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Did the evolution from oblique to normal convergence contribute to the Andean building ?

Etienne Jaillard

Université Grenoble-Alpes, ISTerre, IRD-CNRS-USMB, CS 40700, 38058 Grenoble Cedex 9, France.

Etienne.Jaillard@univ-grenoble-alpes.fr

Abstract. Although submitted to subduction since at least the Triassic (≈ 230 Ma), the Andean active margin changed in the Late Cretaceous (≈ 80 Ma) from a mainly extension-dominated, marine area to a compressively deformed, mainly emergent zone. At the same time and until the Miocene, convergence direction of the Farallón plate evolved step after step from strongly oblique, to almost orthogonal to the Andean margin. I propose that these major changes are related to each other. Change in a subduction system from an oblique to a more normal convergence direction geometrically implies a decrease in the distance necessary to acquire the dip of the subduction zone. This triggers an increase in the deformation rate of the subducting slab, and therefore, an increase of its elastic modulus, i.e. an increase of forces resisting bending. This should trigger, either the more rapid bending of the subducting slab, or the contractional deformation and subduction erosion of the upper plate. Because the oceanic lithosphere is mechanically stronger than the continental lithosphere, the second consequence is more probable, and would cause the flattening of the subduction zone, and therefore, subduction erosion of the upper plate. Subduction erosion would explain the periodic compressional deformation and subsequent subsidence observed in the forearc zones, while flattening of the subducting slab would explain the landward migration and widening of the magmatic arc, observed on the Peruvian-Ecuadorian margin during the Late Cretaceous-Paleogene (≈ 80 , $55-60$, ≈ 40 and ≈ 25 Ma). From the Neogene onwards, the direction of ocean-continent convergence did not change significantly, and contractional deformation may be interpreted as almost continuous.

Key-words: Tectono-sedimentary evolution, subduction direction, kinematic change, tectonic deformation, accretion of oceanic material.

1. Introduction

Many studies of the parameters controlling ocean-continent subduction systems addressed present-day subduction zones, i.e. steady-state subduction regime, and conclude that subduction kinematics plays an important role in the tectonic deformation of the upper continental plate. Among the leading kinematic parameters are (1) the age of the subducting slab, which controls its temperature and density, and therefore the dip of subduction zone, and the coupling between downgoing slab and upper plate (Molnar and Atwater, 1978 ; Uyeda and Kanamori, 1979 ; Jarrard, 1986 ; Chemenda, 1994 ; Hassani et al., 1997; Royden and Husson, 2009; Capitanio et al., 2011; Martinod et al., 2020); (2) the relative convergence rate, partly controlled by the absolute trenchward motion of the upper plate (Frutos, 1981 ; Cross and Pilger, 1983 ; Jarrard, 1986; Heuret and Lallemand, 2005; Lallemand et al., 2008; Schellart, 2008; Martinod et al., 2010; Horton, 2018; Wolf and Huisman, 2019); (3) the obliquity of convergence, which determines the parallel and normal convergence rates and thus, the deformation style of the upper plate (Jarrard, 1986 ; McCaffrey, 1996 ; Chemenda et al., 2000; Balázs et al., 2021); and (4) the role of mantle convective motion that may induce shear stress at the base of lithospheres and modify the contractional regime in the subduction zone (Mitrovica et al., 1989; Schellart, 2024).

Studies of the evolution of long-lasting past subduction zones are scarcer, and few studies addressed the consequences of a change in the convergence direction in an ocean-continent system. During Late Cretaceous to Paleogene times, subduction of the oceanic plate beneath northern Peru and Ecuador evolved from a strongly oblique to an almost orthogonal convergence direction (Pilger, 1984; Pardo-Casa and Molnar, 1987; Seton et al., 2012; Wright et al., 2016; Müller et al., 2019). Based on these observations, some workers proposed that changes in the convergence rate (acceleration or slowdown) or convergence direction are responsible for the tectonic erosion and deformation of the upper plate (Jaillard and Soler, 1996; Jaillard et al., 2000; Lallemand et al., 2008; Horton, 2018; Herbozo et al., 2020; Quiero et al., 2022). They also pointed out that subduction erosion and correlative forearc subsidence (Von Huene and Scholl, 1991) are coeval with the landward shift and widening of the magmatic arc and therefore, may be related to a decrease of the subduction dip beneath the continental plate.

The Andean active margin of northern Peru and Ecuador is submitted to subduction at least since Triassic times. Nevertheless, it recorded a mainly extensional regime during Jurassic and early Cretaceous times. During the latest Cretaceous (≈ 80 Ma), the tectonic regime of the Andean margin began to change to contractional, causing compressional deformation, terrane accretion and uplift. This paper proposes that compressional deformation, accretion of oceanic terranes, subduction erosion, and landward migration and widening of the magmatic arc may be, at least partly, accounted for by the step by step evolution from an oblique to a grossly orthogonal convergence. This work first discusses the main periods of change in the convergence direction (Late Cretaceous-Paleogene), then proposes an overview of the correlative effects on the Andean margin of Ecuador and northern Peru, and exposes the theoretical geometric and kinematic effects of a change in the convergence direction. The Late Cretaceous-Oligocene evolution of this part of the Andes is eventually proposed to be partly controlled by the step by step evolution toward an orthogonal convergence direction between ≈ 80 and 25 Ma.

2. Geological setting

After a probable hiatus in Late Triassic times (Riel et al., 2018), oceanic subduction occurred continuously beneath the South American continental margin since ≈ 220 Ma (e.g. Jaillard et al., 2000). Jurassic and Early Cretaceous times are marked by the development of a magmatic arc and by a mainly marl-carbonate sedimentation in the backarc basins in a mainly extensional tectonic regime (Jaillard et al., 2000). Mild movements began in Albian times (Bussel and Pitcher, 1985; Mégard, 1987; Jaillard et al., 2000; Carlotto et al., 2023), but actual deformation began in Late Cretaceous times and provoked the progressive emergence and deformation of the whole active margin (Steimann, 1929; Benavídes, 1956; Mégard, 1987; Jaillard et al., 2000; Horton, 2018).

During Cretaceous and Paleogene times, the NW-trending Peruvian margin included well-developed forearc, arc, and backarc zones (Fig. 1A). The latter comprises a western strongly subsiding basin and an eastern moderately subsiding basin, separated by a positive swell (Eastern Cordillera or Marañón “geanticline”; Benavídes, 1999; Jaillard et al., 2000; Mišković et al., 2009; Eude et al., 2015; Baby et al., 2024). The Cretaceous-Paleogene arc zone was located at the eastern foot of the present-day Andean cordillera, while the coeval forearc zone corresponded to the present coast and offshore areas (Fig. 1A).

Ecuador presents a different organization, since most of the present-day forearc and arc zones are floored by oceanic terranes accreted in Late Cretaceous times (Goossens and Rose, 1973; Feininger and Bristow, 1980; Reynaud et al., 1999; Kerr et al., 2002; Jaillard et al., 2009;

Vallejo et al., 2009; Jaillard, 2022; Spikings et al., 2024). The moderately subsiding backarc basin was separated from the accreted oceanic terranes by a positive swell (Eastern Cordillera or “Cordillera Real”, Litherland et al., 1994) (Fig. 1A), which presently hides the lateral relations between the forearc and backarc zones. In both countries, the deformed arc zones and Eastern Cordilleras were thrust since about 25 Ma onto the eastern backarc areas, which, from then on, behaved as a foreland basin (Jaillard, 2022) (Fig. 1B). The tectonic, magmatic and sedimentary features of this evolution can be summarized as follows.

