Ma, Bonhomme *et al.*, 1985; Kontak *et al.*, 1986; Carlier *et al.*, 1996). The latter would reflect the thickening of the lithosphere due to incipient

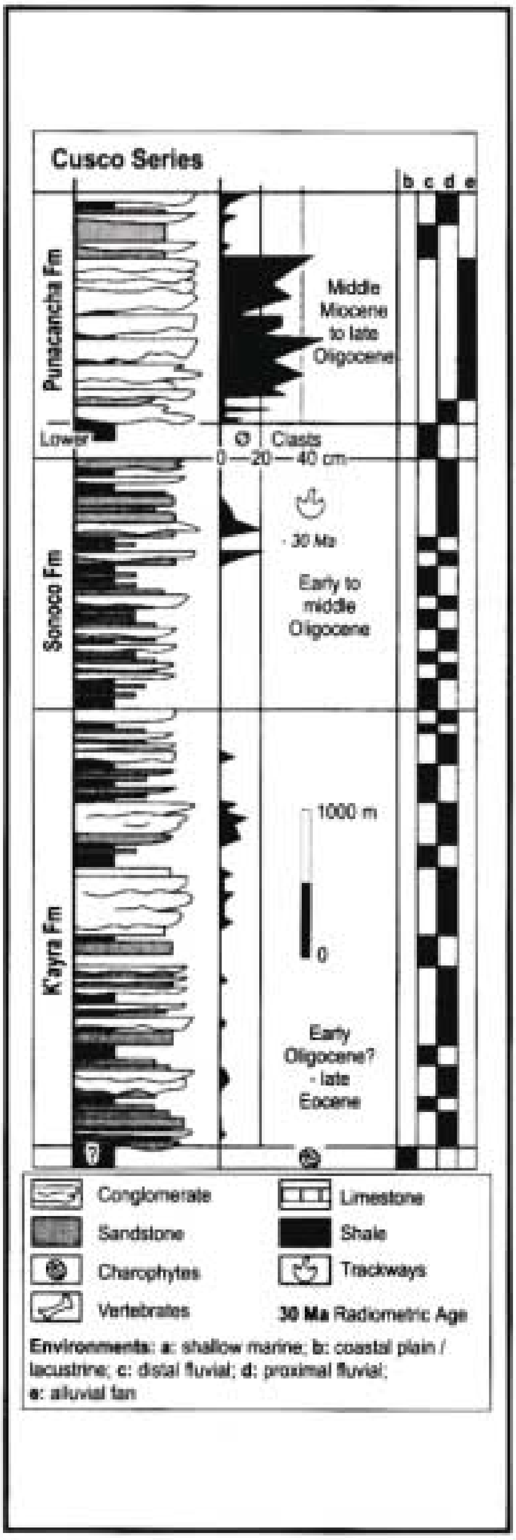


FIGURE 34 - Stratigraphic succession of the Eocene-Miocene "Red Beds" of the Cusco area (Andes of southern Peru) (after Carlotto, 1998).

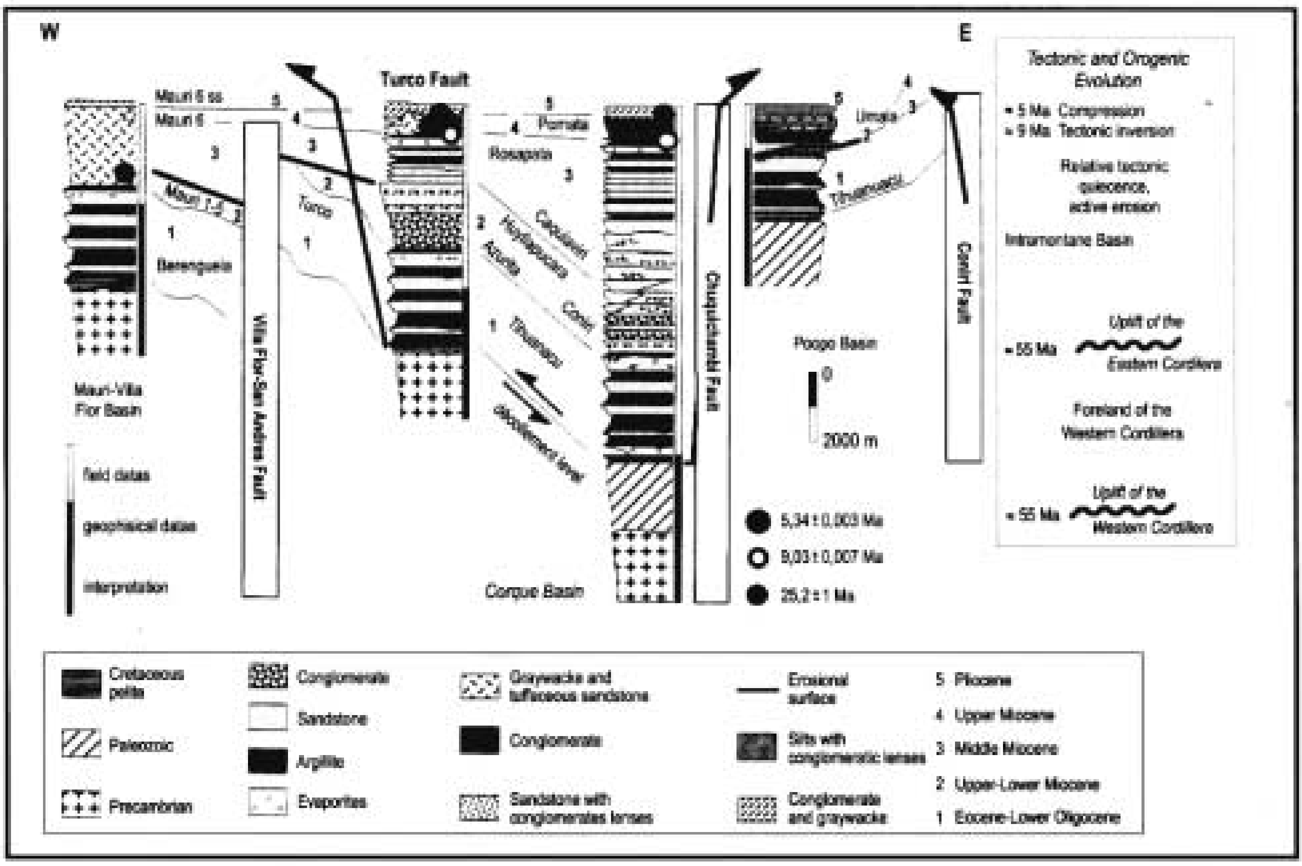


FIGURE 35 - Representative stratigraphic successions accross the Altiplano Basin of Bolivia (after Lamb et al., 1997, Rochat et al., 1998).

thrusting to the NE of the Altiplano onto the Brazilian Shield, triggering also the partial melting of the continental crust (Carlier *et al.*, 1996).

In northernmost Chile, volcanic activity resumed by 25 Ma Oligocene terrestrial sediments (Azapa Formation, Muñoz and Charrier, 1996; Garcia, 1997; Garcia *et al.*, 1999; Fig. 38) are unconformably overlain by late Oligocene-middle Miocene acid tuff associated with fluvial sediments deposited in an extensional environment (Lupica and Oxaya formations, 25 - 18 Ma). Farther to the S (Atacama area), no deposits are known between the playa and fan sediments of the upper Paciencia Group (28 Ma) and the unconformable volcanic rocks of the San Bartolo Group, the base of which is dated at 17 Ma (Flint *et al.*, 1993).

In the present-day Andes of Peru and Ecuador, the late Oligocene-early Miocene was thought to be marked by the creation of a remarkable belt of intermontane basins (Marocco *et al.*, 1995; Noble *et al.*, 1999). However, recent F-T dates led to the consideration that most intermontane basins of Ecuador are late early to early middle Miocene in age (18 - 9 Ma, Hungerbühler, 1997; Steinmann, 1997; Steinmann *et al.*, 1999; Fig. 40).

In the Cuzco area, the Oligocene red beds are disconformably overlain by a fine-grained unit, in turn unconformably overlain by a 4000 m thick series of sandstone and conglomerate beds of late Oligocene-early Miocene age (Punacancha Formation, Carlotto, 1998; Fig. 34). These mainly represent reworked volcanic rocks and exhibit compressional syn-sedimentary deformation, thus evidencing noticeable coeval volcanic and tectonic activity (Marocco and Noblet, 1990; Carlotto, 1998).

On the Altiplano and in the Eastern Cordillera, the late Oligocene-early Miocene sequence is marked by the arrival of coarse-grained conglomerate units (Azurita and Coniri formations, Kennan *et al.*, 1995; Rochat *et al.*, 1998; Fig. 35).

The back-arc area is marked by a strong increase in the

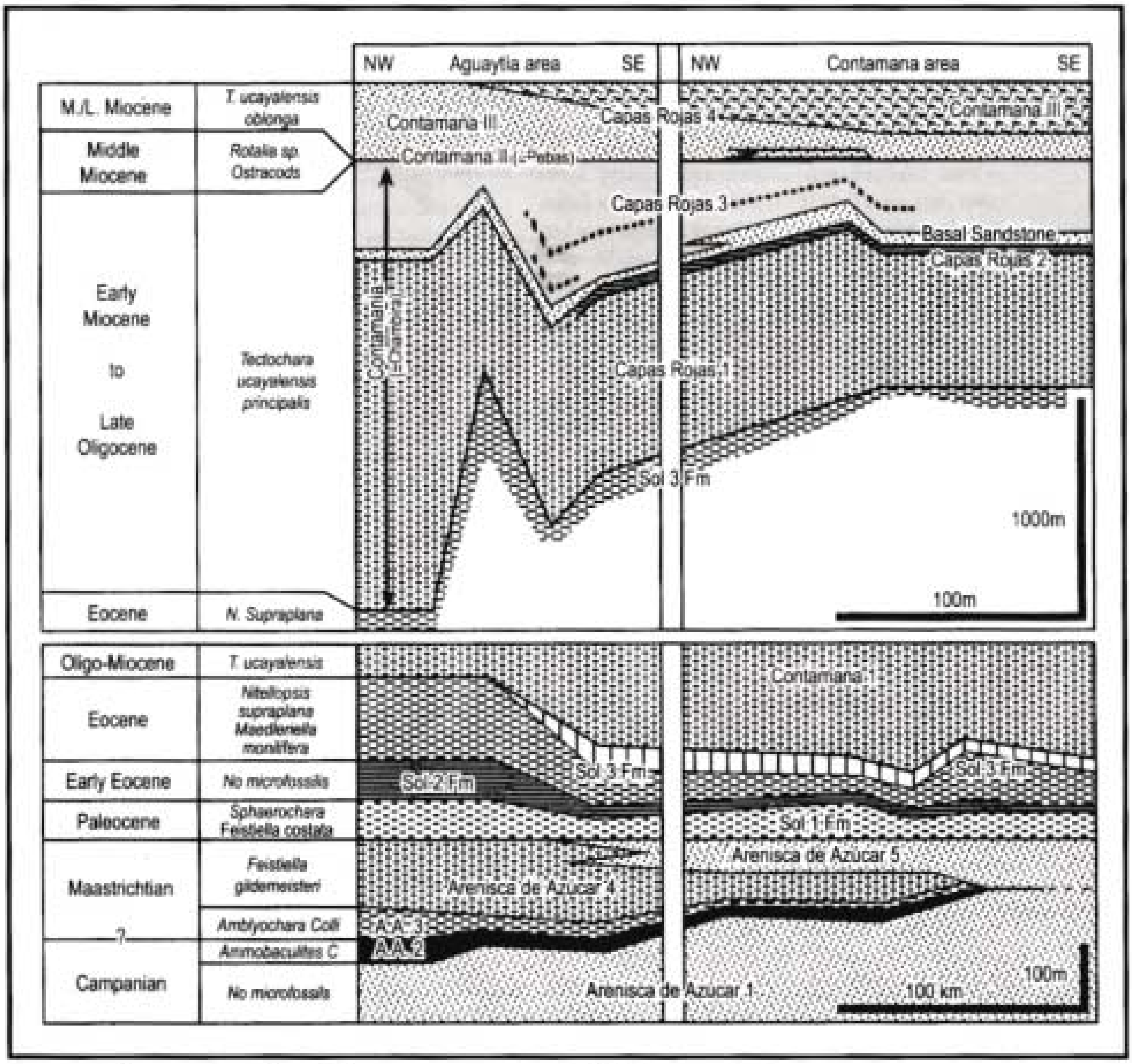


FIGURE 36 - Paleogeographic profiles of the latest Cretaceous-Miocene "Red Beds" of the Ucayali Basin (eastern Central Peru) (after Koch and Blissenbach, 1962; after Müller and Aliaga, 1981).

tectonic subsidence, which marks the start of its evolution as a true retro-arc foreland basin (Sempere *et al.*, 1990; Dashwood and Abbots, 1990; Thomas *et al.*, 1995; Baby *et al.*, 1995; Berrones and Cotrina, 1996; Contreras *et al.*, 1996; Fig. 39). The latest Oligocene-middle Miocene sequences overlie a marked unconformity representing a long period of non-deposition. In Ecuador, Eocene red beds are unconformably overlain by 500-1000 m of siltstone with thin interbeds of sandstone or evaporite, of lacustrine to continental environment, ascribed to the early (to middle?) Miocene (Chalcana Formation). Near Rentema (northern Peru), a 1000 m thick series of alluvial siltstone, sandstone and conglomerate beds (Upper Sambimera Formation) is probably of late Oligocene to middle Miocene age (Naeser *et al.*, 1991; Fig. 31). Farther to the E and SE (Marañon, Ucayali basins), the distal equivalent of this series (Chambira Formation, Contamana I) consists of red siltstone, shale and thin-bedded sandstone, with local evaporite and coal beds of brackish to lacustrine environment. They overlie directly the Eocene beds (Koch and Blissenbach, 1962; Marocco *et al.*, 1993; Fig. 36). These deposits express a resumed erosion of the paleo-Andes subsequent to the late Oligocene phase (Marocco *et al.*, 1993). Farther to the SE (Madre de Dios Basin), the Tertiary succession has never been studied.

In the Cordillera Real of Bolivia, an important magmatic pulse is responsible for the intrusion of numerous S-type granitoid plutons dated between 28 and 23 Ma (Ávila-Salinas, 1990). Farther to the E, the Eastern Basin of Bolivia is marked by a late Oligocene-early Miocene succession of continental sandstone associated with subordinate conglomerate and shale, interpreted as early foreland deposits related to the uplift of the Eastern Cordillera (Petaca Formation, Marshall *et al.*, 1993; Sempere, 1995).

FIGURE 37 - Interpretative structural section and well successions across the fore-arc zone of northern Peru (Trujillo and Naquina basins) (after Suess et al., 1988).

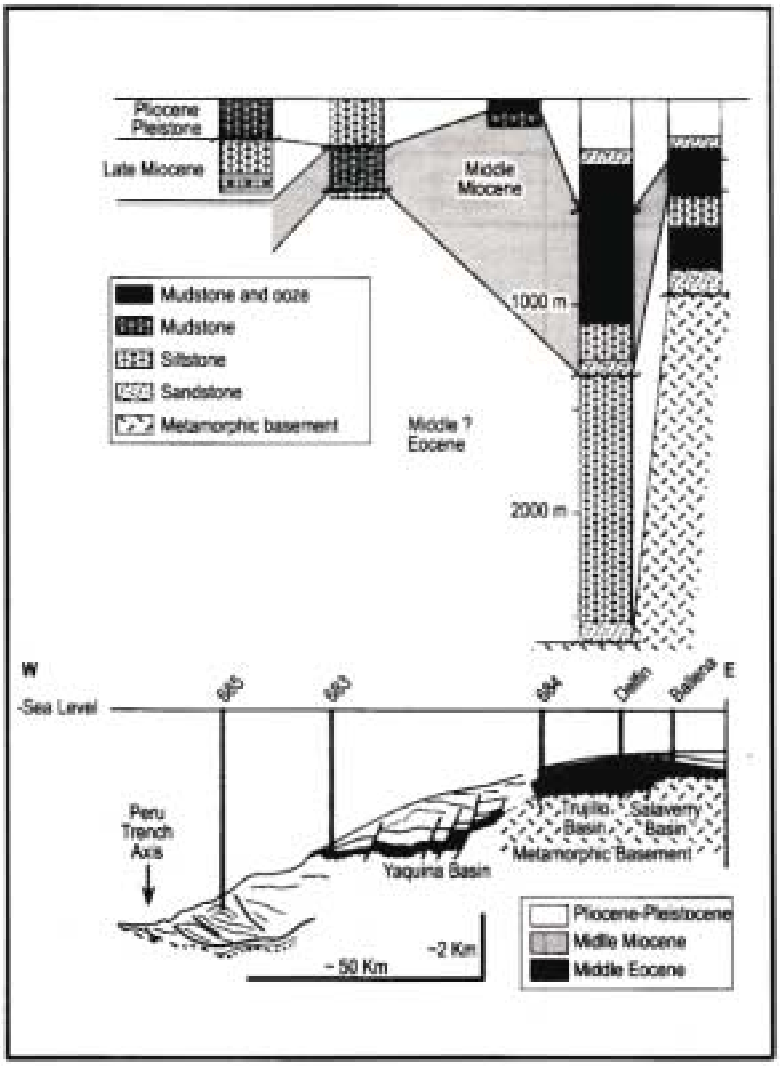
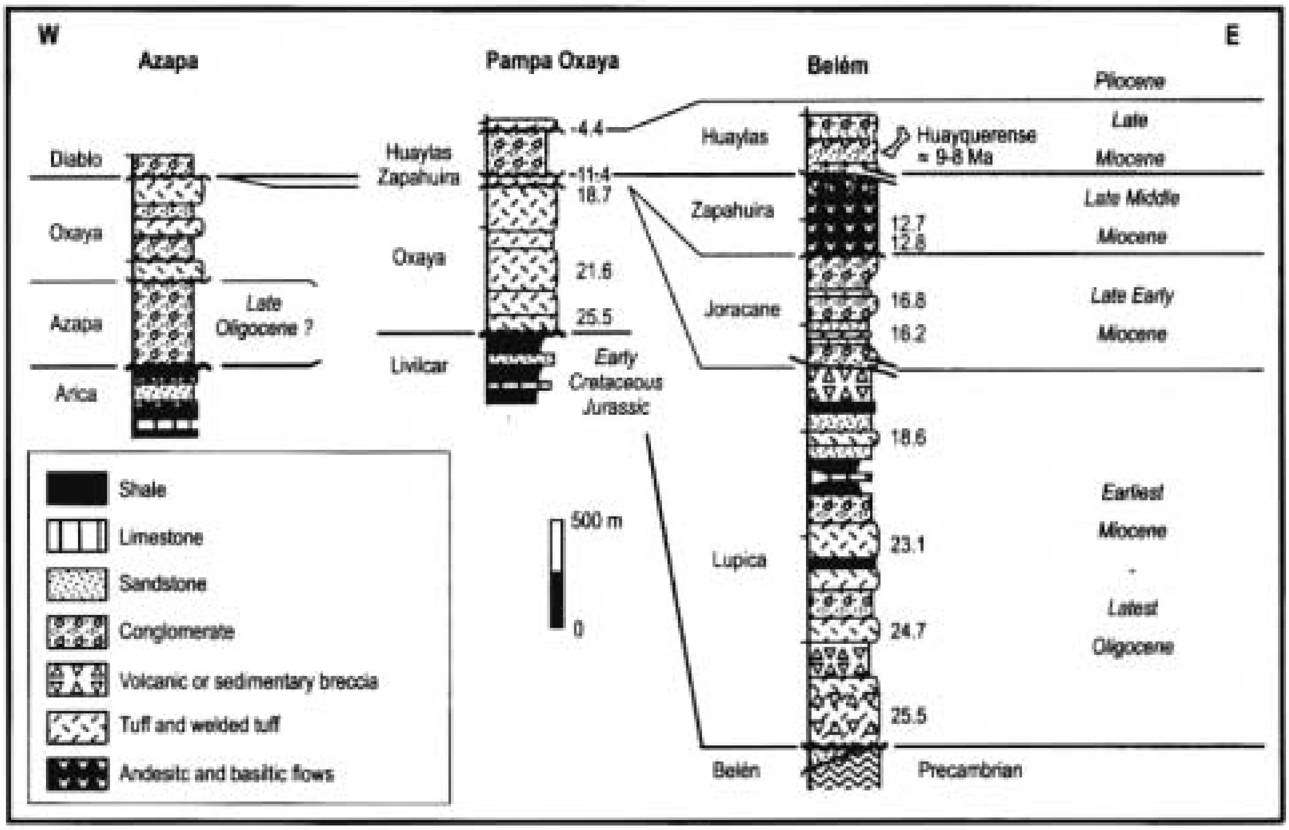


FIGURE 38 - Representative sections in the eastern fore-arc zone of northernmost Chile (after Garcia, 1997; Garcia et al., 1999).



"Quechua 1" event (17 - 15 Ma) and middle to late Miocene evolution (15 - 9 Ma)

In the fore-arc basins of Ecuador, no good record of the event, which occurred near the early-middle Miocene boundary (17 - 15 Ma), is known. This event may be recorded by the arrival of moderate amounts of detrital sediments and by local transgression (Progreso Basin). In the Andes of Ecuador, F-T dates on sediments indicate an uplift stage around 18 Ma (Steimann *et al.*, 1999). In the Eastern Cordillera of Ecuador, resets of K/Ar ages at 20 - 15 Ma may be due to compressional deformations (Litherland *et al.*, 1994). This interval is followed by the opening of the Guayaquil Gulf (Deniaud *et al.*, 1999).

In central Peru, the middle-late Eocene structures were re-activated in the Western Cordillera and in the Marañon Fold and Thrust Belt (Mégard, 1984; Mégard *et al.*, 1984), which fold late Cretaceous red beds. A possible magmatic gap occurred around 19 - 18 Ma in the arc of central Peru (Soler, 1991). In southern Peru, it is responsible for monoclinal folds and reverse faults, large scale incisions due to the resumption of erosions triggered by a pulse of uplift (Sébrier *et al.*, 1988). The latter is thought to be responsible for a 400 m uplift (Sébrier *et al.*, 1988).

In the arc and fore-arc zones of northern Chile (Fig. 42), W-verging fault thrusts are assumed to have began around 18 - 15 Ma (Garcia *et al.*, 1996; Muñoz and Charrier, 1996; Charrier *et al.*, 1999). This event is associated in northernmost Chile, with unconformable late middle Miocene lava flows (Zapahuira Formation, 13 - 11 Ma) on *the early Miocene tuff and conglomerate* (Garcia, 1997; Fig. 38). Farther to the S (Antofagasta), the volcanic rocks of the San Bartolo Group (17 Ma) overlie Oligocene fluvial sediments with an angular unconformity (Flint *et al.*, 1993).

In the fore-arc basins of Ecuador, early Miocene marlstone beds are disconformably overlain by middle Miocene sandstone and marlstone (Subibaja-Angostura Formation). They are in turn overlain by early late Miocene conformable mudstone (Onzole Formation, Benítez, 1995). In the Progreso Basin, the middle Miocene disconformity (Subibaja and San Antonio transgressive limestone members) is followed by deposition of nearshore sandstone of late Miocene age (Progreso Formation), related to the beginning of the first opening stages of the Guayaquil Gulf. After an opening stage (late Oligocene-early Miocene) the Tumbes Basin of northernmost Peru received deep-marine turbidite beds (León, 1983). In the fore arc of central and northern Peru, the early Miocene deposits are conformably overlain by thick middle Miocene marine mudstone (Macharé *et al.*, 1986; Ballesteros *et al.*, 1988; Fig. 37).

Except in Ecuador, the arc zones are marked by the resumption of significant amounts of volcanic products. In central Peru, a magmatic pulse occurred in the Eastern Western Cordillera and the Altiplano between 18 and 13 Ma, which comprises abundant volcanism (Soler, 1991; Fig. 20). In southern Peru, the shoshonitic and High-K magmatism came to an end, while calc-alkaline magmatism went on (Carlier *et al.*, 1996). In the arc and fore-arc zone of northernmost Chile, late middle Miocene lava flows (Zapahuira Formation, 13 - 11 Ma) unconformably overlie early Miocene tuff and conglomerate (Garcia, 1997; Fig. 38). Farther to the S, a volcanic unit (San Bartolo Group, 17 Ma) overlies Oligocene fluvial sediments with an angular unconformity (Flint *et al.*, 1993; Fig. 30).

In the paleo-Andes of Ecuador and Peru, the opening of the "Miocene" intermontane basins had been dated as mainly early Miocene (K/Ar, 26 - 22 Ma, Lavenu *et al.*, 1992) and were regarded as pull-apart basins, opened by the play of an oblique NE trending strain exerted on pre-existing NNE and ENE trending faults (Noblet *et al.*, 1988; Baudino *et al.*, 1994; Marocco *et al.*, 1995; Barragán *et al.*, 1996). However, recent F-T concordant dates support a late early to early middle Miocene age (16 - 14 Ma) for the creation of most intermontane basins of Ecuador (Figs. 33 and 40), which have been alternatively regarded as the result of an E-W extensional regime (Steinmann, 1997; Hungerbühler, 1997; Steinmann *et al.*, 1999). An extensional regime is also thought to have governed the creation of many Miocene intermontane basins of Peru (Noble *et al.*, 1999).

In Ecuador, the rapidly opened intermontane basins (16 - 15 Ma) were filled by middle Miocene fine-grained lacustrine deposits (14 - 10 Ma) representing a period of relative tectonic quiescence (Noblet *et al.*, 1988; Marocco *et al.*, 1995). The fine-grained deposits are overlain by a coarsening-upward sequence of sandstone and conglomerate, coeval with the compressional *closure of these basins (Mégard et al., 1984; Noblet et al., 1988;* Baudino *et al.*, 1994; Marocco *et al.*, 1995; Hungerbühler *et al.*, 1995; Hungerbühler, 1997; Fig. 40). The Cuenca and Loja basins of southern Ecuador contain middle Miocene marine intercalations, indicating that they were located at or very close to sea level and are regarded as embayments of fore-arc basins (15 - 11 Ma, Hungerbühler 1997; Steinmann *et al.*, 1999; Fig. 40). In Peru, the non marine volcaniclastic and sedimentary in-fill of the middle Miocene basins is usually unconformably overlain by mainly volcanic deposits dated at 10 to 7 Ma (Mégard *et al.*, 1984; Marocco *et al.*, 1995; Noble *et al.*, 1999).

The Altiplano Basin of Bolivia is a peculiar case of intermontane basin consisting of N-S-trending half-grabens (Fig. 35). In this basin, shale, sandstone and subordinate conglomerate of middle Miocene age rest conformably on the early Miocene deposits (Lamb *et al.*, 1997; Rochat *et al.*, 1998). The middle Miocene sequence was deposited with very high sedimentation rates, especially in the Corque Basin of southern Bolivia (Roperch *et al.*, 1999a). Clastic sediments mainly derived from the Eastern Cordillera, the erosion of which allowed the development of regional-scale flat morphological surfaces. These are the Chayanta (13 - 14 Ma) and San Juan de Oro (10 Ma) surfaces, which can be observed and traced from northern Argentina up to the La Paz region (Servant *et al.*, 1989; Hérail *et al.*, 1993; Gubbels *et al.*, 1993). This period coincides with a low uplift rate of the Eastern Cordillera.

In the back-arc basins, a conspicuous shallow marine transgression is recorded during the late middle Miocene (15 Ma, Pebas Formation of Peru, Hoorn, 1993), which was connected to the open marine realm through the Maracaibo area (Hoorn *et al.*, 1995) and possibly the Guayaquil seaway in southern Ecuador (Steinmann *et al.*, 1999). A similar shallow marine invasion of early late Miocene age is known in the Eastern Basin of Bolivia (Yecua Formation, Marshall *et al.*, 1993), which was connected southward to the Atlantic Ocean. In Ecuador and northern Peru, this period is marked by slight decrease of the tectonic subsidence (Thomas *et al.*, 1995; Contreras *et al.*, 1996; Fig.39). However, in eastern Bolivia, a strong increase of tectonic subsidence has been related to the deformation of the eastern Cordillera (Marshall *et al.*, 1993; Sempere, 1995).

"Quechua 2" event (9 - 8 Ma) and late Miocene evolution (9 - 6 Ma)

This period begins with the Quechua 2 tectonic phase (9 - 8 Ma, Mégard, 1984; Sébrier *et al.*, 1988). Rather than a major deformational event, the Quechua 2 event is a turning point in the evolution of the northern Central Andes, which corresponds to the change from a depositional period characterized by thick and relatively widespread fining-upward sequences, to a compressional and uplift period marked by erosions and depositional areas restricted to the fore arc and retro arc domains. This is interpreted as the result of the beginning of the nearly *en bloc* eastward thrusting of the Paleo-Andes onto the Brazilian and Guiana shields, which resulted in crustal thickening and rapid uplift of the arc zones and paleo-Andes, and the transfer of active deformation into the Subandean thrust and fold belts.

The fore-arc zone are mainly marked by uplift (Sébrier *et al.*, 1988), unconformities and reverse faulting. In coastal Ecuador, the middle Miocene disconformity is overlain by transgressive marine sandstone (Angostura Formation) and then by a thinning-upwards succession of marine nearshore sandstone of late Miocene age (Progreso and Lower Onzole formations), related to the beginning of the opening of the Guayaquil Gulf (Deniaud *et al.*, 1999). In the fore-arc zone of Peru, the unconformity between middle and late Miocene marine deposits in the fore-arc basins of central Peru (Lima Basin, 11°S) has not been recognized farther N (Yaquina Basin, 9°S, Ballesteros *et al.*, 1988). In the Lima Basin, low energy shelf deposits of late Miocene age evolved toward shelf to slope deposits (Ballesteros *et al.*, 1988), indicating a noticeable deepening of the depositional environment, due to the subsidence related to tectonic erosion of the fore-arc zones (Von Huene *et al.*, 1988). In the Yaquina Basin, middle Miocene low energy turbidite beds grade upwards into higher energy turbidites of late Miocene-Pleistocene age (Ballesteros *et al.*, 1988).

In the fore-arc zone of northern Chile, where W-verging thrusting went on (Fig. 42), a same kind of unconformity is recognized between late middle Miocene lava and late Miocene alluvial deposits (Huaylas Formation, 9 - 8 Ma, Garcia, 1997; Fig. 38). In the magmatic arc of Peru, no location change is observed in the magmatic activity, with respect to the former periods. A major magmatic pulse occurred between 12 and 7 Ma (peak activity around 10 Ma). It corresponds to numerous intrusions, and very abundant effusive products (Soler, 1991). A possible gap, or at least magmatic quiescence, is mentioned at 9 - 8 Ma, which may coincide with a compressional event (Soler, 1991; Fig. 20).

The Paleo-Andes and surrounding areas are marked by a general and rapid uplift (Sébrier *et al.*, 1988; Steinmann *et al.*, 1999), the local rates of which remain to be specified. In Ecuador, the estimates of mean rock uplift rate since the late Miocene (9 Ma) is of 0.7 mm/y and the mean surface uplift is of 0.3 mm/y (Hungerbühler, 1997; Steinmann *et al.*, 1999), which is consistent with estimates by Delfaud *et al.* (1999) in the same area, by Sébrier *et al.* (1988) in southern Peru, and by Parraguez *et al.* (1997) in northern Chile. In the latter area and in Bolivia, uplift involved both the Western (Sébrier *et al.*, 1988) and Eastern (Benjamin *et al.*, 1987) Cordilleras.

In Ecuador, this period is marked by the compressional closure of the Miocene intramontane basins (Noblet *et al.*, 1988; Marocco *et al.*, 1995; Hungerbühler, 1997; Steinmann *et al.*, 1999), which are filled by coarsening and thickening-upward clastic deposits. This is interpreted as the result of a change from an extensional (15 - 10 Ma) to a compressional regime (9 - 8 Ma), which resulted in the uplift of southern Ecuador, the establishment of terrestrial conditions in the intermontane basins of southern Ecuador and rising relief in the Eastern Cordillera (Hungerbühler, 1997). Latest Miocene times are then marked by the development of smaller-scale intermontane basins filled with mainly volcanic and volcanogenic rocks (Lavenu *et al.*, 1996; Hungerbühler, 1997; Fig. 41).

In northern Peru, although timing constraints are poorer, and the change from extensional to compressional regime is assumed to be of late Miocene age. In central Peru, N-S shortening induced mainly dextral movements along NW trending faults (Mégard, 1984). In the Ayacucho Basin, compressional deformation occurred between 9.5 and 8.5 Ma (Mégard *et al.*, 1984). In southern Peru, latest Miocene transpressional stress is thought to be responsible for the closure of intermontane basins (Paruro Basin) opened about 12 Ma ago (Carlotto, 1998). Farther to the S, late Miocene times are marked by the contraction of the Altiplano, related to the tectonic inversion of the pre-existing normal faults defining the hemi-grabens located W of the Altiplano Basin (Kennan *et al.*, 1995; Lamb *et al.*, 1997; Rochat *et al.*, 1998, 1999; Fig. 35). In the Altiplano Basin, tuff beds dated at 9 Ma are disconformably overlain by a thin sequence of conglomerate reworked Paleozoic basement rocks (Lamb *et al.*, 1997; Rochat *et al.*, 1998).

In northern Chile, compressional tectonic activity went on in the Pre-Cordillera, with the development of the western thrusts of the W-vergent thrust system (Fig. 42), whereas an extensional regime prevailed farther to the W (Longitudinal Valley and Coastal Cordillera; Muñoz and Charrier, 1996; Garcia *et al.*, 1999).