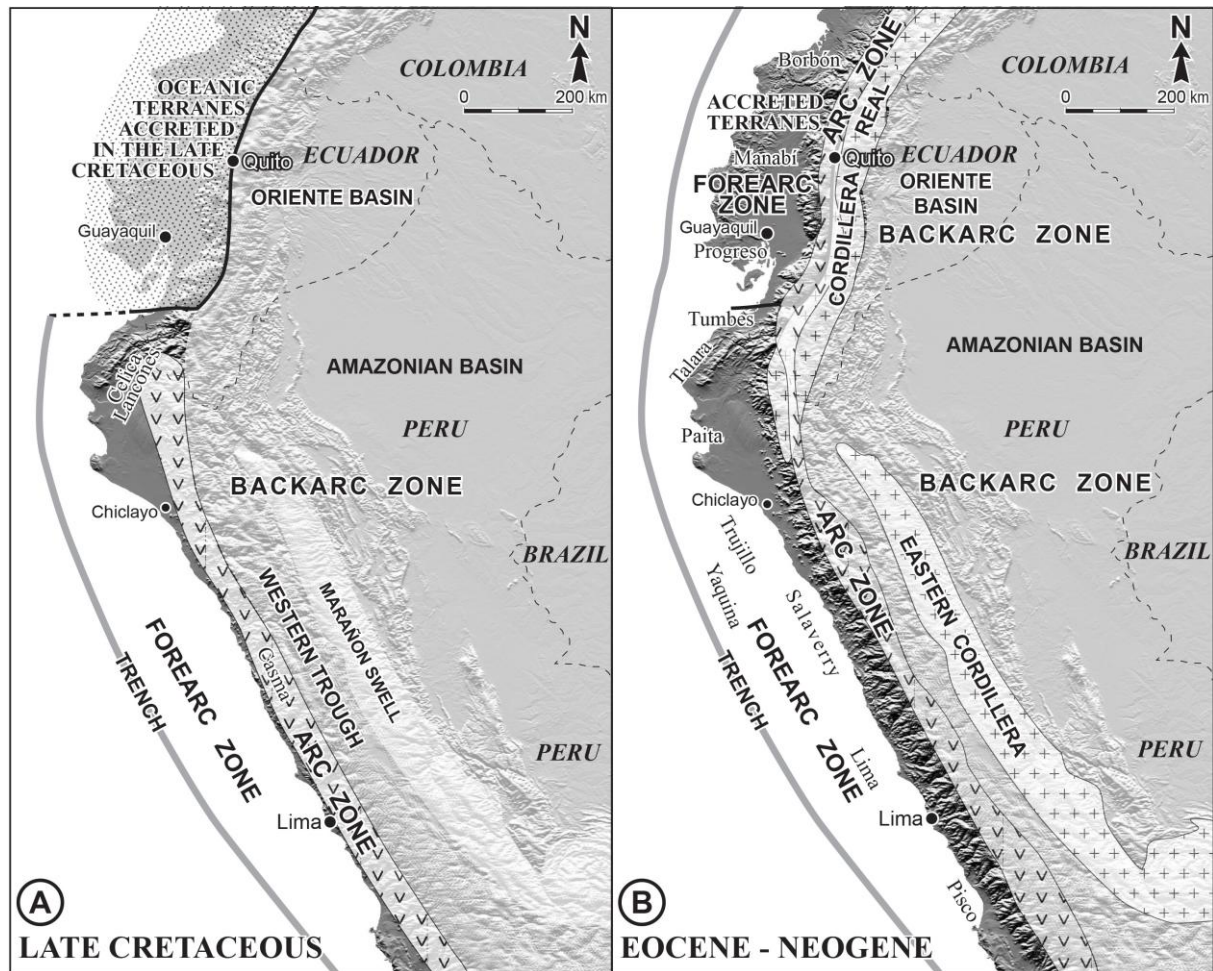


Fig. 1: Main structural zones of northern Peru and Ecuador during the Late Cretaceous (A) and Eocene-Recent times (B). Location of the main sedimentary basins cited in the text.

3. Cretaceous-Neogene evolution of northern Peru and Ecuador

3.1. « Middle » to early Late Cretaceous (≈ 100-85 Ma)

During this time-span, compressional deformation was restricted to the forearc and arc zones, and seems to have been associated with strike-slip movements.

The Peruvian magmatic arc was narrow and the location of the magmatic front remained stable (Cobbing et al., 1981; Soler and Bonhomme, 1990; Jaillard and Soler, 1996) (Fig. 2). During Albian times, the intense activity of this arc was coeval with the development of deep intra-arc basins, interpreted as strike-slip basins (Matalaque and Casma Troughs: Myers, 1975; Soler, 1991; Jaillard, 1994; Polliand et al., 2005; León et al., 2008; Mamani and Jacay, 2015; Jaillard et al., 1996; 1999; Granier and Jaillard, 2018). During the Late Albian (≈ 100 Ma), wrench tectonics provoked the local emergence of the arc zone (Cobbing et al., 1981; Mégard, 1984; Bussel and Picher, 1985; Carlotto et al., 2023), and the creation of a

forearc basin in northwesternmost Peru and southwesternmost Ecuador (Celica-Lancones Basin, Fig. 1, 2) interpreted as a pull-apart basin (Jaillard et al., 1996; 1999). The back-arc areas of Peru and Ecuador recorded subsidence and mild extensional synsedimentary deformation (Mathalone and Montoya, 1995; Mello et al., 1995; Jaillard, 1997; Barragán et al., 2004).