Regarding the Eastern Basin, the eastward migration of the deformation front during the late Miocene is the prevailing feature (Sheffels, 1990; Baby *et al.*, 1997). From this time onwards, most of the deformation and shortening of the Andean margin is accommodated by the eastern areas, which received W-proceeding coarse-grained clastic, terrestrial sediments. In the Oriente Basin of Ecuador, late Miocene deposits consist of thick sequences of poorly dated coarse-grained conglomeratic sequences separated from each other by disconformities (Christophoul, 1999). In the western part of the Eastern Basin of northern Peru (Bagua

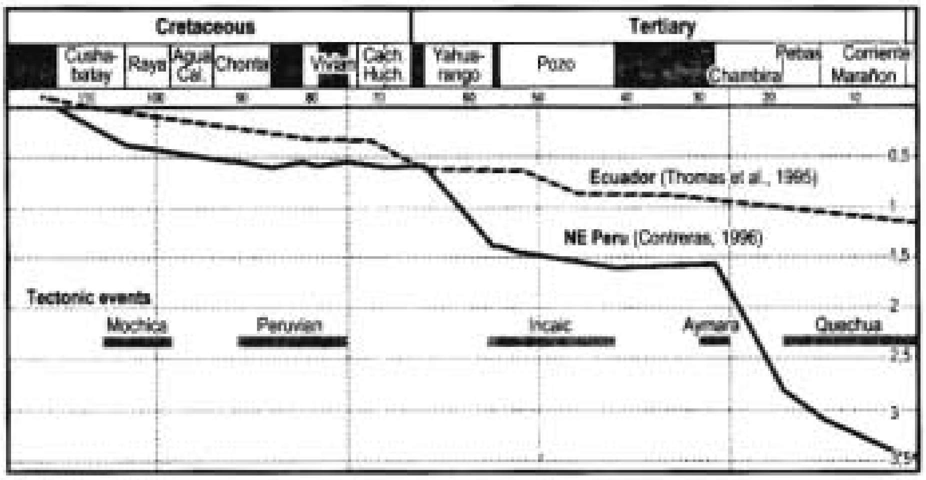
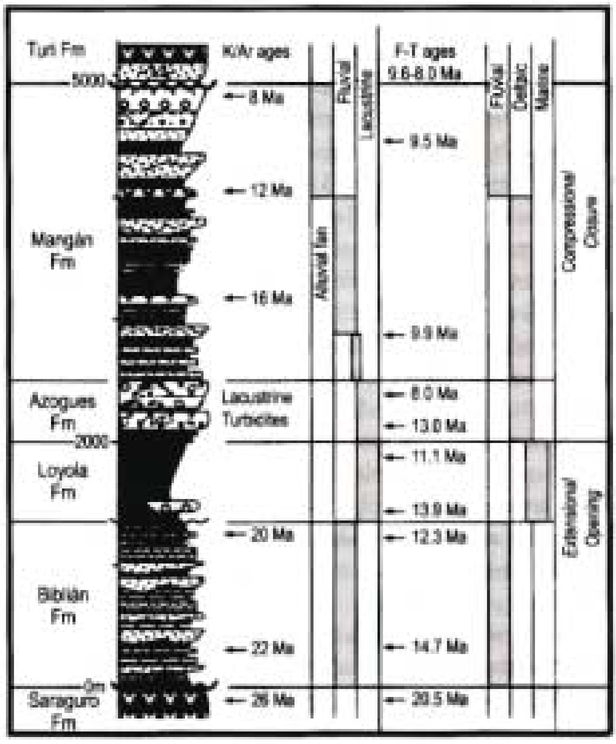
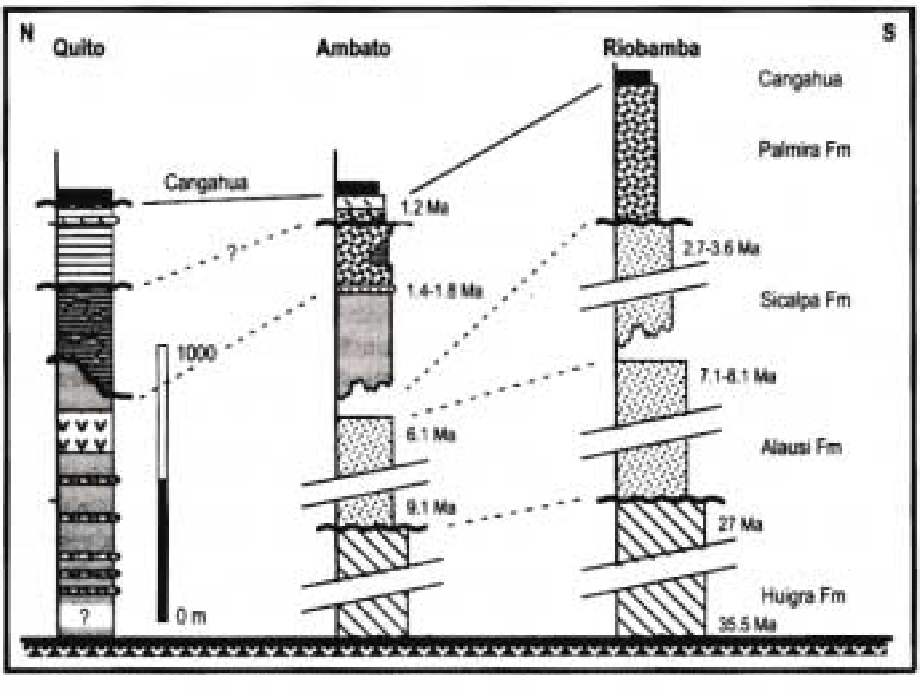


FIGURE 39 - Average tectonic subsidence curves for the Oriente Basin of Ecuador (Thomas et al., 1995) and the Marañon Basin of northern Peru (Contreras et al., 1996).

FIGURE 40 - Lithology, age, environments and evolution of the sedimentary infill of the Cuenca Basin (Ecuador); after Marocco et al., 1995; Steinmann et al., (1999).

FIGURE 41 - Representative successions of the volcanogenic infill of the inter-Andean basins of central Ecuador (after Lavenu et al., 1996).





area), a marked unconformity (10 Ma) separates fluvial sandstone and conglomerate of middle Miocene age, from late Miocene coarse-grained fanglomerate (San Antonio Formation, Mourier *et al.*, 1988; Naeser *et al.*, 1991; Fig. 31). In the Ucayali Basin, fluvial siltstone and sandstone of late Miocene age abruptly overlie marine to brackish beds (Koch and Blissenbach, 1962; Fig. 36).

Quechua 3 event (7 - 5 Ma) and latest Miocene-Present evolution (6 - 0 Ma)

At the beginning of this period, the significant 7 - 5 Ma contractional event is marked by a chiefly E-W shortening (Mégard, 1984; Sébrier *et al.*, 1988). It is marked by folding and reverse and strike-slip faulting in southwestern Peru (Sébrier *et al.*, 1988), and by the onset of the sub-Andean thrust and fold belts, which accommodate most of the shortening in the Andean Chain during the Pliocene (Roeder, 1988; Sheffels, 1990; Baby *et al.*, 1992).

In the fore-arc zone the deformation and uplift were influenced by the incipient subduction of the Nazca and Carnegie aseismic ridges (7 to 3 Ma ago, Suess *et al.*, 1988; Von Huene and Scholl, 1991, Benítez, 1995), and the ongoing tectonic erosion of the fore-arc zones. In the fore-arc zones of Ecuador, a regional disconformity dated at the Miocene-Pliocene boundary (5.5 Ma) precedes the deposition of a coarsening and shallowing-upward sequence, related to the increased uplift and erosion of the Andean Chain from 9 Ma (Benítez, 1995; Deniaud *et al.*, 1999). In the Guayaquil Gulf, however, a strong subsidence due to transtensional movements allowed the deposition of huge volumes of clastic sediments, especially during the early Pleistocene (Deniaud *et al.*, 1999). A similar disconformity and hiatus seem to be recorded at 5 Ma in the offshore basins of Peru (Von Huene *et al.*, 1988).

In central Peru (6° - 14°S), the subsidence of the fore-arc zones related to the tectonic erosion is recorded in the Lima Basin by Pliocene low energy turbidite beds, which indicate deepening of the environment (Ballesteros *et al.*, 1988), by the local transition from uplift to subsidence at 6 Ma (Von Huene *et al.*, 1988), and by the lack of uplifted terraces in the coastal zone (Macharé and Ortlieb, 1993). On the contrary, significant uplift movements affected the coast of northern (4° - 6°S, < 0.2 mm/y) and southern Peru (14° - 18°S, < 0.7 mm/y) since the late Pliocene (Macharé and Ortlieb, 1992; Von Huene *et al.*, 1988). In this latter case, uplift is due to the subduction of the Nazca aseismic ridge, which induced a regional extensional stress regime (Macharé and Ortlieb, 1992)

All segments of the Andes are marked by the continuation of the major and rapid uplift. In Ecuador, uplift rate estimated by F-T evidence a slow down between 6 and 4 Ma, and an increase from 3 Ma (Steinmann *et al.*, 1999). Other methods estimate 1000 - 1200 m of net uplift since 5 Ma (0.2 mm/y, Delfaud *et al.*, 1999). Pliocene-Quaternary times are also marked by the uplift of the sub-Andean Zone of Ecuador (Baby *et al.*, 1999). In the Andes of southern Peru, uplift is estimated at 1300 m since the late Miocene, of which 200 - 300 m would be of Quaternary age.

The arc zones of Ecuador (Steinmann *et al.*, 1999), and Northern and Central Peru are marked by an effusive pulse centred around 5 - 4 Ma. In Peru, it corresponds mainly to ignimbritic tuff associated with rhyolitic dykes in the Western Cordillera (Soler, 1991). In southern Peru, the emplacement of alkaline, peraluminous and shoshonitic suites along major fault systems suggests that an extensional regime prevailed around 6 - 5 Ma (Carlier *et al.*, 1996). Effusions of shoshonite, minette, lamproite and peraluminous rhyolite and dacite went on during the past 3 Ma, and took place along the fault systems limiting the Altiplano, and interpreted as sinistral wrench-faults (Carlier *et al.*, 1996; Carlotto, 1998).

In the eastern basins, latest Miocene-Pliocene times are marked by a strong flexural subsidence allowing thick accumulations of foreland clastic deposits (Thomas *et al.*, 1995; Contreras *et al.*, 1996; Baby *et al.*, 1995; Fig. 39), whereas Recent times are marked by a strong decrease of the sedimentation rate and local uplifts. However, Recent sedimentation continues in restricted and/or more easterly areas (Ucamara depression of easternmost Peru; Pastaza-Marañon alluvial fan).

In the Oriente Basin of Ecuador, localized coarse-grained fanglomerate are incised by present-day rivers (Christophoul, 1999). In northeastern Peru, apatite fission tracks data indicate that the Santiago Basin underwent a rapid uplift (0.4 mm/y) during the last 10 Ma, probably related to the onset of the Santiago Fold and Thrust Belt during the latest Miocene (Pardo, 1982; Mégard, 1984). In the Marañon Basin, a sedimentary hiatus separates middle-late Miocene fine-grained deposits from disconformable coarse-grained fanglomerate of latest Miocene-Recent age (Mathalone and Montoya, 1995) and Pliocene times are marked by the uplift of the area (Contreras *et al.*, 1996). In the Ucayali Basin, no post-Miocene deposits are known (Koch and Blissenbach, 1962). In the Madre de Dios Basin, late Miocene deposits fill incised valleys, and the recent alluvial terrace morphology shows a Pleistocene uplift of the area. In the sub-Andean foreland basin of northern Bolivia, a significant increase in the subsidence allowed the accumulation of about 5000 m of late Miocene-Pliocene clastic sediments. This is interpreted as the result of the rapid eastward migration of the Andean deformation 10 to 6 Ma ago (Gubbels *et al.*, 1993; Baby *et al.*, 1995). However, no foredeep sedimentation occurs at present (Roeder, 1988).

Tectonic and kinematic evolution of the North-Central Andes

From Bolivia to Ecuador, structural style of the Andes changes dramatically. Geometry of the present-day deformation of the Bolivian and North-Chilean Andes results from Neogene thin-skinned tectonics, whereas the Ecuadorian Andes have been structured by thick-skinned and wrench tectonics since Cretaceous times. Figures 2A, 2B and 2C illustrate global changes in structural geometry and chain width. These two parts of the Andes form two extremes. Neogene tectonic events seem to occur contemporaneously, but express two shapes of orogenic belt. Three Neogene orogenic stages, late Oligocene-early

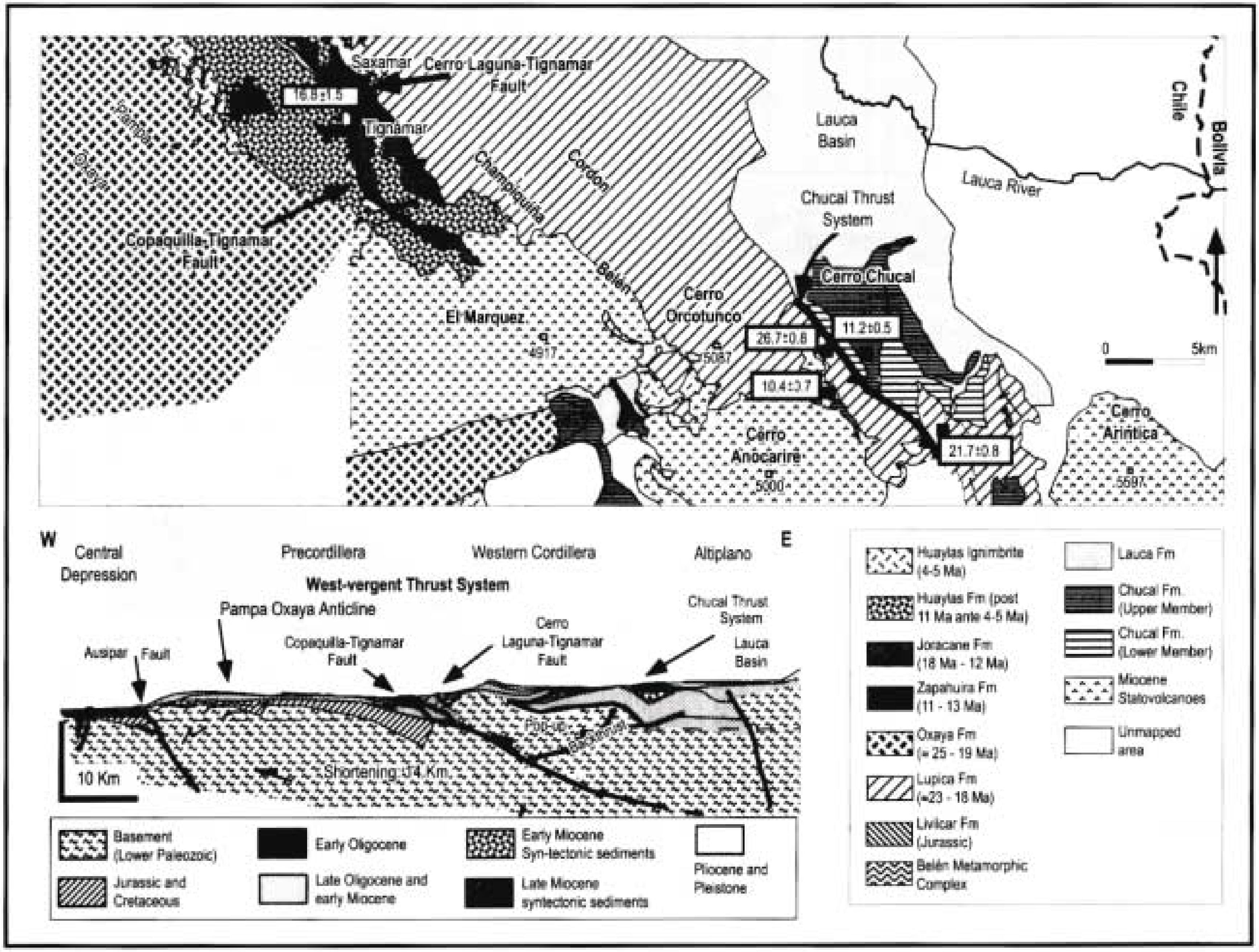


FIGURE 42 - Geology and structural sketch of the Altiplano and Precordillera of northern Chile.

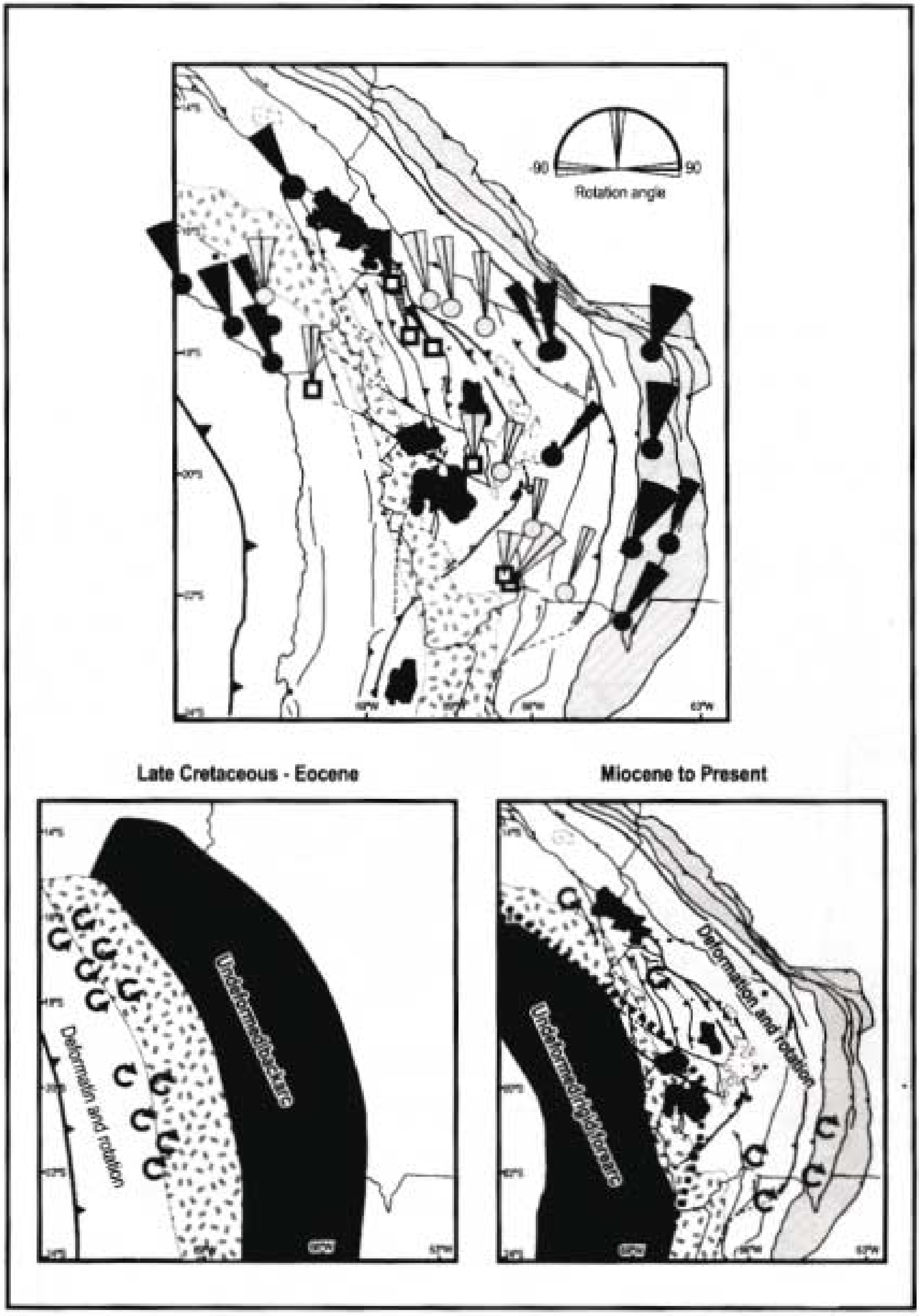


FIGURE 43 - Tectonic rotations in the Central Andes. Important rotations characterize the fore-arc's evolution before the Miocene. The Arica elbow probably developed during the Upper Cretaceous and Eocene. The rotations found in the Altiplano-Puna are linked to the pushing of the rigid fore-arc and to the paleogeographic control of the propagation of the deformation in the subandean belt (after Roperch et al., 1999b).

Miocene, late Miocene, and Pliocene-Quaternary, have been distinguished according to the deformation style and rate. They have been recorded in fore-arc, intermontane and back-arc basins. Sedimentary successions involved in the deformations consist of Cambrian to Oligocene pre-orogenic strata, and Oligocene-Miocene to Recent continental syn-orogenic in-fill.

Central Andes of Bolivia

Recent work has shown the importance of crustal shortening for the development of the structural pattern of this part of the Andes (Allmendinger *et al.*, 1983; Isacks, 1988; Roeder, 1988; Sheffels, 1990; Sempere *et al.*, 1990; Baby *et al.*, 1992, 1997; Gubbels *et al.*, 1993; Schmitz, 1994; Kley and Reinhardt, 1994; Dunn *et al.*, 1995; Rochat *et al.*, 1999; Giese *et al.*, 1999).

Crustal structures and Neogene shortening

The Central Andes are divided from E to W into several morpho-tectonic units (Fig. 43). The Chaco and Beni plains correspond to a slightly deformed Neogene foreland basin underlain by the Brazilian Shield. It is overthrust by the Subandean Zone, a complex thin-skinned fold and thrust belt characterized in its central part (Santa Cruz elbow) by large-scale transfer zones (Baby *et al.*, 1996). The northern branch of the Subandean Zone is characterized by large scale thrust sheets (10 - 20 km of offset) and broad synclines (Roeder, 1988) filled by up to 6000 m of syn-tectonic Neogene sedimentary rocks (Baby *et al.*, 1995a). Surface mapping, seismic reflection data, and drilling information show that the main detachment planes are located in the Ordovician, Silurian, Devonian and Permian shaly levels (Baby *et al.*, 1995b). The slope of the base of the foredeep is 4° toward the SW. The amount of shortening is 74 km, i.e. 50%.

In the southern branch of the Subandean Zone, a regional E-verging thrust (Mandiyuti Thrust) divides the southern Bolivian Zone into two fold and thrust belts, which differ according to their thrust system geometry. Mainly fault-propagation folds and fault-bend folds characterize the western belt, whereas fault-propagation folds and passive-roof duplexes characterize the eastern belt. Main detachments are located in Silurian dark shale, Lower Devonian shale, and at the base and top of the Middle to Upper Devonian dark shale. The Silurian-Devonian succession is covered by more than 2000 m of upper Paleozoic and Mesozoic sandstone with no potential detachments; in some places it is also covered by several thousand metres of syn-orogenic Neogene sedimentary rocks (Moretti *et al.*, 1996). The base of the foredeep slopes at 2°W. Total shortening decreases southward from 140 km (50%) at 20°S, to 86 km (35%) at 22°S.

The Interandean Zone and Cordillera Oriental are deformed by E-vergent thrusts which involve basement rocks (Kley, 1996), and associated thin-skinned thrusts and back thrusts. Mainly Silurian, Devonian, and Carboniferous strata are exposed in the Interandean Zone. In the Cordillera Oriental, the Neogene thrust system is superimposed on a deeply eroded pre-Cretaceous fold belt that deformed Ordovician anchimetamorphic sedimentary rocks. Shortening is concentrated in the W-verging thrust system of the western part of the Cordillera Oriental and of the Interandean Zone. The Cordillera Oriental is characterized by small Neogene piggyback basins (Fornari *et al.*, 1987; Hérail *et al.*, 1996). Good surface data allowed us to construct some balanced cross sections, according to which total shortening may be estimated between 80 and 100 km. The Altiplano is a complex Neogene intermontane basin deformed by both extensional and compressional tectonics. Surface mapping, seismic reflection data, and drilling information made possible the construction of balanced cross-sections. The total shortening calculated is 20 km and 13 km in the southern and northern parts of the Altiplano, respectively.

The western areas include several morphological units. The Cordillera Occidental includes Plio-Quaternary volcanoes. Its western part is formed by a W-verging thrust system (Muñoz and Charrier, 1996; Fig. 42), characterized by the reactivation of high angle faults and the lack of Paleozoic cover. On the eastern part of the Precordillera, back thrusts limit a blind pop-up structure below the Tertiary deposits (Riquelme and Hérail, 1997). Shortening is close to 18 km. In the Central Valley, within which Late Cretaceous-Paleocene magmatic arc and associated deposits are deformed by a mild Plio-Pleistocene extensional tectonics (Parraguez *et al.*, 1997). The Coastal Cordillera shows low relief constituted by Jurassic - Early Cretaceous magmatic arc rocks. The Chilean margin exhibits a horst and graben topography, and in its central part, a well expressed extensional, asymmetric basin (Muñoz and Fuenzalida, 1997), similar to the Neogene basins located farther N (Von Huene and Scholl, 1991).

Crustal balancing across the Central Andes between 15°S and 18°S (Fig. 2) on the basis of a normal pre-orogenic crustal thickness (according to the location of the Palaeozoic basin and the lack of significant Meso-Cenozoic extension) allows us to calculate 210 km of shortening during the Neogene (Baby *et al.*, 1997). At the latitude of the Arica elbow, shortening is associated with the clockwise rotation of crustal blocks controlled by inherited faults (Fig. 43), due to the compression exerted by the fore-arc zone that behaves as a rigid buttress. These rotations are coeval with the compressional deformation, but the elbow shape of the Bolivian orocline has been acquired prior to this deformation, and is probably of Late Cretaceous or Eocene age (Roperch *et al.*, 1999b).

The Moho shape and the Nazca Plate geometry at this latitude are well constrained by geophysical studies (James, 1971; Cahill and Isacks, 1992; Dorbath *et al.*, 1993; Beck *et al.*, 1996; Zandt *et al.*, 1996). Deep crustal structures are imaged by lower crust reflectors located at different structural levels (Wigger *et al.*, 1994; Allmendinger and Zapata, 1996). The crustal duplexes below the Eastern and Western Cordillera are insufficient to produce the crustal thickening evidenced by geophysical data below the Altiplano and the fore-arc zone. Duplex structures in the lower crust (Lamb and Hoke, 1997) cannot explain the over balanced volume (7216 km^2 in cross-section; Fig. 2), since the lower crust structures have been taken into account in the crustal balancing. Asthenosphere wedge as well as significant volumes of magmatic addition cannot account for the observed thickness (Rochat *et al.*, 1999). The significant tectonic erosion of the Chilean margin (Rutland, 1971; Cloos and Shreve, 1996; Von Huene and Scholl, 1991) and associated extensional deformations suggest that deep crust material removed from the continental edge has been underplated below the fore-arc zone and Altiplano (Schmitz, 1994; Baby *et al.*, 1997).

Timing of Neogene deformations

In the Central Andes, the back-arc thrusting started in the late Oligocene (Sempere *et al.*, 1990; Baby *et al.*, 1997). The first W-vergent thrust motions in the fore-arc zone occurred in the late Oligocene-lower Miocene along the median thrust plane (Garcia *et al.*, 1996). Meanwhile, the Altiplano corresponded to an endorheic basin (Rochat *et al.*, 1998, 1999) situated at the back of the more internal crustal thrust of the Eastern Cordillera. During the upper Miocene, the median thrust plane of the W-vergent thrust system was reactivated (Garcia *et al.*, 1996) and crustal back thrusts produced the partial expulsion of the Altiplano, which represented, therefore, a broad piggy-back basin carried over the crustal duplex of the Eastern Cordillera (Baby *et al.*, 1997). Activity of the Subandean fold and thrust system started at the same period (Gubbels *et al.*, 1993); its eastward propagation accelerated in the Pliocene and continues presently.

Kinematic and dynamic analysis

Tectono-sedimentary studies of the Altiplano (Rochat *et al.*, 1998) indicate a local type isostatic behaviour (deep basin controlled by vertical motion along pre-existing high angle faults). Predicted topography from 10 km of deposits, assuming a normal crust isostatically compensated, is 1.5 km (Rochat *et al.*, 1999). However, no significant absolute subsidence and uplift occurred in the Altiplano during the Neogene. The continuity of the sedimentation in the centre of the Altiplano shows that the topography was archived by filling up of the thick syn-orogenic deposits and progradation over the uplifting borders.

The Neogene filling of the fore-arc extensional basin (Von Huene and Scholl, 1991) is coeval with the sedimentary overloading of the Altiplano, which corresponds to 30% of the volume eroded from the back-arc and the Cordillera Occidental (Rochat, 1999). Timing of both processes indicates that deep tectonic erosion and underplating were able to maintain isostatic equilibrium and consequently the vertical aggradation of the Altiplano level. Along the fore-arc zone, structural traps (like the Altiplano crustal piggyback basin) do not exist. The intensity of Neogene erosion shows that these areas were overcompensated by the deep underplating. The upper Plio-Pleistocene decrease of sedimentation areas in the Altiplano (Rochat *et al.*, 1998) was associated with exorheic drainage. Consecutive minor uplift, as is shown by lacustrine over-deepening and extensional deformations (Lavenu, 1995), show that the equilibrium between superficial sedimentation-erosion and deep erosion underplating was broken. In an ongoing convergence tectonic context, deep up-drive will involve destruction of the Altiplano by erosion and associated collapse.

Andes of Ecuador

The Ecuadorian Andes (1°N - 4°S), are one of the narrowest and most active part of the Andean Belt. It is deformed by NNE-SSW right-lateral transpressive shear zones (Tibaldi and Ferrari, 1992) and underwent an intense Holocene tectonic and volcanic activity. The Dolores-Guayaquil Megashear constitutes an important dextral transcurrent boundary which marks roughly the suture zone between the South American continental margin and the Coastal Block with oceanic basement, accreted during Late Cretaceous-Paleogene times (Juteau *et al.*, 1977; Lebrat *et al.*, 1987; Cosma *et al.*, 1998; Reynaud *et al.*, 1999). Deep geophysical data are not numerous enough to constrain the Moho geometry. Below the chain, the average depth of the Moho is about 50 km (Prévot *et al.*, 1996).

Crustal structures and Neogene deformations

The Ecuadorian Andes are divided from E to W into six morphotectonic units (Fig. 2). The Amazonas foreland basin is deformed by two major NNE-SSW trending, transpressional right-lateral fault zones, which correspond to inverted Mesozoic rift systems (Baby *et al.*, 1999). Positive flower structures were developed along these trends and formed the main oil fields of Ecuador. No Quaternary sedimentary sequences are cropping out in this basin, which seems to undergo uplift presently.

The Subandean Zone is formed by two en echelon NNE-SSW trending positive flower structures (Napo and Cutucú uplifts, Baby *et al.*, 1999), which are still seismically and volcanically active. They result also from transpressional dextral movements, and are separated by a Quaternary pull-apart basin (Pastaza Depression).

The Cordillera Real is a metamorphic belt, intruded by Jurassic batholiths and strongly deformed by wrench tectonics. A W-dipping, high angle reverse fault zone separates it from the Subandean Zone. In the Interandean Valley, thick alluvial, lacustrine and volcaniclastic continental sediments were deposited in several Neogene intermontane basins, controlled by regional strike-slip faults limiting the Interandean Valley (Marocco *et al.*, 1995; Barragán *et al.*, 1996; Hungerbühler, 1997).

The Western Cordillera and Coastal area are part of the allochthonous oceanic terranes accreted to the Andean margin during Late Cretaceous-early Tertiary times (Lebrat *et al.*, 1987; Cosma *et al.*, 1998; Hughes and Pilatasig, 1999; Reynaud *et al.*, 1999). The Western Cordillera is made of oceanic plateau and island arc magmatic rocks and their Cretaceous-Eocene flysch cover, overlain and crosscut by continental arc magmatic rocks. The Coastal area is characterized by four main Neogene fore arc basins (Borbón, Manabí, Progreso, Guayaquil; Fig. 1) related to dextral strike-

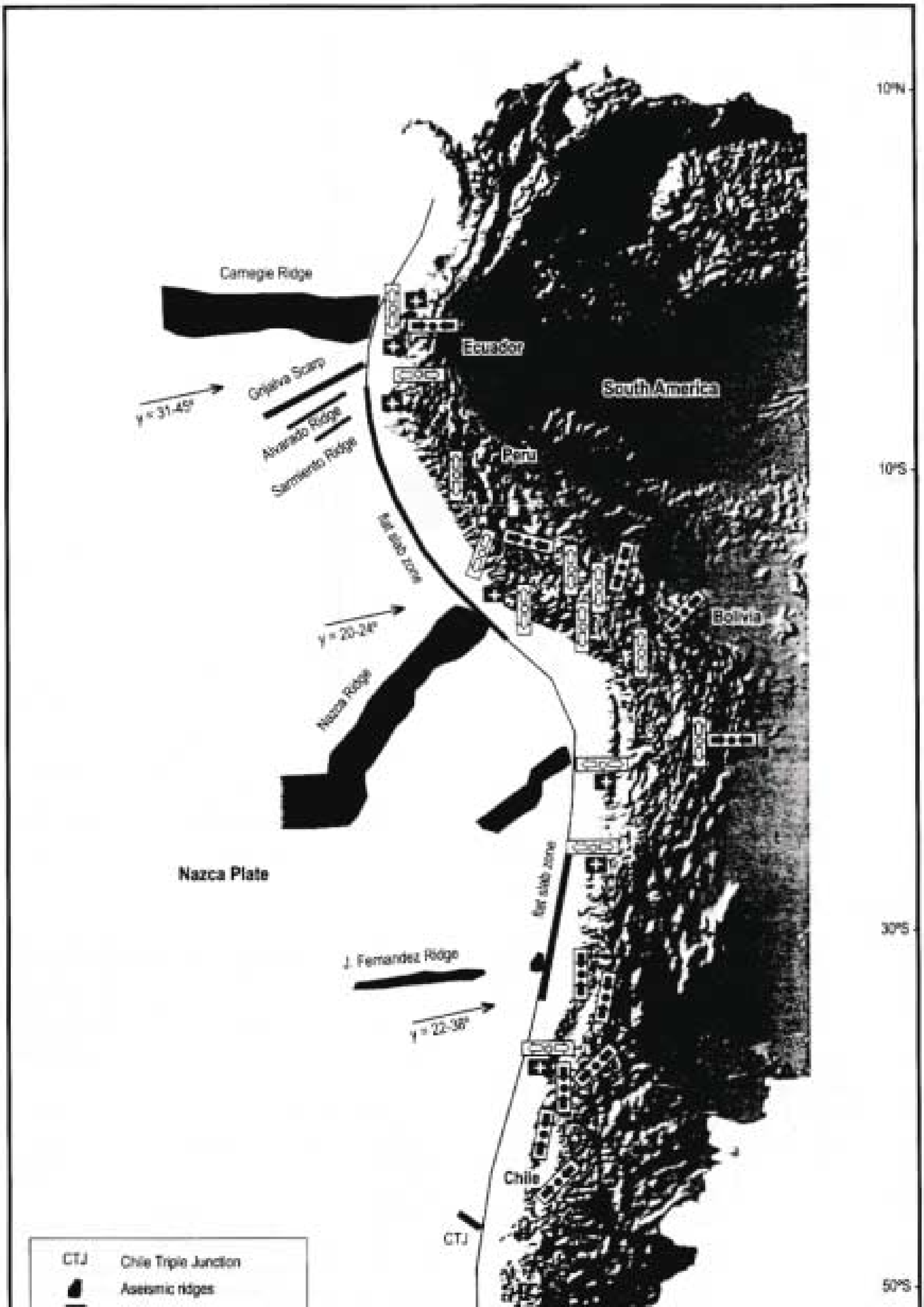


FIGURE 44 - Principal directions of stress deduced from structural analysis of Quaternary and active faults.