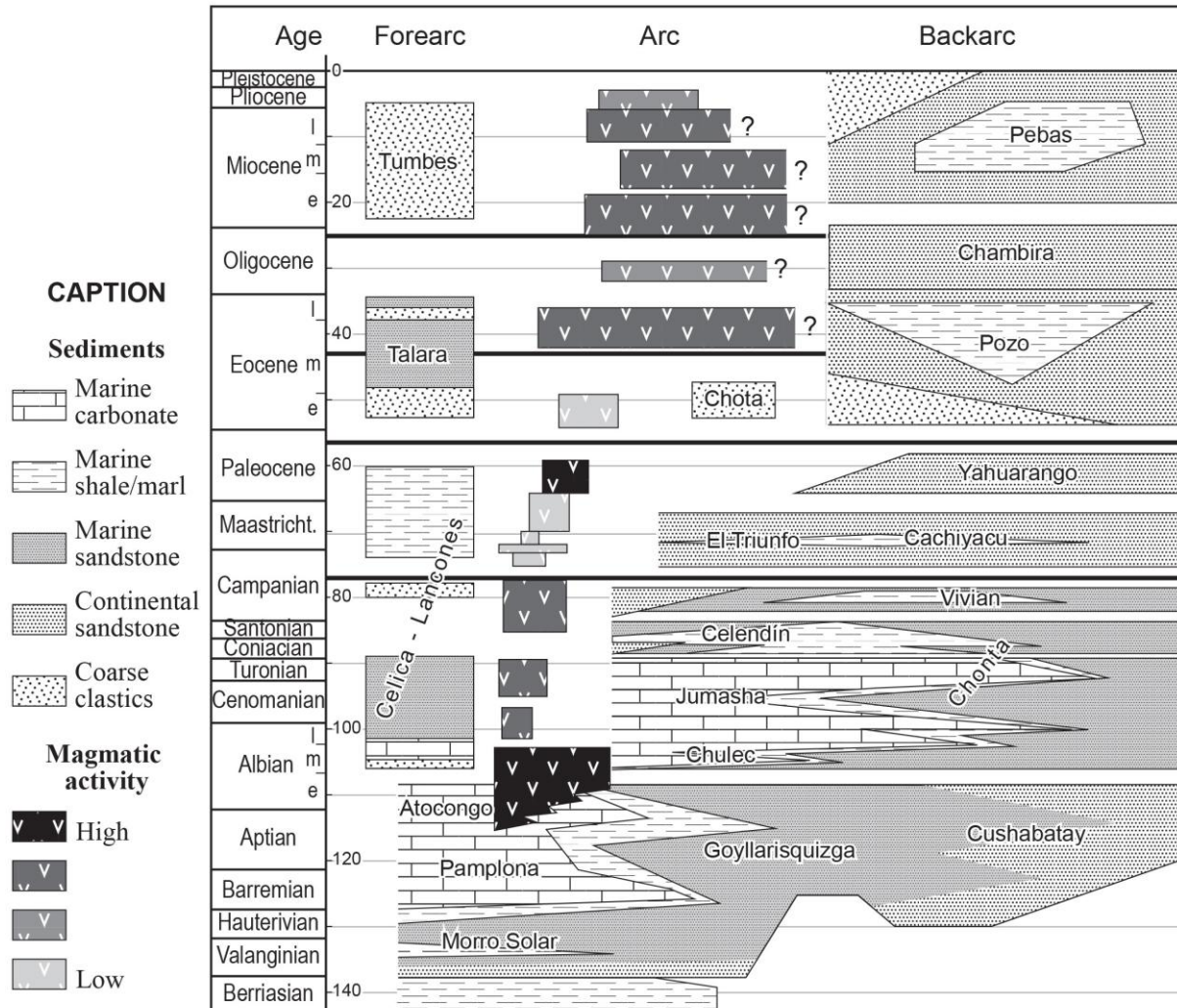


Fig. 2: Sketchy evolution of northern Peru in Cretaceous-Tertiary times (references in the text).

About the Turonian-Coniacian boundary (≈ 88 Ma), a regional, although poorly understood tectonic event led to strike-slip movements and a plutonic gap in the arc zone of Peru (Bussel and Pitcher, 1985; Soler and Bonhomme, 1990). The forearc Celica-Lancones Basin still received Coniacian marine sediments (Petersen, 1949; Reyes and Vergara, 1987) (Fig. 3). Meanwhile, in the backarc areas of Peru and Ecuador, transpressive deformation was marked by the mild inversion of pre-existing extensional structures, local disconformities and fine-grained detrital sedimentation, which brought to end the carbonate shelf sedimentation (Kummel, 1948; Seminario and Guizado, 1976; Mathalone and Montoya, 1995; White et al., 1995; Gil et al., 1996; Jaillard, 1997; Rivadeneira et al., 1999; Barragán et al., 2004; Chacaltana et al., 2005) (Fig. 2, 4). In southern Peru, this tectonic instability may have provoked large-scale sliding of the Albian-Turonian sediments (Callot et al., 2008) and progressive unconformities in Coniacian strata (Carlotto et al., 2023).

3.2. Latest Cretaceous-Paleocene (≈ 85 -55 Ma)

The latest Cretaceous-Oligocene period is a key period, marked by significant changes in the tectonic regime and paleogeography of the Andean margin. This evolution is marked by several tectonic events.

- *Late Santonian-Early Campanian* (≈ 85 Ma). Emergence and mild deformation of the forearc Celica-Lancones Basin are post-dated by a transgression of middle Campanian age (Jaillard et al., 1999; 2005; Valarezo et al., 2019; Jaillard, 2022), which expresses a noticeable subsidence (Fig. 3). This event is also known in northwestern Peru (Paita area: Macharé et al., 1986; Taïpe et al., 2004) and locally in the Andes (Bagua area: Naeser et al., 1991). In Peru and northern Chile, the magmatic arc remained narrow but began to shift eastward (Mukasa, 1986; Soler and Bonhomme, 1990; Hammerschmidt et al., 1992; Scheuber et al., 1994; Capaldi et al., 2021) (Fig. 2). A thermal event has been identified in the Ecuadorian margin (Litherland et al., 1994). In the backarc areas, transpressional inversion of old extensional structures induced weak uplifts, sedimentary hiatuses and mild disconformities, which are sealed by the mid-Campanian marine transgression (Seminario and Guizado, 1976; Mathalone and Montoya, 1995; Naeser et al., 1991; Balkwill et al., 1995; Gil et al., 1996; Jaillard, 1997; Rivadeneira et al., 1999; Chacaltana et al., 2005; Jaillard et al., 2000; 2005) (Fig. 2, 4).

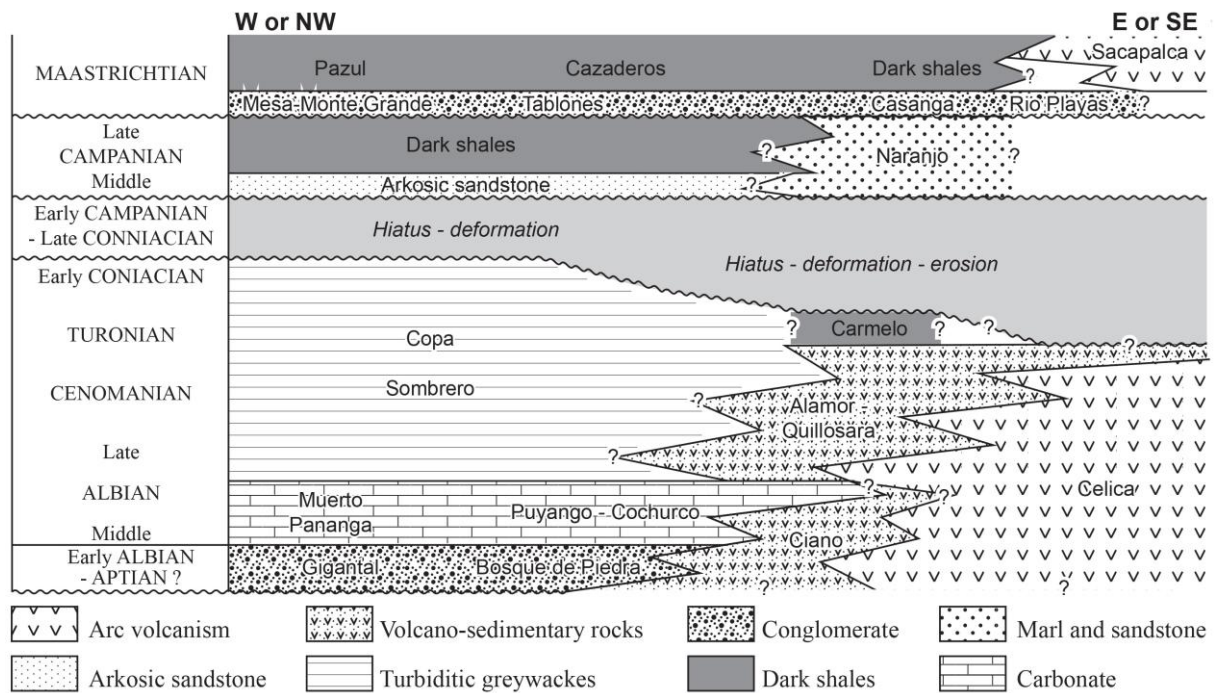


Fig. 3: Stratigraphy of the Lancones-Alamor Basin of northernmost Peru and southernmost Ecuador (mainly according to Jaillard et al., 1996; 1999). Location on Fig. 1A.

- *Late Campanian* (≈ 75 Ma). Late Campanian, coarse-grained quartz-rich deposits are known in the forearc zones of northern Peru and southern Ecuador (Jaillard et al., 1999; 2005; 2008; Taïpe et al., 2004). The rapid uplift the Eastern Cordillera of Ecuador is interpreted as resulting from the accretion of an oceanic terrane (Jaillard et al., 2004; 2009; Spiking et al., 2005; 2010; 2024; Vallejo et al., 2009; 2019; Jaillard, 2022; 2024) (Fig. 3, 4). This collision event resulted in the end of island arc activity on the oceanic terranes (present-day arc and forearc zones of Ecuador) and in the clockwise rotation of the latter (Roperch et al., 1987; Luzieux et al., 2006). The eastern part of the newly accreted oceanic terranes received quartz-bearing sediments (Yunguilla Fm: Toro Álava and Jaillard, 2005; Jaillard et al., 2004; 2009; Vallejo, 2007), whereas quartz-free pelagic cherts deposited on their western part (Guayaquil

Fm: Benítez, 1995; Jaillard et al., 1995; Keller et al., 1997). This indicates that only the eastern part of the oceanic terranes was in contact with the continental margin at that time (Jaillard, 2022). This event is also identified in the backarc areas of Ecuador and Peru, where a widespread sedimentary gap of late Campanian age ($\approx 75-72$ Ma) is thought to result from the uplift related to this tectonic event (Koch and Blissenbach, 1962; Seminario and Guizado, 1976; Raynaud et al., 1993; Jaillard, 1997; 2022; Zuñiga et al., 2021) (Fig. 2, 4).

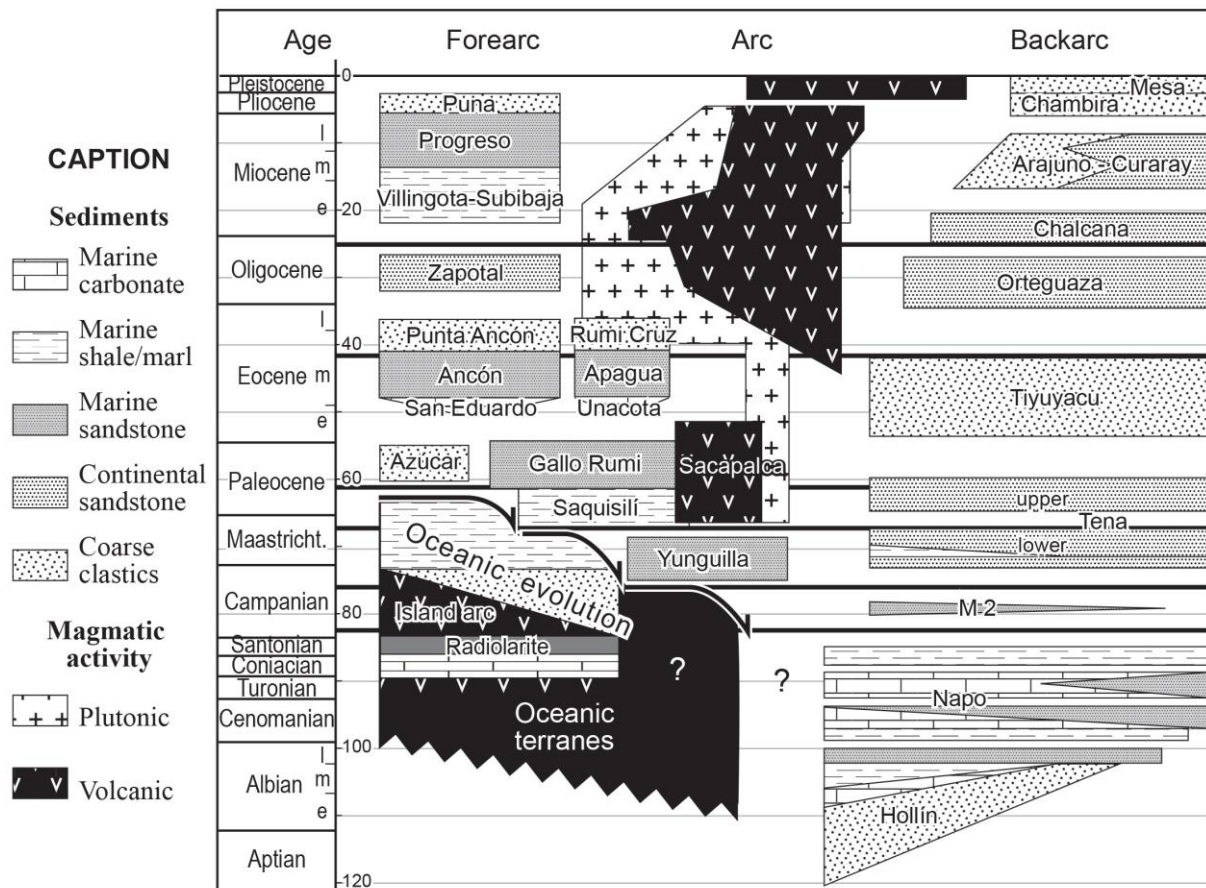


Fig. 4: Sketchy evolution of Ecuador in Cretaceous-Tertiary times (references in the text).

- Late Maastrichtian-Early Paleocene ($\approx 68-63$ Ma). The upper Campanian emergence of the forearc zones of Peru (Talara, Lancones, Fig. 1) was followed by local subsidence, which allowed deposition of marine clastic sediments of Maastrichtian age (Macharé et al., 1986; Morales, 1993; Jaillard et al., 1999; Alemán, 1999) (Fig. 3). The arc zone of Peru was affected by dextral wrench movements (Bussel and Pitcher, 1985; Romero et al., 2002) and a new thermal event occurred in the present-day Andes of Ecuador (70-60 Ma; Litherland et al., 1994), now interpreted as a rapid exhumation event (Spikings et al., 2005; 2010; Vallejo et al., 2009; 2019). In the backarc basins of Peru and Ecuador, sedimentary hiatuses are followed by the resumption of deposition in Maastrichtian times (Seminario and Guizado, 1976; Contreras et al., 1996; Jaillard, 1997; Jaillard et al., 2005) (Fig. 4). This new tectonic event (Spikings et al., 2005; 2010; Vallejo et al., 2009; 2019) is coeval with an accretionary event in Ecuador (Jaillard et al., 2004; 2009; Jaillard, 2022).