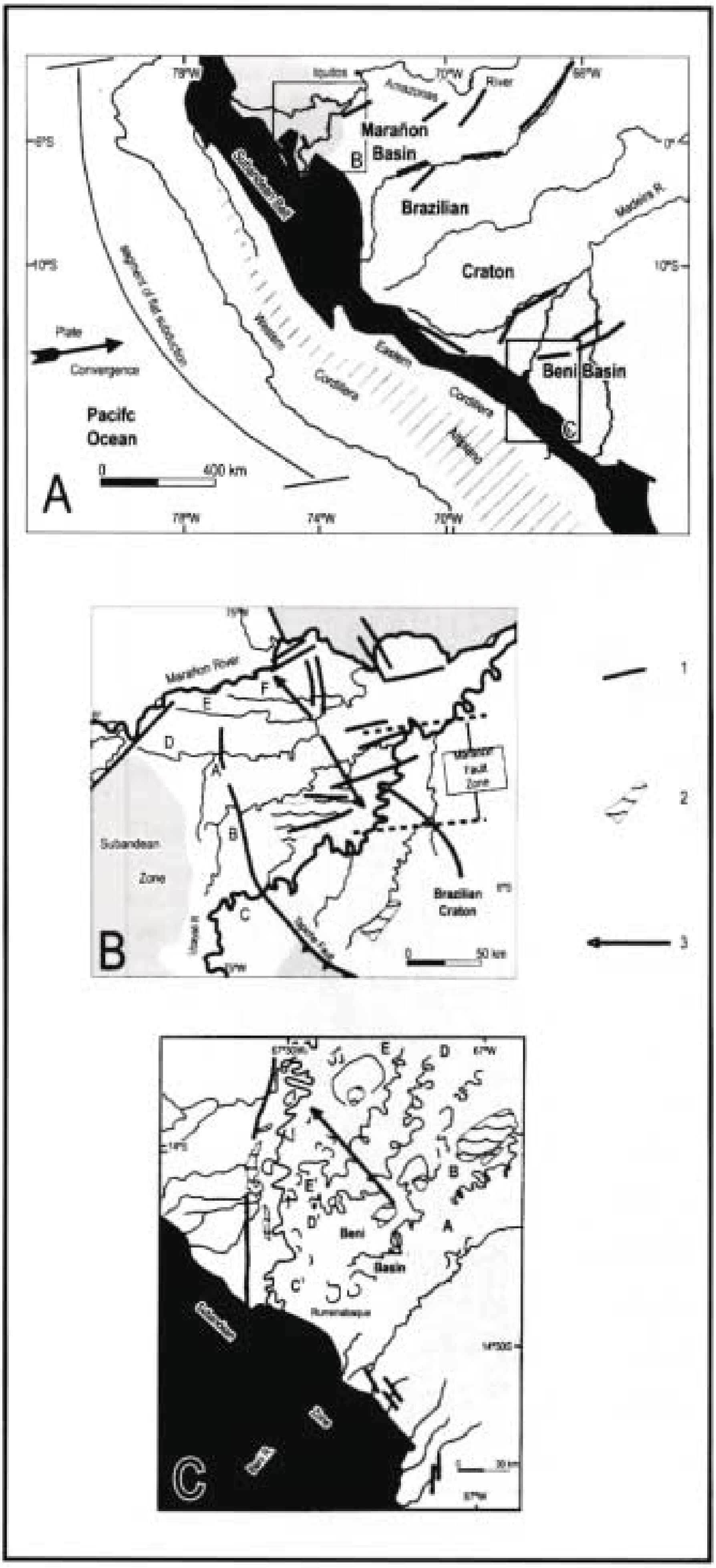


FIGURE 45 - *Geodynamic environment of the Marañon and Beni basins.*
A: Location. B: Rivers and basement structures in the Marañon Basin. Shift of the inflection points from A to C (Ucayali River) and from D to F (Marañon river). 1 - Tapiche Fault, front of Subandean thrust tectonics;
2 - main basement faults;
3 - structural elongated lakes ;
4 - main geographical direction of river shifts.
C: Beni Basin and shifts of the Beni River. A to E show the successive lateral shifts of the river, which correspond to the deflection points C' to E'. Lakes in the Beni Basin are not of tectonic origin.

slip displacements (Deniaud *et al.*, 1999). The Gulf of Guayaquil is the deepest Neogene fore-arc basin. It corresponds to a pull-apart basin developed between the Dolores-Guayaquil Megashear Zone to the E, and the oblique convergent Nazca-South America Plate boundary to the W (Deniaud *et al.*, 1999).

Timing of Neogene deformations

Neogene deformations have been recorded in the fore-arc, intermontane and foreland basins. The creation of the intermontane basins started at about 28 - 26 Ma (Marocco *et al.*, 1995), like the Bolivian Altiplano Basin. In the coastal area, the evolution of the Manabí and Progreso Neogene basins began in the early Miocene. In the Amazonas Basin, stratigraphic data are insufficient to specify the age of the onset of the Neogene foreland basin. It is marked by the eastward wedge of the subaerial Arajuno Formation (Petroproducción seismic information), of probable early Miocene age.

The upper Miocene is characterized by the closure and the piggyback evolution of the intermontane basins (Marocco *et al.*, 1995), and the closure of the Manabí and Progreso fore-arc basins (Deniaud *et al.*, 1999). At the same time, the Amazonas foreland was mildly deformed and invaded by marine incursions (Hoorn *et al.*, 1995).

The Pliocene showed an acceleration of the deformation and marked the onset of the strongest orogenic stage of the Andes, which is still active. The opening of the pull-apart basin of the Gulf of Guayaquil started during the Pliocene, and sedimentation rate reached a maximum (8600 m/Ma in the depocenter) in the lower Pleistocene (Deniaud *et al.*, 1999b). The uplift of the Subandean Zone (Napo and Cutucú uplifts) occurred during this period and continues presently.

Overview of the neotectonics of the North-Central Andes (Ecuador, Peru, Bolivia, and Northern Chile)

The main features of the Andean Cordillera were acquired during the Miocene, and few changes occured since then. However, significant modifications of the topography are produced by neotectonic deformations, resulting in the present-day topography. During this period, altiplano basins are formed or maintained in Ecuador and Bolivia, the Nazca and Carnegie aseismic ridges are introduced in the subduction zone, leading the coast to rise, and the two depressions of the Marañon and Beni basins are individualized, giving birth to the present Amazonas River.

This evolution of the landscape is better approached and understood considering three different aspects of neotectonic studies and methods. The first determines the state of stress (Fig. 44), as deduced from fault analysis. As far as Quaternary terranes are considered, a comparison with the present state of stress can be carried out. The second deals with the vertical movements along the coast, as determined from the study of marine terraces. The third type of neotectonic studies is dedicated to the foreland basins, where river locations and shifts are controlled by neotectonic deformation of the basin surface (Fig. 45).

The convergence vector is oblique with respect to the plate boundary zone. The mode of the oblique accommodation is problematic, specially the relationship between the overriding plate deformations and the subduction. The geometry of the coast and the subduction system suggests that it strongly controls the building of the range. The coastal areas of South America are generally submitted to extensional tectonics, mostly because it lies over the subducted plate without lateral constraints. However, the direction of extension is variable. In areas of oblique convergence and relatively wide coastal zone (comprised between trench and Western Cordillera) it is orthogonal to the direction of plate convergence (Ecuador, southern Peru). In Chile, where there is a narrow belt between the trench and the Main Cordillera, the extension is orthogonal to the margin, and interpreted as related to gravitational post-seismic effects.

In the coastal region of Ecuador, the stress pattern is dominated by a N-S extension (Dumont *et al.*, 1997), due either to the general northward escape of the Andean Block, or, more locally, to the northward increasing obliquity of the convergence, from the Gulf of Guayaquil to Colombia. The triangle-shaped Andean Block accommodates the deformation at the triple junction between the South American, Caribbean, and Nazca plates. At the southern tip of the Ecuadorian Coastal Block, which forms the southern corner of the Andean Block, the Gulf of Guayaquil opened as a result of the right lateral movement of the Andean Block with respect to the South American Plate. This right lateral movement is accommodated along the Pallatanga Fault, which extends northeastwards towards the Interandean Depression (located between the Eastern and Western Cordilleras) and farther N to other fault segments (*e.g.*, Chingual-Sofia Fault). Southwestwards, the Pallatanga Fault extends into the Gulf of Guayaquil, by the means of a system of transcurrent and normal faults. The calculated average Quaternary extensional rate of the Gulf of Guayaquil is of 2.5 ± 1.1 mm/y.

The subduction of the Carnegie Ridge during early and middle Pleistocene is an important parameter of the coastal uplift. Several Quaternary abrasion surfaces at elevations ranging from 7 m to as much as 330 m (*e.g.*, the surfaces of the Tablazos Formation) are observed between the Gulf of Guayaquil and Esmeraldas, suggesting a maximum uplift rate of about 0.2 mm/y during the Quaternary. Along the Pacific coast of Peru, the Quaternary faults evidence a N-S trending extension. In the Pacific lowlands of southern Peru, this state of stress is about neutral, due to a topographic effect related to the proximity of the deep Peru-Chile Trench. On the northern Peruvian Coast, the present-day elevation of the abrasion surfaces suggests an uplift rate of 0.2 mm/y during the Quaternary. In southern Peru, in front of the Nazca Ridge, uplifted marine terraces located at 300 to 700 m high, suggest an average uplift rate of 0.18 mm/y and a maximum uplift rate of 0.7 mm/y for the same period (Macharé and Ortlieb, 1993).

In Chile, in localised coastal areas, which are the closest to the trench (80 to 100 km); the observed state of stress during the upper Pleistocene is an E-W extension. This E-W *stretching is related to uplifted terraces, located over a* crustal bulge due to the subduction. Thus the extension is interpreted as resulting from an accommodation of the rising topography, related to body forces. The E-W trending sxx is s3 (sHmin, Horizontal minimum principal stress, or tensional deviatoric stress), syy N-S trending is s2 (intermediate deviatoric stress) and szz, vertical, is s1 (Hmax, maximum principal stress or compressional deviatoric stress). The state of stress is szz > syy > sxx. In northern Chile (23° to 27°S) the Quaternary marine abrasion surfaces, located at an elevation of 200 m, suggest a maximum uplift rate of about 0.2 mm/y for the Quaternary (Ortlieb *et al.*, 1996).

In the Main Range of Central Andes, present-day stress field and crustal deformations are not homogeneous along strike. In the whole Andean range of Ecuador, the present-day stress field appears to be homogeneous and the Quaternary dominant tectonic regime is an E-W trending compression. Near the trench, the state of stress is s1 = N81°E; in the High Cordillera between 0° and 1°S, the state of stress is N77°E < s1 < N120°E; and in the Subandean Zone, s1 is N99°E (Ego *et al.*, 1996). In the northern part of Ecuador, along the right lateral Chingal-La Sofia Fault, the lateral slip rate displacement is of 7 ± 3 mm/y for the last 37 ka BP (Ego, 1995). In the Interandean Depression, in the restraining bend of the Latacunga Zone, the shortening rate is of 1.4 ± 0.3 mm/y since 1.4 Ma (Lavenu *et al.*, 1995). Along the right lateral strike-slip Pallatanga Fault, the horizontal rate motion is of 4 ± 1 mm/y for the same period (Winter and Lavenu, 1989; Winter, 1990; Winter *et al.*, 1993).

The present-day state of stress in the Peruvian Andes has been deduced from the structural analysis of Quaternary and active faults and seismic data (Sébrier *et al.*, 1985, 1988, Mercier *et al.*, 1992). The crustal deformation of the High Andes is characterized by normal faulting, excepted within the Eastern Cordillera of central Peru, whereas the western and eastern boundaries of the High Andes (fore-arc and foreland) are characterized by thrust mechanisms that indicate compressional deformations.

In the High Andes, two tectonic regimes occur. In the Western Cordillera, recent and active deformations result from a N-S trending extensional tectonics. In the Eastern Cordillera, seismicity and active strike-slip faults result from both a N-S trending extension and an E-W trending compression. In the Subandean Zone, reverse faulting is in agreement with an E-W trending compression. Close to the trench, at the contact between the Nazca and South American plates, focal mechanisms of earthquakes evidence an E-W trending compression roughly parallel to the convergence between the two plates. In southern Peru, the state of stress in the High Andes as well as in the Pacific Lowlands results from a N-S trending extension.

Thus, the state of stress in the Andes of central Peru and those of the Andes of southern Peru may be interpreted as an effect of compensated high topography. However, compressional tectonics affects the High Andes of central Peru but not those of southern Peru.

In central Peru, the stress model is such that the vertical stress szz increases with the topography and the compressional stress sHmax is considered constant and *trends E-W, i.e., roughly parallel to the convergence* direction. In the Western Cordillera of the High Andes, szz becomes 1, Hmax is s2 and Hmin is s3, trending N-S. In the Eastern Cordillera, szz becomes s2, sHmax is s1 and trends E-W, and sHmin is s3 trending N-S. In the Eastern Cordillera, the compressional strike-slip faulting may be explained by an effect of topography, between the high Western Cordillera and the Subandean Lowlands. The Eastern Cordillera being undercompensated, its elevation should be lower in an isostatic equilibrium. The change between the compressional regime in the Subandean Zone and the strike-slip regime in the Eastern Cordillera should take place between 1000 and 2000 m in elevation.

In southern Peru, the state of stress is different in the High Andes; there, szz is s1, sHmax is s2 and sHmin is s3 and trends N-S, *i.e.*, roughly perpendicular to the convergence vector (Sébrier *et al.*, 1985, 1988; Mercier *et al.*, 1992). In the Bolivian High Andes, during the Pliocene (6 - 3 Ma), the tectonic regime was extensional; s3 is sHmax and trends E-W (Lavenu and Mercier 1991). During uppermost Pliocene-lower Pleistocene (3 - 2 Ma) a compressional tectonics affected this region, which is characterized by s1 trending E-W, parallel to the convergence. This tectonic event is characterized by a weak deformation, and by the reactivation of old faults as reverse and strike-slip faults. This stress regime is followed by a nearly coeval N-S trending compressional tectonics. Since lower Pleistocene to Present, the whole range is affected by an extensional tectonics with s3 trending N-S (kilometric normal faults with hectometric throw). In the Altiplano and the High Andes, Quaternary tectonic regime is extensional with sHmin = s3 and trends N-S, sHmax = s2 and trends E-W, and s1 is vertical. As in Peru, this stress field results from body forces due to a compensed high topography. The E-W trending horizontal stress sHmin = s2 is roughly parallel to the convergence direction; szz (s1) increases with the topography due to the range load.

The intermediate zones (*e.g.*, Tarija, 1900 m in elevation) are characterized by two superposed stress regimes. One is a relatively weak strike-slip compressional stress, with s2 vertical, s1 = sHmax, E-W trending, and s3 = sHmin, N-S trending. The other one, more intensive, is an extensional, axial stress, with s1 vertical; s2 trends E-W and is equivalent to s3, which trends N-S. If we admit that the vertical stress szz is the result of the weight of an isostatically compensated topography, the strike-slip state of stress is consistent with the intermediate location of the basin, between the Subandean Zone and the High Andes (Lavenu and Mercier, 1991).

Along the Chilean coast, the Quaternary regime is extensional and of an E-W strikes. This deformation characterizes the westernmost portions of the continental fore-arc, close to the trench axis (80 km). This deformation does not appear to be directly linked to boundary forces due to the convergence, but could be the consequence of co-seismic crustal bending with subduction-related earthquakes. It could be topographic accommodation to the uplift of this part of the coast (body force due to topography), sxx striking E-W becomes s3, syy striking N-S is s2, and szz is s1. The state of stress is such that szz > syy > sxx. This phenomenon could be related to the zones of maximum coupling between the oceanic and continental plates in the Central Andes, which could act as a buttress zone.