- Late Paleocene (≈ 57 Ma). The late Paleocene tectonic event is recorded by a regional hiatus postdated by widespread unconformable conglomerates, known as far as Bolivia (Marocco et al., 1987; Carlotto et al., 2023). In the forearc zones, it is marked by local coarse-grained conglomerates (Azúcar, Mogollón Fms: Macharé et al., 1986; Séranne, 1987; Jaillard et al.,

1995; Ordoñez et al., 2006; Aizprua et al., 2019; Witt et al., 2019; Hernández et al., 2020), and elsewhere by a late Paleocene-early Eocene hiatus (Fig. 4), probably associated with a tectonic event affecting accreted oceanic terranes (Jaillard, 2022; Witt et al., 2022). In the arc zone of Peru, contractional deformation of late Paleocene age was associated with a major magmatic gap and by the incipient widening of the magmatic arc (Cobbing et al., 1981; Soler and Bonhomme, 1990; Noble et al., 1990) (Fig. 2). In the present-day Andean areas of Peru and Ecuador, a new thermal event would have occurred (55-50 Ma) (Litherland et al., 1994) and compressional deformation was sealed by subaerial, unconformable conglomerates of early Eocene age (Jacay et al., 1996; Jaillard et al., 2004; Toro Álava and Jaillard, 2005; Vallejo, 2007). In the backarc areas of Peru and Ecuador, the late Paleocene event is recorded by a long-lasting hiatus (Custódio et al., 2023), or by disconformable, coarse-grained conglomerates of early Eocene age that overly variably eroded beds of Paleocene to Cretaceous age (Rentema-Cajaruco, Tiyuyacu Fms: Naeser et al., 1991; Gil et al., 1996; Rivadeneira et al., 1999; Christophoul et al., 2002; Ordoñez et al., 2006; Moreno et al., 2020) (Fig. 4).

3.3. Eocene-Oligocene (\approx 55-25 Ma)

- *Early to middle Eocene boundary (\approx 50-48 Ma)*. Extensional subsidence of the continental forearc zones led to the creation in the middle Eocene of a belt of marine forearc basins in northern Peru (Salaverry, Talara basins, Fig. 1), the deposits of which may overlay unconformably Late Cretaceous beds (Macharé et al., 1986; Ballesteros et al., 1988; Von Huene et al., 1988; Morales, 1993; Higley, 2004a; Hessler and Fildani, 2015; Viveen and Schlunegger, 2018; Lemgruber-Traby et al., 2020) (Fig. 5). Similar forearc basins are also known on the recently accreted oceanic terranes of Ecuador (Early Progreso, Manabí and Borbón basins, Fig. 1) (Evans and Whittaker, 1982; Jaillard et al., 1995; 2000; Higley, 2004b; Witt et al., 2019; Hernández et al., 2020; Chalampunte, 2020), as well as in central Peru (Lima basin (Fig. 1): Thornburg and Kulm, 1981; Kulm et al., 1988; Genge et al., 2020; Early Pisco Basin: Dunbar et al., 1990; Devries, 1998; 2017; León et al., 2008; Devries et al., 2021; Malinverno et al., 2021). In the eastern basins, compressional inversion of normal faults was followed by the deposition of widespread, unconformable conglomerates of middle Eocene age (Mathalone and Montoya, 1995; Christophoul et al., 2002; Iribarne et al., 2018; Custódio et al., 2023) (Fig. 2, 4).

- *Late Middle to Late Eocene (\approx 42-35 Ma)*. In the forearc zones, mild deformation and sedimentary gaps were sealed by disconformable late Eocene to Oligocene sandstones (Becerra et al., 1990; Benavídes, 1999; Fildani et al., 2008; Witt et al., 2019) (Fig. 5, 6). Arc magmatism was marked in Peru by a chemical change and by a drastic widening of its activity (Soler and Bonhomme, 1990; Soler, 1991; Carlier et al., 1996; Bissig et al., 2008) (Fig. 2). In the present-day Western Cordillera (arc zone) of Peru and Ecuador, the development of east-vergent fold and thrust belts was associated with significant uplift (Cobbing et al., 1981; Mégard, 1984; Carlier et al., 1996; Spikings et al., 2005; 2010; Toro Álava, 2007; Vallejo et al., 2019), and was postdated by unconformable conglomerates and volcanic rock of late Eocene age (Noble et al., 1990; 1999; Naeser et al., 1991; Mégard et al., 1996; Toro Álava and Jaillard, 2005; Vallejo, 2007; Custódio et al., 2023) (Fig. 2, 4, 5). In the eastern basins, reduced subsidence provoked widespread sedimentary gaps or reduced sedimentation in late Eocene-Oligocene times (Dashwood and Abbots, 1990; Mathalone and Montoya, 1995; Contreras et al., 1996; Christophoul et al., 2002), although marine influences are locally recorded (Pozo Fm: Roddaz et al., 2010; Antoine et al., 2016) (Fig. 2, 4).

- *Late Oligocene (\approx 29-25 Ma)*. The late Oligocene tectonic event is better expressed in southern Peru and Bolivia. The forearc zones of Peru and Ecuador recorded a widespread sedimentary hiatus, followed by the unconformable deposition of coarse-grained deposits

(Fig. 5). The arc zone of northern Peru was marked by a 30-26 Ma magmatic gap, but compressional deformation was gentle (Clark et al., 1990; Sébrier and Soler, 1991; Benavides, 1999). The proto-Andes recorded a slight thermal event in Ecuador ($\approx 35-25$ Ma) (Litherland et al., 1994) and pre-Miocene unconformities in Ecuador and Peru (Noble et al., 1990; Mégard et al., 1996; Steinmann et al., 1999). Late Oligocene movements were recorded in the backarc basins by angular unconformities and coarse-grained deposits (Arajuno Fm: Christophoul et al., 2002) and by a strong increase of the subsidence rate (Dashwood and Abbotts, 1990; MATHALONE and Montoya, 1995; Contreras et al., 1996; Custódio et al., 2024), interpreted as the beginning of flexural subsidence (Jaillard, 2022) (Fig. 2, 4).

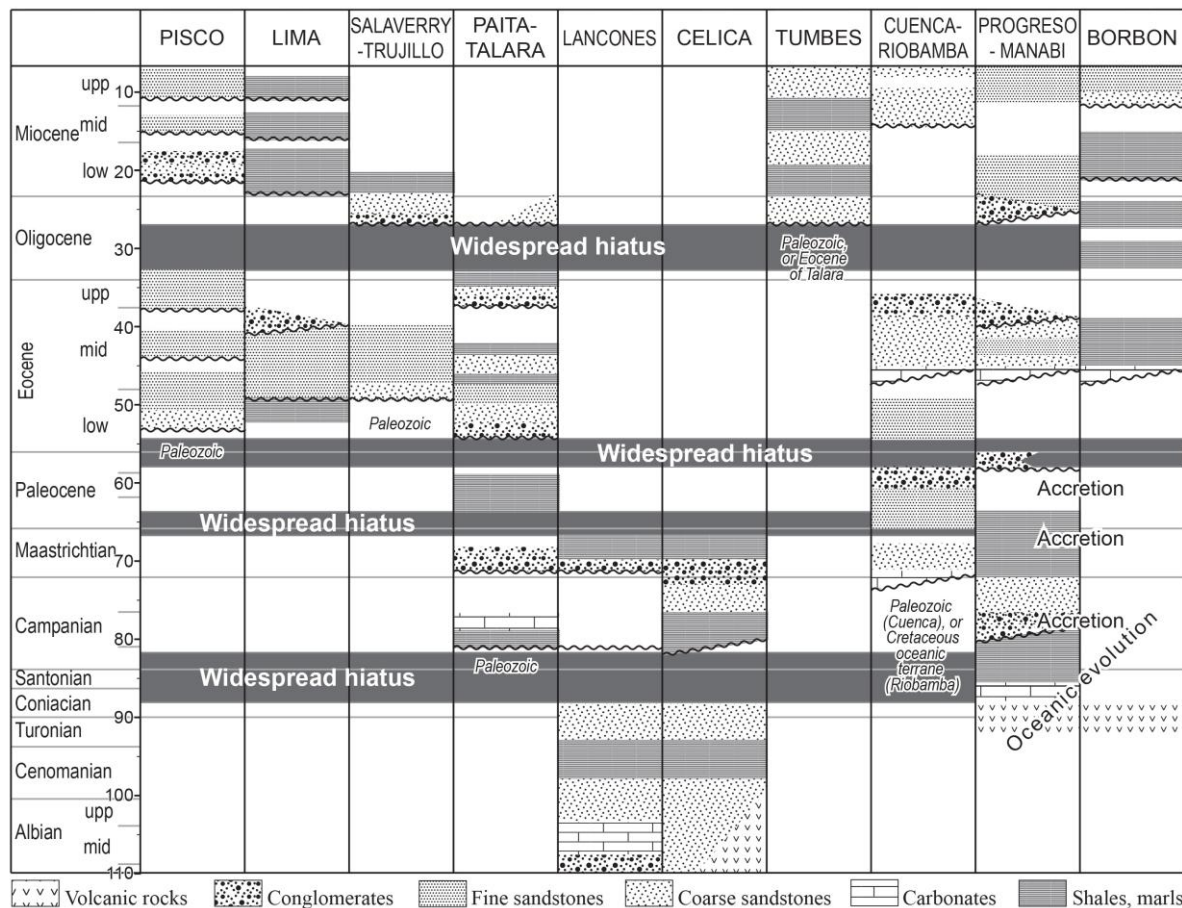


Fig. 5: Chronostratigraphic sketch of the forearc areas of northern Peru and Ecuador since the Late Cretaceous.

References : Pisco : Bosio et al., 2004; DeVries, 1998; 2017; DeVries et al., 2021; Dunbar et al., 1988 ; 1990 ; León et al., 2008; Lima : Thornburg and Kulm, 1981 ; Genge et al., 2020 ; Salaverry-Trujillo : Genge et al., 2020 ; Kulm et al., 1988 ; Moberly et al., 1982 ; Suess et al., 1988 ; Thornburg and Kulm, 1981 ; von Huene et al., 1985 ; 1988 ; Paita-Talara : Fildani et al., 2008 ; González, 1976 ; Marsaglia and Carozzi, 1990 ; Morales, 1993 ; Séranne, 1987 ; Taipe et al., 2004 ; Lancones-Celica : Chalco, 1955 ; Jaillard et al., 1996 ; 1999 ; 2005 ; Reyes and Vergara, 1987 ; Valarezo et al., 2019 ; Winter et al., 2010 ; Tumbes : León, 1983 ; Cuenca-Riobamba : Bristow, 1973 ; Bristow and Hoffstetter, 1977 ; Faucher et al., 1971 ; Steinmann et al., 1999; Hungerbühler et al., 2002 ; Jaillard et al., 2004 ; Progreso-Manabí : Benítez, 1995 ; Jaillard et al., 1995 ; Ordoñez et al., 2006; Aizprua et al., 2019 ; Borbón : Evans and Whittaker, 1982 ; Chalampiente, 2020.

3.4. Miocene to present ($\approx 25-0$ Ma)

Since the Late Oligocene, compressional deformation went on throughout the paleo-Andes and progressively migrated eastward.

In the forearc zones, the late Oligocene tectonic event was followed by the creation of a new belt of marine forearc basins, allowing early to middle Miocene beds to be deposited unconformably on Eocene to Paleozoic rocks (Tumbes, late Progreso, Borbón basins; Fig. 1) (Evans and Whittaker, 1982; León, 1983; Macharé et al., 1986; Ballesteros et al., 1988; Benítez, 1995; Dunbar et al., 1990; Viveen and Schlunegger, 2018; Witt et al., 2019; Bosio et al., 2020; Chalampunte, 2020; Fig. 5). From early Miocene onwards, compressional deformation remained mild. Arc volcanism resumed in the early Miocene; the location and width of the magmatic arc remained grossly stable in Peru (Clark et al., 1990; Soler and Bonhomme, 1990; Benavídes, 1999; Bissig et al., 2008), but it migrated eastward and enlarged in Ecuador (Chiaradia et al., 2004; Schütte et al., 2010; Fig. 4) and northern Chile (Hammerschmidt et al., 1992).

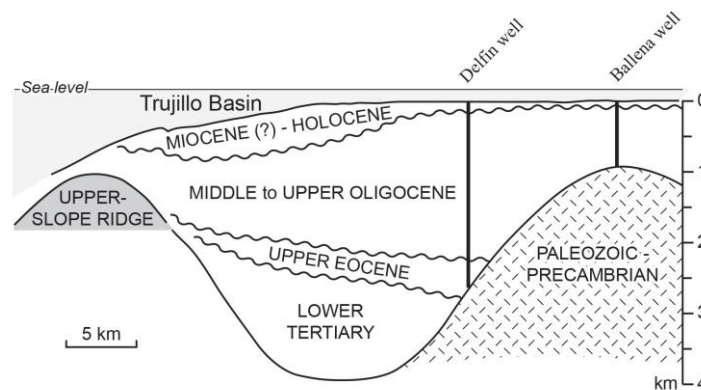


Fig. 6: Sedimentation in the Trujillo Basin (offshore northern Peru, location of Fig. 1B), after Thornburg and Kulm (1981), showing main unconformities in the late Paleocene-early Eocene ?, late Eocene-early Oligocene and early Miocene times.