The partition of the deformation across the plate boundary zone shows that the tectonic regime of the Quaternary is more complex than previously recognized. In the southern Andes (Chile), as well as in the northern Andes (Ecuador), the Cordilleran segments, linked to large strike-slip faults and high angle convergence obliquity, slides toward the North. A part of the energy, transmitted from the subducting plate to the overriding plate, is absorbed by the free escape of fore arc slivers, parallel to the margin. The lack of important crustal thickening and widening of the range characterizes these parts of the Andes. On the contrary, in the Cordilleran segments linked to low angle convergence obliquity, the progressive stop of the lateral movements is due to buttress zones and the energy is absorbed by the crustal thickening and widening of the range (Bolivia).

The Subandean Zones of the Central Andes are dominated by a compressional stress regime. In the Subandean Zone of Ecuador, the Quaternary stress field is compressional and trends E-W. In the Subandean Zone of central Peru, reverse faulting is in agreement with an E-W trending compression, whereas deformations result from a *N-S trending compression (s1 is sHmax) in the Subandean* Zone of southern Peru. In the Subandean lowlands of Bolivia, deformations are compressional, with sHmax as sxx, is s1, E-W trending.

The Marañon and Beni basins are respectively situated at the northern and southern ends of the Peru-Bolivia Andean segment (Fig. 45A). This segment corresponds to the flat slab subduction of the Nazca Plate beneath the Andes. Specific structures of the foothills of the Subandean Zone control two basins, each one having only one outlet, the Amazonas and Madeira rivers, respectively. The flat surface of these basins shows a complex network of flowing and fossil river traces. These active and abandoned fluvial traces are used, together with neotectonic, seismotectonic and subsurface structural data (Dumont and Fournier, 1994), to determine the neotectonic evolution of the Peruvian and Bolivian foreland basins (Dumont, 1996). The phenomena exemplified below refer to short term neotectonics, occurring during the Holocene (0.1 - 0 Ma).

In both basins, recent directional shifts of the main rivers are controlled by the offset of faults. The Ucayali River flows northwards along a N-S intra-subandean basin, then enters the Marañon Basin where it has been deflected to the NE (Fig. 45B). The successive deflection points shifted upstream and along the foothills. The line joining the deflection points (Fig. 45B, points A, B and C) lies just behind the Andean Frontal Thrust, represented here by the Tapiche Fault (Fig. 45B). Contemporaneously, the Marañon River was deflected to the NE, lined up with the straight, NE trending lower reaches of the Huallaga River, which is controlled by a fault observed on satellite images. In the eastern part of the Marañon Depression, the rivers trend NE-SW, parallel to the strike of the main basement faults of the Marañon Structural Zone (Laurent, 1985). Elongated lakes are situated over structurally downwarped blocks (Dumont, 1993).

Successive shifts of the Beni River (Fig. 45C) show the northward migration of the deflection point made by the Beni River entering the basin. A N-S trending fault crossing the foothill margin controls this downstream increment. The present regional state of stress in the Subandean region is roughly E-W, except in the southern part of the Marañon Basin where it is NE-SW (Assumpção, 1992). Quaternary normal faults displaying a NNW-SSE extension in the distal part of the Marañon Basin, as well as rising of bulges on the eastern margins of both basins are consistent with the present-day state of stress. The interpretation emphasizes that in the distal areas of the Marañon Basin the river traces are guided by topographic lows along tensional faults, or basement blocks, uplifted or downwarped by tensional faulting. Near the piedmont, river shifts are controlled by the increment of fault movements toward the basin. However, the effect of faults in the near piedmont is more difficult to explain than in the distal areas. Topographic effect between the High Andes and the foreland basin (Assumpção and Araújo, 1993) may explain that the geometry of active faults on the foothills piedmont depend also on the local trend of the Cordillera. The interpretation of the successive shifts involves the knowledge of paleoclimate oscillations. It appears that a river moves toward a new formed depocenter at the onset of a wetter period, and can stay in place, *even if tectonic deformation progress, during* relatively dryer periods (Schumm *et al.*, 1998).

In summary, the study of the Recent state of stress in the Andes shows several types of behaviour of the continental plate along the active margin. This behaviour is linked to the dip of the subducted plate, the obliquity of convergence between the oceanic and continental plates, the body forces and boundary forces, the presence or absence of buttress zones in the upper plate, and the possibility for the coastal blocks of free escaping (Fig. 46).

In Ecuador, where the convergence obliquity is very high ($g = 31°$ to $45°$), the Coastal Block is pushed northwards, and is affected by a N-S trending extension. Since the elevation of the Andean range is relatively low, the topographic effect is weak. The range is separated from the Coastal Block by a large strike-slip fault, and an E-W trending compressional stress developed, due to boundary forces.

In Peru and Bolivia, convergence obliquity is relatively low ($g = 20°$ - $24°$). The High Andes, which present the highest average altitude, are affected by a N-S extension, mainly due to the body forces. A compressional regime is observed only along the boundary between the range and the Brazilian Shield (boundary forces).

In central and southern Chile, convergence obliquity is intermediate ($g = 22°$ to $30°$). The fore-arc and intra-arc zones of the Cordillera, the topography of which is lower than in Peru and Bolivia, are affected by a N-S to NE-SW compression.

Regarding the uplifted marine terraces, the type and rate of vertical movements are thought to be relatively independent on the rate and direction of convergence, the convergence obliquity, and the age and dip of the subducted plate (Macharé and Ortlieb, 1993). Conversely, these vertical movements are tightly dependent on the morphology of the subducting plate (aseismic ridges), and on the structure and

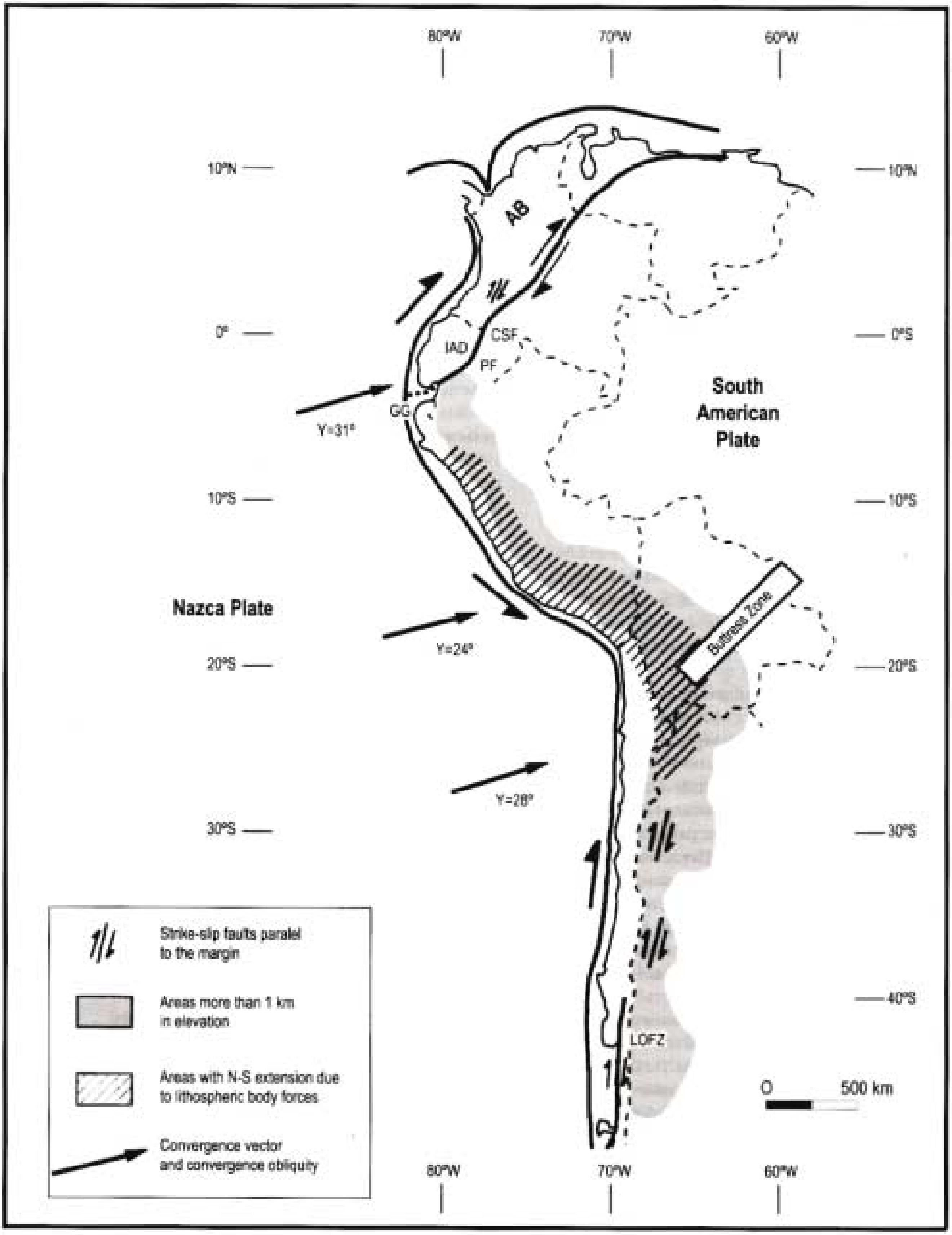


FIGURE 46 - Geodynamic features and relationships between convergence obliquity and Andean Cordillera. AB - Andean Block; CSF - Chingual-La Sofia Fault; GG - Gulf of Guayaquil; IAD - Interandean Depression; LOFZ - Liquiñe-Ofqui Fault Zone; PF - Pallatanga Fault (after Ego, 1995).

density distribution in the overriding plate. In Chile, as well as in Peru, the vertical motion of coastal areas is dependent on the distance from the coast to the trench, and this deformation characterizes the westernmost portions of the continental fore arc, 80 km close to the trench axis. The relationships between Quaternary vertical movements and seismic activity are still poorly understood.

Conclusions : Geodynamic Processes of the Northern-Central Andean Orogeny

Plate Kinematics Framework

The evolution of the central and northern Andean system can be divided into four main periods with different sedimentary, tectonic and magmatic characteristics, indicating distinctive geodynamic situations and convergence directions.

Late Permian - Late Jurassic: Tethyan Period

The Early Mesozoic evolution of the Andean margin is influenced by the Tethyan rifting and evolution (Jaillard *et al.*, 1990, 1995), and is characterized by the onset of a southeastwards subduction along the Colombian, Ecuadorian and Peruvian margins (Fig. 47). After the Late Paleozoic coalescence of Pangea, the Triassic evolution of the northern and central Andes was dominated by an extensional regime responsible for the creation of grabens and the extension of alkaline volcanics. This tectonic context is clearly related to the westward propagation of the Tethyan break-up between Laurasia and Gondwana.

The Early Jurassic evolution of the northern Central Andes was dominated by the destruction of the Late Triassic-Liassic carbonate platform, caused by a general extensional tectonic activity that progressed diachronously southwards. This is thought to have been induced by the rifting of the E-W trending Tethyan system. Meanwhile, no significant absolute motion of the South American Plate occurred relative to the surrounding continental plates (Africa, North America).

Between late Early and early Late Jurassic times, the Tethyan breakline resulted in the opening of NE-trending oceanic-floored rhombochasms (Alpine and central Atlantic oceans, Bernoulli and Lemoine, 1980) linked by E to ENE trending sinistral transform zones (*e.g.*, the Caribbean Transform Zone). The opening of the central Atlantic Ocean began before the late Middle Jurassic (157 Ma, Klitgord and Schouten, 1986) and possibly as early as latest Liassic (180 Ma, Scotese *et al.*, 1988). In the Andes, this period (190 - 140 Ma) was marked by the emplacement of I-type plutons and calc-alkaline volcanics along the NNE trending Ecuadorian-northern Peruvian margin, which should have been coeval with an active subduction beneath this part of the Andean margin. According to the pre-break-up reconstruction, this situation can be interpreted in two ways:

1 - If the Colombian segment was facing continental blocks, the subduction must have involved the new oceanic crust of the Tethyan arm created between the Colombian segment and these blocks (Jaillard *et al.*, 1990; Litherland *et al.*, 1994).

2 - If the Colombian segment directly faced the oceanic paleo-Pacific Plate, subduction was probably active before the Early Jurassic, and the creation of a magmatic arc may have resulted from more rapid subduction, due to an accelerated accretion rate in the paleo-Pacific system.

Whatever the case, the roughly southeastward subduction beneath the Ecuadorian segment must have induced oblique subduction along the Peruvian margin, associated with a strong sinistral strike-slip component, manifested by the creation of the large NW trending south-Peruvian turbiditic pull-apart basin (Vicente *et al.*, 1982) and by transtensional features in the back-arc zones (Sempere *et al.*, 1998; Fig. 47). The Kimmeridgian-Berriasian time-span is transition period. Along the Colombian-Ecuadorian segment, this period was marked by accretions of displaced terranes, compressional deformation, and the end of magmatic activity, while along the Peruvian segment varied tectonic events were associated with the resumption of subduction-related volcanic activity (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995). All this clearly resulted from an important, global-scale geodynamic change. In the west-Tethyan realm (central Atlantic, Alpine oceanic ridges), spreading rates significantly decreased (Olivet *et al.*, 1984; Klitgord and Schouten, 1986; Savostin *et al.*, 1986). If a Tethyan-Colombian oceanic arm did exist, the motion vector of the Phoenix oceanic plate was the sum of the expansion vectors of the Tethyan and Pacific ridges (Duncan and Hargraves, 1984). As a result, slowdown of Tethyan expansion would have induced a northeastward convergence between the Phoenix and South American plates (Duncan and Hargraves, 1984; Jaillard *et al.*, 1990). Moreover, the outpouring of a large oceanic plateau along the Pacific Ridge in Tithonian times may have modified the accretion direction of the East-pacific paleo-plate (Nakinishi *et al.*, 1989).

Early Cretaceous-Paleocene: South Atlantic Period

During this period, the development of the South Atlantic Ocean controlled the westward drift of the South American Plate and the variations in the convergence rate along the subduction zone. These are thought to determine the sedimentary, tectonic and magmatic evolution of the Andean margin. During the Early Cretaceous, the sudden arrival of a great amount of east-derived sands can be interpreted as the result of the westward doming of eastern South America due to the incipient rifting of the South Atlantic Ocean. Although no reliable geodynamic reconstruction is available, the lack of significant tectonic or magmatic activity along the Pacific margin of the South American Plate N of 18°S would indicate a slow, steep-dipping subduction of the paleo-Pacific slab.

The definitive opening of the South Atlantic Ocean at equatorial latitudes during Albian times (Emery and Uchupi, 1984; Scotese *et al.*, 1988) induced the beginning of the absolute westward motion of the South American Plate. Therefore, as noted by various authors (Frutos, 1981; Mégard, 1987; Soler and Bonhomme, 1990), the beginning of compressional deformation along the Peruvian and Colombian segments during the Late Albian (100 - 95 Ma) coincides with the onset of the trenchward motion of the upper plate (Uyeda and Kanamori, 1979; Cross and Pilger, 1982; Jarrard, 1986).

The Albian-Turonian period coincides with a period of high convergence rate and with the mid-Cretaceous magnetic quiet zone (Larson, 1991). In the Central Andean margin, it is characterized by important magmatic activity, a high average subsidence rate (Jaillard and Soler, 1996) and probably significant dextral strike-slip movement (Bussel and Pitcher, 1985). The latter are probably related to the north-northeasterly motion assumed for the paleo-Pacific slab during Late Cretaceous times (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Fig. 48). This convergence direction accounts for the lack of Cretaceous arc magmatism along the NNE trending Ecuadorian margin. The NW trending subduction zone along the Peruvian margin extended probably northwestwards into the oceanic domain as an intra-oceanic subduction zone allowing the development of Cretaceous island arcs, such as those known in western Ecuador.

The Coniacian-late Paleocene interval was marked by a significant slowdown in the convergence rate (85 - 75 Ma), followed by a period of low mean convergence rate between the Phoenix and South American plates (80 - 58 Ma, Soler and Bonhomme, 1990). This period was characterized, however, by the beginning of the Late Cretaceous Andean compressional events. In the northern part of the studied area, significant deformation was restricted to the fore-arc and arc zones. However, the Late Cretaceous and Paleogene tectonic events are coeval with a noticeable decrease of the subsidence rate in the back-arc areas, which favoured detritital deposits, sedimentary hiatus and unconformities. This suggests that during this period of oblique subduction, most of the convergence was accommodated by lateral displacements of fore-arc slivers along the edge of the margin, rather than by shortening and thickening of the upper plate. In the southern Peru and northern Chile, however, deformations seem to have been more important, and a significant increase of the subsidence rate in Bolivia is interpreted as the result of a foreland-type subsidence related to the deformation and tectonic loading of the margin (Sempere, 1994).

Late Paleocene-late Oligocene: Transition Period

The late Paleocene to late Oligocene interval (55 - 25 Ma) is a key period in the whole Andean evolution. Displaced terranes were accreted or obducted along the Colombian segment, important compressional deformation occurred in the Andean realm and sedimentary gaps and unconformities occurred in the eastern domains.

These events coincided with global plate kinematic reorganization (Scotese *et al.*, 1988). In late Paleocene-Eocene times, the convergence of the paleo-Pacific plate changed from N or NNE to NE or ENE (Pilger, 1984; Pardo-Casas and Molnar, 1987), provoking the change from a dominantly dextral transform zone to a nearly normal convergent regime in the Colombian-Ecuadorian segment (Figs. 47 and 48). Such dramatic changes explain how terranes that were previously situated W of the Andean margin, were drifted eastwards and accreted to the northern Andean margin at that time (Jaillard *et al.*, 1995). On the other hand, the more easterly convergence direction allowed subduction to take place beneath the NNE trending Ecuadorian margin and triggered the resumption of arc magmatism in this area by early Eocene times.

A second major reorganization occurred by late Oligocene times, as the Farallón Plate splitted into the Cocos and Nazca plates (Wortel and Cloething, 1981). Convergence direction evolved from ENE to nearly W-E and convergence rate subsequently increased (Pilger, 1984; Pardo-Casas and Molnar, 1987; Tebbens and Cande, 1997; Somoza, 1998; Fig. 48). As a consequence, this period is marked by a significant increase of the orthogonal component of the convergence velocity between the paleo-Pacific oceanic plate and the Andean margin. The subsequent increased coupling along the subduction zone was responsible for a significant eastward migration of the deformed zone, which involved the former arc zone and proximal back-arc areas, *i.e.*, the present-day Eastern Cordillera of Ecuador, Western *Cordillera of Peru and Bolivia, and eventually the Eastern* Cordillera of Bolivia. This suggests a significant decrease of the play of lateral displacement along the Andean margin in the accommodation of convergence, and a correlative increase of the shortening and thickening of the overriding continental plate. Note that this period does not correspond to the subduction of a younger plate (Fig. 49).

Late Oligocene to Present: Pacific Period

From the late Oligocene onwards, the northern central Andean margin was completely controlled by the W to WNW motion of South America and the E to ENE motion of the paleo-Pacific Plate, that determined a roughly E-W couple and a nearly normal subduction system (Fig. 48). During this period, the subducting slab is rejuvenating, the convergence rate is relatively high (Fig. 49) and aseismic ridges arrived in the subduction zone (Fig. 3). This geodynamic pattern, which remains relatively stable and differs significantly from the preceding ones, corresponds to the classical Chilean-type convergent margin.

From 30 Ma onwards, the age of the oceanic plate rejuvenated slightly, becoming probably more buoyant, and favouring, therefore, a low-dipping angle of subduction. Although the convergence rate did not change significantly, late Oligocene-early Miocene times are marked by an acceleration, while the Pliocene is marked by a slight deceleration (Fig. 49). Correlation of these rate variations is difficult to link with specific tectonic events. During this period, the deformed zone significantly migrated eastwards and enlarged, eventually involving crustal-scale thrusting in the Peruvian and Bolivian parts of the chain (Figs. 51 and 52). Due to this large-scale thrust movement, the chain was considerably uplifted, most of the present-day altitude being acquired during the last 8 to 9 Ma. On the other hand,

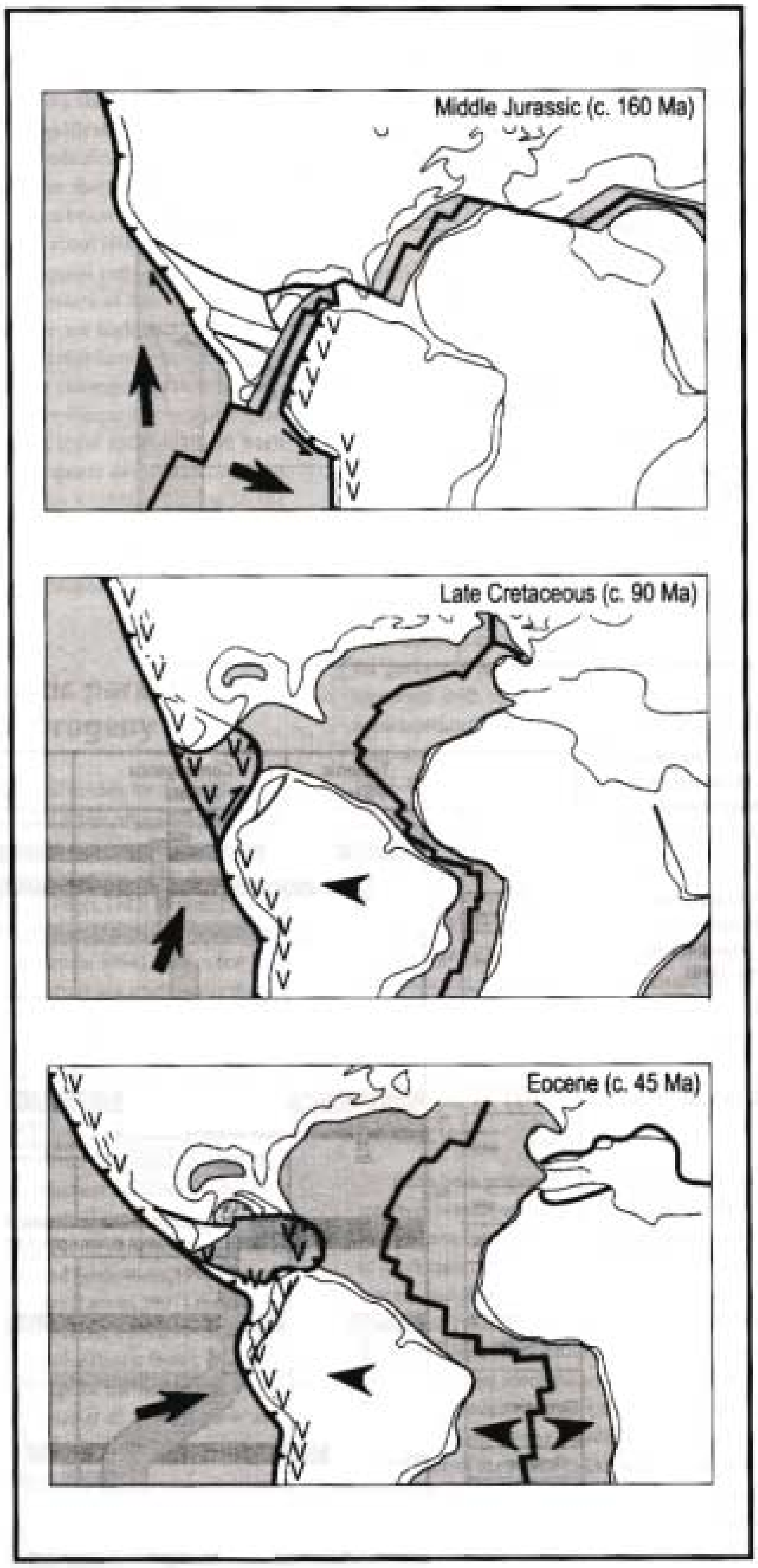


FIGURE 47 - Sketch of the plate tectonic evolution of the Andean margin since Early Mesozoic times (after Jaillard et al., 1990, 1995).

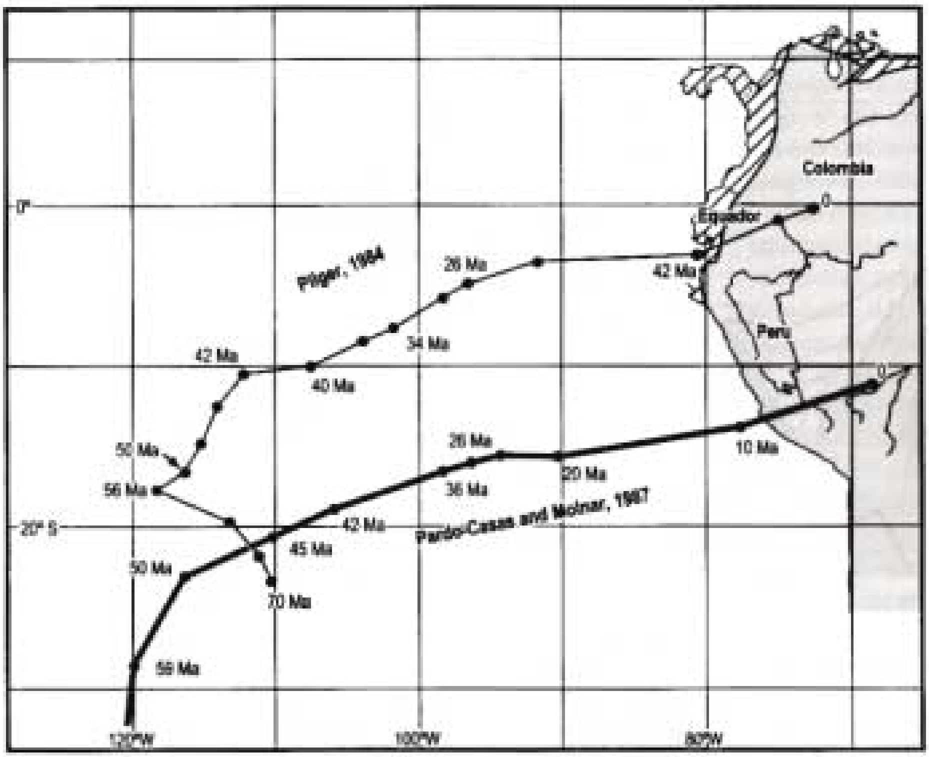
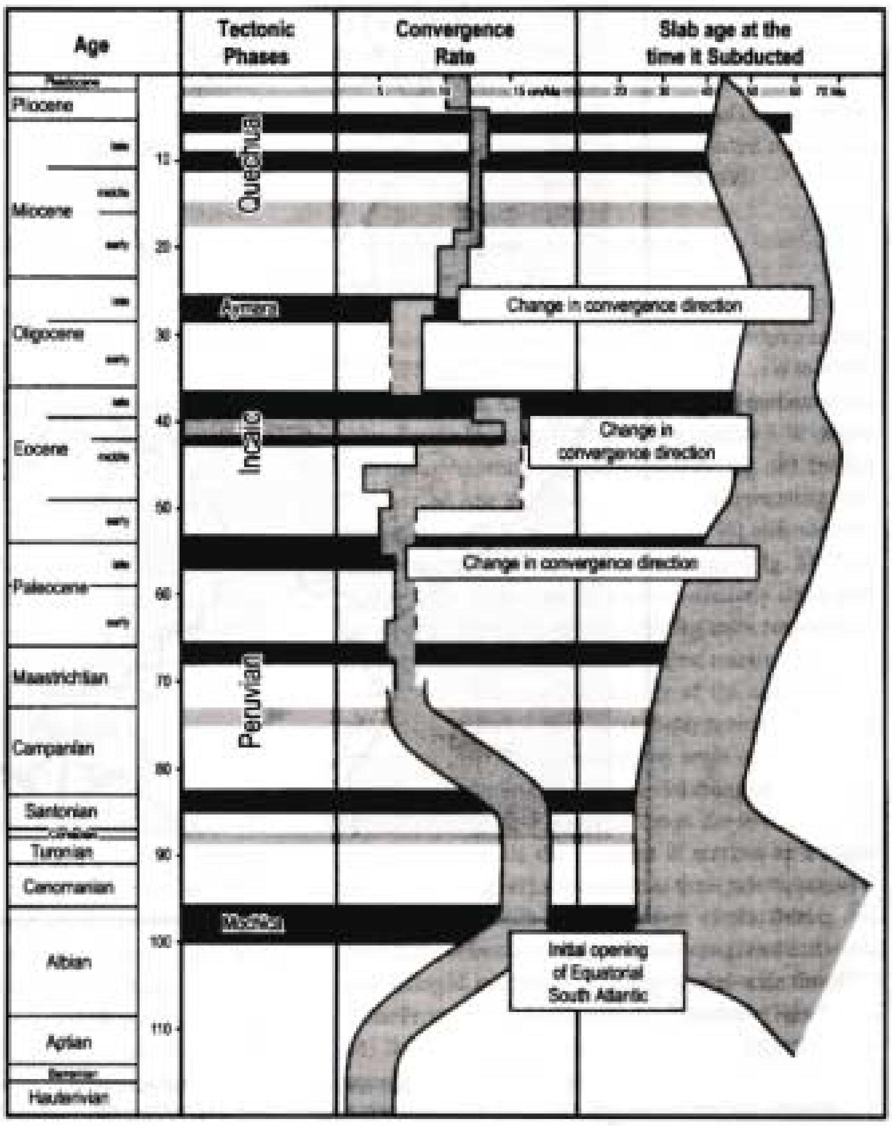


FIGURE 48 - Evolution of the convergence direction of the paleo-Pacific oceanic plate relative to the Andean margin since the latest Cretaceous, (after Pilger 1984; Pardo-Casas and Molnar, 1987).

FIGURE 49 - Evolution of the convergence rate and the age of the slab while subducted, since Middle Cretaceous times (after Soler and Bonhomme, 1990).



subsidence in the eastern basins of Bolivia and southern Peru increased, due to flexural loading. In contrast, subsidence did not increase in Ecuador and even decreased in northeastern Peru (Fig. 39), thus indicating that the tectonic processes differ significantly in both regions. Meanwhile, the fore-arc zones subsided considerably, in such a manner that Pliocene outer shelf deposits are found presently at more than 4000 m below sea-level, indicating an average subsidence rate of about 1000 m/Ma.

In spite of a nearly orthogonal convergence direction, the rate of dextral displacements of fore-arc slivers or terranes of western Ecuador are high (0.5 cm/y). This suggests that such movements may have been much higher in the Late Cretaceous, when convergence rate was much higher (Fig.49) and much more oblique than in the Neogene (Fig. 48). In the same way, local rotations have been significant during the Neogene, demonstrating that contractional shortening was a leading process in the thickening and bending of the Altiplano Orocline, and suggesting that the cumulated amount of rotation may have been significant since the Cretaceous.

Role of kinematic parameters in the Andean Orogeny

Most classical geodynamic models for the origin of the tectonic phases in continental active margins are based on the observation and comparison of various present-day active margins (Uyeda and Kanamori, 1979; Scholl *et al.*, 1980; Uyeda, 1982; Cross and Pilger, 1982; Jarrard, 1986), or through physical modelling (Bott *et al.*, 1989; Whittaker *et al.*, 1992; Cloos, 1993; Shemenda, 1994). Only a few have been elaborated through the study of a single active margin evolution through a long period of time. The study of the Andean margin since earliest Mesozoic times, however, provides some geological constraints on the origin and nature of the tectonic phases of continental active margins.

Plate tectonic reconstructions are poorly constrained for the Late Cretaceous period (Pardo-Casas and Molnar, 1987), especially as regards subduction of ridges, dip of the subducting slab and direction of convergence. However, quantitative approximation of some parameters, such as the convergence velocity (Soler and Bonhomme, 1990) deduced from the global spreading rates (Larson, 1991), the absolute motion of the South American Plate driven by the opening and ridge activity of the South Atlantic Ocean (Nürnberg and Müller, 1991) and the age of the oceanic slab while subducted, calculated by Soler *et al.* (1989), allow us to analyze them in relation to the early tectonic evolution of the northern Central Andean margin.

Age of the subducted slab

Classical models assume that the subduction of a young, buoyant oceanic lithosphere induces a contractional strain in the overriding continental plate (Molnar and Atwater, 1978; Cross and Pilger, 1982; Sacks, 1983). According to Soler *et al.* (1989), the beginning of the contractional period (Albian) and the late Oligocene to Recent contractional phases roughly coincide with the rejuvenation of the oceanic plate subducting at that time. However, the Late Cretaceous and major Paleogene shortening phases occurred during a continuous increase in the relative age of the subducted slab (Fig. 49). Therefore, the lithosphere age of the subducted slab may contribute to the appearance of a long-termed contractional regime, but cannot account for short-termed shortening phases.

Absolute trenchward motion of the overriding plate

As noted by many authors, the opening of the South Atlantic Ocean at the equatorial latitudes during Albian times, which provoked the beginning of the westward shift of the South American Plate, roughly coincides with the initiation of the contractional deformation along the Peruvian-Ecuadorian margin. Thus, this parameter seems to control the long-termed contractional regime of the continental active margin.

As emphasized by Sébrier and Soler (1991) for the late Tertiary Andean contractional phases, only a slight shortening occurs in the Andean retro-arc foreland during the periods of tectonic quiescence, and then most of the westward drift of the South American Plate should be accommodated by an absolute westward overriding of the continental plate over a retreating oceanic slab. On the other hand, the amount of tectonic shortening observed during the contractional phases implies that virtually all the westward drift of the South American Plate is accommodated by the shortening. Therefore, during the contractional phases, the western continental margin of the South American Plate is virtually motionless in an absolute reference frame (*i.e.*, there is a stopping of the slab retreat). This recurrent stopping of the slab retreat, the mechanical origins of which are unclear, might be one of the driving phenomenon of the short-lived contractional tectonic crisis.