The Miocene-Present interval was marked by numerous, variably intense compressional tectonic events, dated at ≈ 22 , 17-15, 12-10, 8-7 and 2-1.5 Ma (Fig. 2, 4, 5) (e.g., Sébrier and Soler, 1991; Benavídes, 1999; Spikings et al., 2010). These deformation events affected mainly the present-day Andes (≈ 22 -8 Ma), before to shift toward the eastern basins (≈ 10 -0 Ma) (e.g. Eude et al., 2015; Moreno et al., 2020; Jaillard et al., 2022; Baby et al., 2024). The forearc zones were marked by widespread hiatuses or discontinuities near the early-middle Miocene boundary (17-14 Ma) and by a new hiatus and discontinuity in the latest Miocene-Pliocene (7-5 Ma, Fig. 5, 6) (Dunbar et al., 1990; Devries, 1998; Deniaud et al., 1999; Hernandez et al., 2020).

Within the present-day chain, a late early Miocene magmatic gap (Lavenu et al., 1996; Chiaradia et al., 2004) was followed by the creation of a belt of intramontane basins of middle Miocene age, which closed ≈ 8 Ma ago (Marocco et al., 1995; Steimann et al., 1999; Hungerbühler et al., 2002; Moreno et al. 2020; 2022) (Fig. 2, 4). Rapid uplift of the eastern parts of the chain began in the early Miocene (≈ 22 Ma) and resumed in the late Miocene (≈ 12 Ma) (Laubacher and Naeser, 1994; Hoorn et al., 1995; Steinmann et al., 1999; Spikings et al., 2010; Vallejo et al., 2019; Moreno et al. 2020; Custódio et al., 2024). It resulted from the thrusting of the Eastern Cordillera onto the foreland basin (Naeser et al., 1991; Rivadeneira et al., 1999), which triggered the development of the Subandean Fold and Thrust Belt 10-8 Ma ago (Alemán and Marksteiner, 1993; Mathalone and Montoya, 1995; Benavídes, 1999; Baby et al., 2013; Eude et al., 2015; Iribarne et al., 2018; Moreno et al., 2020; Baby et al., 2024). In the eastern basins, this may have triggered flexural subsidence and deposition of widespread brackish to marine deposits (Pebas Fm: Roddaz et al., 2010; Wesselingh et al., 2006; Hoorn et al., 2010; Antoine et al., 2016) (Fig. 2, 4). The eastern

foreland basin is then deformed by compressional structures in the latest Miocene-Pliocene (Gil et al., 1996; Rivadeneira et al., 1999; Baby et al., 2024).

3.5. Synthesis

From this overview, several features arise. First, a comparable tectonic evolution is recorded in Ecuador and northern to central Peru, although the former underwent the accretion of oceanic terranes, whereas the latter did not, except, maybe, northernmost Peru. Although tectonic events may be of different nature and intensity, they appear to be grossly coeval in both areas. There, major tectonic deformation occurred in the Middle Campanian (≈ 80 -75 Ma; Peruvian phase of Steinmann, 1929), late Maastrichtian (≈ 68 Ma), late Paleocene (≈ 58 Ma), middle Eocene (≈ 40 Ma; Inca phase of Steinmann, 1929) and late Miocene times (≈ 10 Ma; Quechua 2 phase of Mégard, 1984). Mégard (1984) identified a Quechua 1 phase (early Miocene, 20 Ma) that may be represented in the study area by the widespread hiatus of Oligocene-early Miocene age. This suggests that terrane accretion is not the main driving mechanism of Andean tectonic events in the northern Andes. Rather, a phenomenon common to both kinds of margin (with or without accretion), is responsible for the observed tectonic events. I propose that the origin of this phenomenon must be sought in the subduction dynamics.

Second, forearc zones show a cyclic evolution, with alternation of creation of forearc basin and subsidence, and emergence and deformation of these basins. Although the Cretaceous evolution of forearc zones is poorly known, the evolution of “middle” Cretaceous forearc basins seems to be controlled by transcurrent movements. Afterwards, a first generation of forearc basins created in Campanian times (≈ 80 Ma; Celica-Lancones, Yunguilla) and became emergent in the Paleocene. The formation of new belt of forearc basins in the middle Eocene (≈ 40 Ma; early Pisco, Salaverry, Talara, Ancón, Manabí, Borbón) is followed by a widespread hiatus of Oligocene age. Finally, a third generation of basins occurred in the Miocene (≈ 25 -20 Ma; late Pisco, Tumbes, Progreso; Fig. 5), most of which became emergent in Pliocene times (≈ 5 Ma).

Finally, in central and northern Peru, the location of the magmatic arc remained stable during the “middle” to Late Cretaceous (≈ 100 -75 Ma), and it migrated eastward in the latest Cretaceous-Paleocene interval (≈ 75 -55 Ma), before to enlarge in Eocene to Miocene times (≈ 55 -30 Ma) (Fig. 2). Because of the Late Cretaceous oceanic evolution and subsequent accretion of the terranes in Ecuador, the evolution of the Ecuadorian magmatic arc is somewhat different during the Late Cretaceous, but it is comparable from the Eocene onwards (Fig. 4).

4. Convergence direction of the Farallón-Nazca plate with respect to northern Peru and Ecuador since ≈ 80 Ma

Since the Late Cretaceous, the convergence direction of the Farallón, then Nazca plate evolved step after step from a grossly northward direction, almost parallel to the Peruvian margin, to a nearly east-ward direction, roughly orthogonal to the average trend of the present-day Andean margin (Fig. 7). Significant changes in the convergence direction occurred at ≈ 85 -82 (late Santonian-early Campanian), ≈ 59 -56 and/or ≈ 55 -50 (late Paleocene-early Eocene), ≈ 42 (middle Eocene) and ≈ 25 Ma (latest Oligocene) (see Pilger, 1984; Pardo-Casas and Molnar, 1987; Somoza and Ghidella, 2012; Seton et al., 2012; Wright et al., 2016; Müller et al., 2019) (Fig. 7, 8).

Although poorly constrained for the Late Cretaceous, most reconstruction assume a NW-directed migration of the Farallón plate along South America during this period, thus

implying a strongly oblique subduction (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Lithgow-Bertelloni and Richards, 1998; Somoza and Ghidella, 2012; Seton et al., 2012; Müller et al., 2019) (Fig. 8). This view is supported by geological interpretations of the arc and forearc zones of Peru that are affected by protracted transcurrent motions during the Cretaceous.