Collision of continental or oceanic obstacles

It has been proposed that the arrival in the subduction trench of oceanic or continental obstacles (aseismic ridges, sea-mounts, continental microplates) will lead to blocking of subduction, contractional deformation of the continental margin and plate reorganisation (Scholl *et al.*, 1980; Cross and Pilger, 1982; Ben-Avraham and Nur, 1987). According to Cloos (1993), only continental blocks and oceanic island arcs with a crust more than 15 to 20 km thick or basaltic plateaux of more than 30 km of crustal thickness will provoke a jam in the subduction zone. The current subduction of the 15 km-thick inactive Nazca Ridge results in the extensive subduction erosion of the fore-arc, and local uplift of *c.* 900 m associated with only moderate horizontal compressional stress (Couch and Whitsett, 1981; Macharé and Ortlieb, 1992). Thus, the arrival of moderately high obstacles in the trench seems to have moderate deformational effects on the upper active margin. On the other hand, the accretions of oceanic island arc terranes of Coastal Ecuador (Santonian, late Paleocene, late Eocene) are coeval with contractional phases observed in Bolivia,

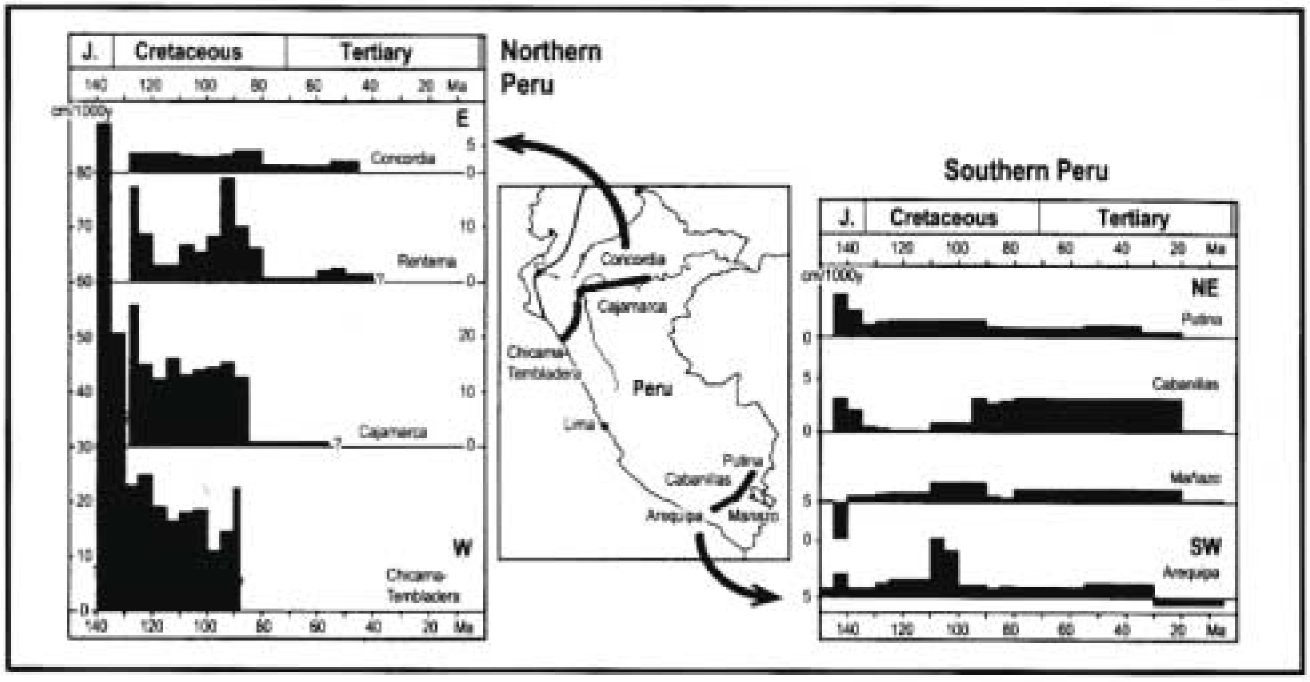


FIGURE 50 - Tectonic subsidence of representative areas of the Peruvian margin (after Jaillard and Soler, 1996).

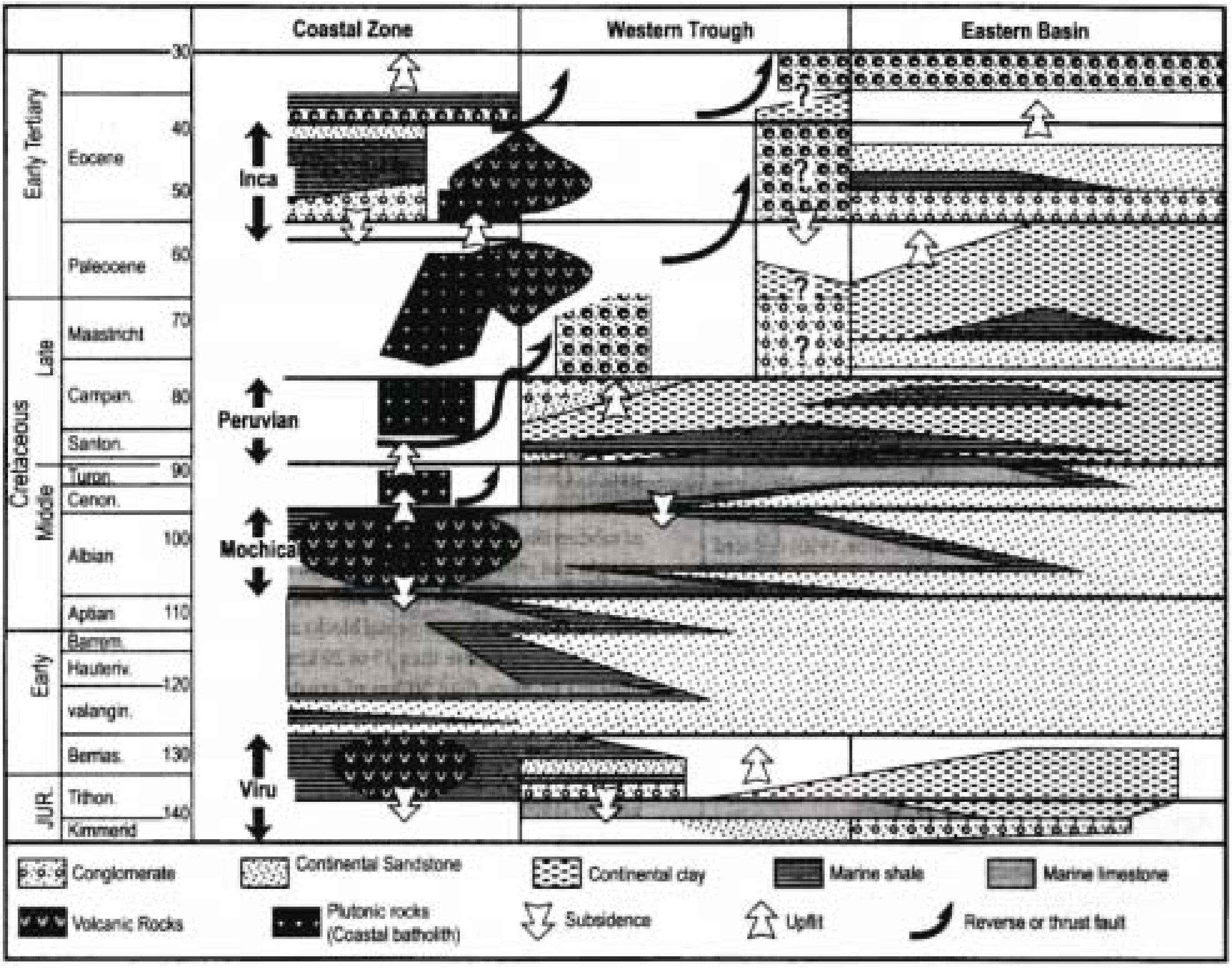


FIGURE 51 - Tectonic, magmatic and sedimentary evolution of the north Peruvian margin between latest Jurassic and Oligocene times (after Jaillard and Soler, 1996).

southern Peru and northern Peru where no collisions are known to have occurred. Thus, in this case, it seems that accretions or collisions of terranes cannot be the cause of regional contractional phases. In the studied area, the fact that contraction took place in non-accretionary settings at the same time as accretions occurred suggests that the accretion events and the coeval contractional phases are consequences of a same global geodynamic mechanism.

Convergence rate

According to Uyeda and Kanamori (1979), Cross and Pilger (1982); Pardo-Casas and Molnar (1987), a rapid convergence between the oceanic and continental plates provokes a compressional stress in the latter. According to Soler and Bonhomme (1990), periods of high convergence rates along the Peruvian margin occurred in Albian-Campanian and late Eocene-early Oligocene times, which coincide roughly with mainly contractional tectonic periods (Fig.49). However, rather than with the convergence velocity itself, the short-lived tectonic events seem to correlate better with changes in the convergence velocity, whichever their sign, either positive (acceleration) or negative (deceleration). If the reconstruction of Soler and Bonhomme (1990) is correct, acceleration occurred in Late Aptian (110 Ma), Late Campanian (75 Ma), early to middle Eocene (50 Ma), latest Eocene (38 Ma), and late Oligocene-early Miocene times (25 - 20 Ma), whereas deceleration occurred in the Late Albian (100 - 95 Ma), Santonian (85 Ma), middle-late Eocene (42 Ma), and Pliocene (4 Ma). All these periods coincide with apparently extensional (Late Aptian, early-middle Eocene boundary) or important contractional tectonic phases. Therefore, short-lived contractional phases as well as extensional tectonic events seem to be mainly controlled by changes in the convergence velocity.

Direction of convergence

The geometry of the geodynamic reconstructions are too poorly constrained to allow a valuable discussion for the Late Cretaceous. The Incaic contractional tectonic phases of late Paleocene (58 - 55 Ma), late middle Eocene (43 - 42 Ma) and late Oligocene age (26 Ma) coincide with successive clockwise rotation in the direction of convergence (Pilger, 1984; Pardo-Casas and Molnar, 1987; Mayes *et al.*, 1990; Tebbens and Cande, 1997). These changes in the convergence direction, which caused successive significant increases of the normal convergence rates seem to have the same effects as those assumed for a convergence acceleration.

Moreover, the important changes in the convergence direction from NNE to ENE by late Paleocene must have induced drastic changes in subduction geometry. The NNE trending Ecuadorian margin changed from a mainly transform to a chiefly convergent regime. This must have induced the eastward drift and accretion of oceanic island arcs along the Ecuadorian margin and the birth of new subduction zones to the W of them (Jaillard *et al.*, 1995). The change in the convergence direction of late middle Eocene age also resulted in a new event of collision of island arcs along the Ecuadorian margin (Bourgois *et al.*, 1990; Hughes and Pilatasig, 1999). Thus, changes in the convergence direction not only control the normal convergence rate, but also play a part in the regional subduction pattern that could in turn influence the tectonic regime. Such changes in the convergence direction during the Paleogene can explain the contemporaneity of the contractional events in non-accretionary settings of the Central Andes and the collisions of island arcs.

Relation convergence rates - subsidence

In northern Peru, periods of slow plate convergence correlate with low subsidence rates (130 to 110 Ma, 75 to 45 Ma, 35 to 25 Ma). Conversely, periods of high convergence velocity are coeval with periods of increased subsidence rate (110 to 85 Ma, 50 to 40 Ma; figs. 49 and 50). This cannot be explained by increased subduction erosion of the deep continental margin (Von Huene and Scholl, 1991), because this latter model is only proved to account for the subsidence of the fore-arc or arc zones, whereas increased subsidence is observed as far as the eastern domain between 110 and 85 Ma (Contreras *et al.*, 1996; figs. 39 and 50). In contrast, these observations are consistent with the thermal model of Mitrovica *et al.* (1989), that assumes that a fast convergence provokes an increase of the subsidence rates along the whole continental margin, through mantle convection (Gurnis, 1992; Stern and Holt, 1994). The lack of such correlation in southern Peru is most probably due to the fact that contractional tectonic events occurred earlier and were stronger than in northern Peru. There, tectonic uplift of the margin by crustal shortening and thickening, and overload tectonic subsidence of the foreland would have prevailed since Senonian times (Sempere, 1994).

Dip of subduction and subduction erosion

The continentward shift of the volcanic front is interpreted classically as a result of the shortening of the continental margin, either by compressive tectonic shortening, or by subduction erosion (Scholl *et al.*, 1980). During Albian and early Late Cretaceous times, the location of the magmatic arc of Peru was stable, indicating that neither significant shortening nor subduction erosion occurred at this time (figs. 20 and 51). The eastward shift of the magmatic arc in the Late Campanian can be explained mainly by the tectonic shortening related to the major Peruvian phase. As a consequence, it seems that no significant subduction erosion took place in Peru before latest Cretaceous, and possibly before Paleocene times, as indicated by the relative stability of the magmatic arc location before this period.

In Eocene times, the ongoing eastward shift of the magmatic belt is associated with its abrupt widening interpreted as the result of a widening of the melting zone in the asthenosphere wedge linked to a decrease of the dip of the Benioff Zone, in turn controlled by the normal convergence velocity (Soler, 1991). The coeval rapid extensional subsidence observed in most of the fore-arc regions by early middle Eocene times is too widespread to result from local tectonic events or paleogeographic effects.

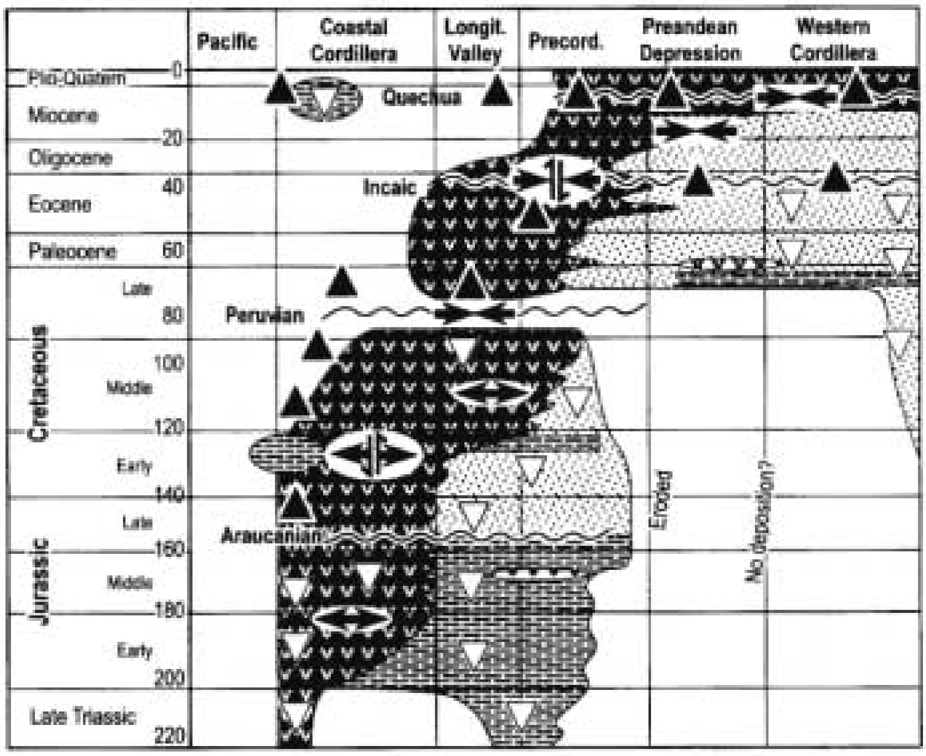


FIGURE 52 - Tectonic, magmatic and sedimentary evolution of the north Chilean margin since Late Triassic times (after Scheuber et al., *1994).*

Since it is associated with the ongoing eastward shift of the magmatic front, we propose that both phenomena are due to the subduction erosion of the margin edge (Von Huene and Lallemand, 1990; Von Huene and Scholl, 1991). Because the widening of the magmatic arc coincides grossly with the assumed initiation of the subduction erosion process, we think that a low-angle subduction plan was a necessary condition for the subduction erosion of the Central Andean margin edge. A low dipping subduction zone would result in increased coupling and shear stress at the contact between the plates, inducing a greater potential of abrasive removal at the base of the overriding plate.

Compressional deformation and subduction erosion

Late early Eocene to early middle Eocene times (50 - 45 Ma) are characterized by the creation of subsiding fore-arc basins along the Andean margin of Peru and Ecuador (Figs. 25, 29, and 51). Such widespread phenomena can be regarded as a consequence of the tectonic erosion of the Andean margin, which is well documented for late Tertiary times (Von Huene and Lallemand 1990; Von Huene and Scholl, 1991). The Peruvian tectonic phase (Steinmann, 1929) is followed by the creation of the Campanian-Maastrichtian Paita-Yunguilla fore-arc basin; the creation of the middle Eocene fore-arc basins is subsequent to the late Paleocene contractional deformation, and the creation of the widespread Miocene fore-arc basins is subsequent to the late Oligocene compressional event. These examples suggest that the creation of fore-arc basins frequently occurs soon after contractional deformational events. Therefore, we propose that, due to increased coupling, tectonic erosion is favoured during contractional deformation periods, whereas the subsequent creation and subsidence of fore-arc basins occur only after the compressional strain has been released. The creation of the latest Cretaceous, Eocene and Miocene fore-arc basins is interpreted, therefore, as a delayed consequence of the tectonic erosion caused by the Peruvian, late Paleocene and late Oligocene contractional phases, respectively.

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