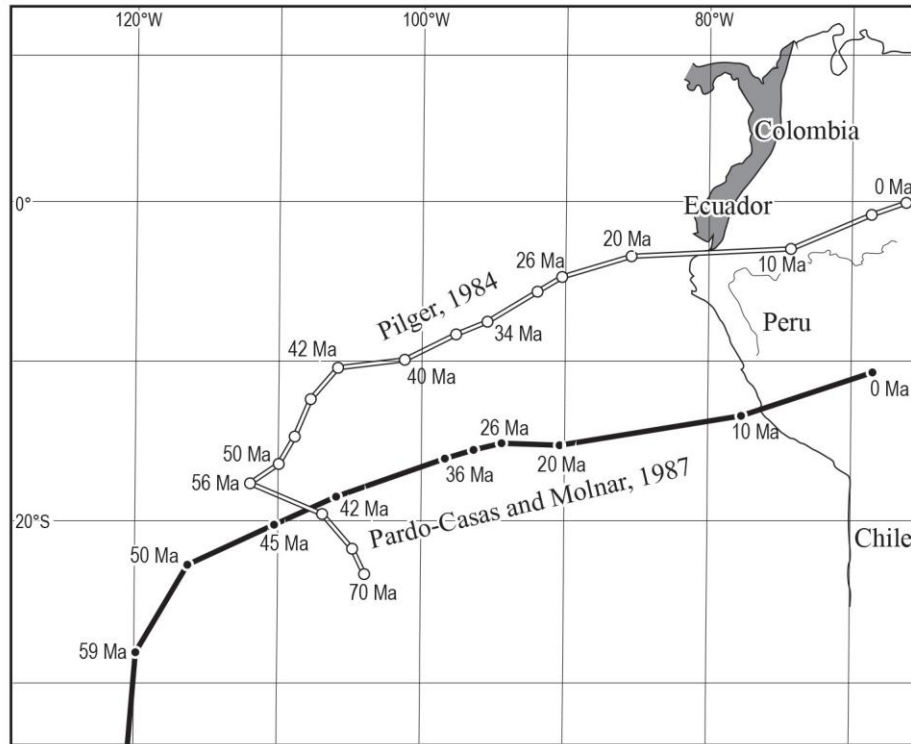


Fig. 7: Reconstructed convergence between the Farallón, then Nazca plate, and South American plates since the latest Cretaceous, after Pilger (1984) and Pardo-Casas and Molnar (1987).

In the central and northern Pacific, the drift direction of the Farallón plate is assumed to have changed to a more easterly direction about 80 Ma ago (middle Campanian) (Debiche et al. 1987; Atwater, 1989; Seton et al., 2012) (Fig. 7, 8), and the Apparent Polar Wander Path of the Pacific plate suggests that the oceanic plate shifted significantly and rapidly toward the East between 85 and 82 Ma (Sager and Pringle, 1988). At the same time, the drift of Africa changed from ESE- to E-ward (Emery and Uchupi, 1984; Steinberger and Torsvik, 2008, Jolivet et al., 2021), suggesting that a global plate reorganisation occurred at that time.

Although poorly dated, a significant change in the convergence direction is recorded in the latest Campanian (≈ 74 Ma) (Lithgow-Bertelloni and Richards, 1998), or in late Maastrichtian times (67 Ma) (Seton et al., 2012), which is not confirmed by Wright et al. (2016).

During the late Paleocene (≈ 60 Ma), the motion of the oceanic Farallón plate drastically changed from N (or even NW (Pilger, 1984)) to NNE or NE (Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Seton et al., 2012; Müller et al., 2019) (Fig. 7, 8).

In the middle Eocene (≈ 45 -40 Ma), the drift direction of the Farallón plate turned again from NNE or NE, to NE or ENE (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987), and its convergence rate increased (Pilger, 1984; Mayes et al., 1990). This is followed by a change in the convergence direction and a slowdown of the convergence rate in the late middle Eocene (≈ 40 -37 Ma) (Pilger, 1984; Pardo-Casas and Molnar, 1987; Mayes et al., 1990).

As the Farallón plate split into the Nazca and Cocos plates around 25 Ma ago (latest Oligocene), the convergence rate of the Nazca plate increased (Somoza and Ghidella, 2012) and its convergence direction became nearly E-W (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Tebbens and Cande, 1997; Seton et al., 2012; Wright et al., 2016) (Fig. 7, 8). From then on, the convergence rate accelerated since ≈ 20 Ma, but convergence direction remained more or less constant, and only slightly turned from E-ward to ENE-ward about 10 Ma ago.

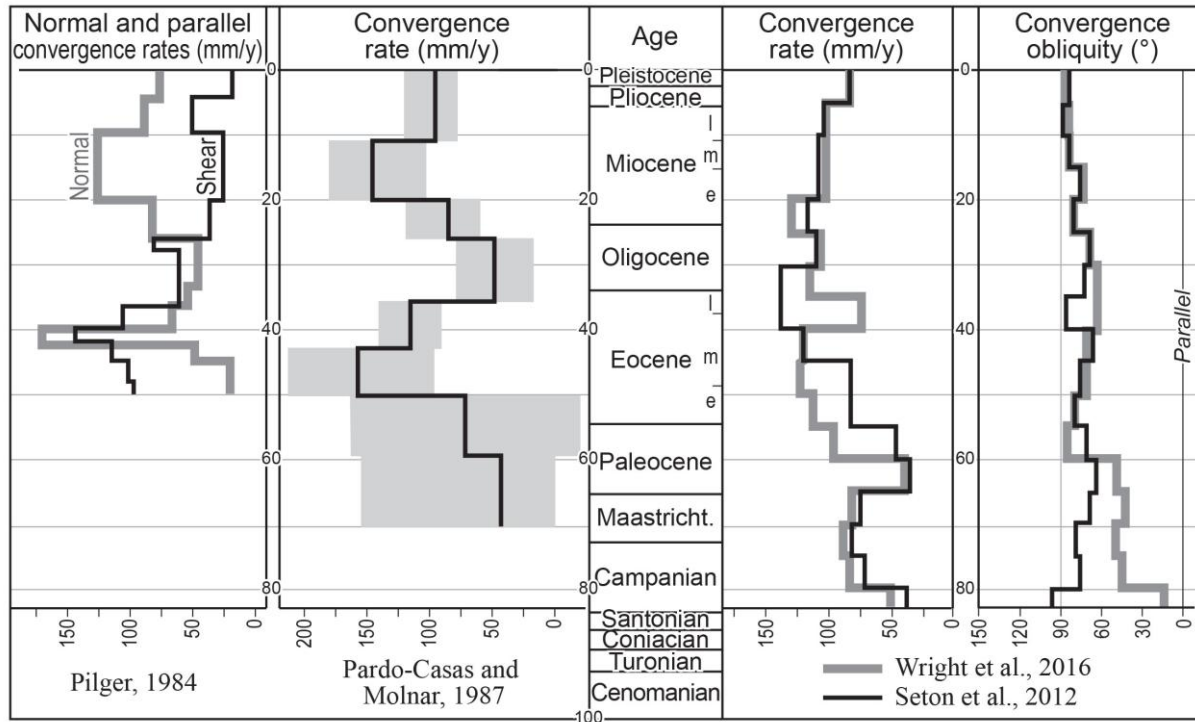


Fig. 8: Convergence rate and obliquity between the Farallón, then Nazca plate, and South America, according to Pilger (1984), and Pardo-Casas and Molnar (1987). Wright et al. (2016) proposed various scenarios according to the adopted time-scale.

Some of these major changes in the convergence direction roughly coincide with significant tectonic phases recorded in northern Peru and Ecuador, in middle Campanian (≈ 80 Ma), late Paleocene (≈ 60 Ma), late middle Eocene (≈ 40 Ma) and latest Oligocene-early Miocene times (≈ 26 Ma).

5. Consequences of changes in the convergence direction

The geometry of a subduction zone is frequently defined by the angle α (slab dip) that the oceanic slab acquires before entering the subduction zone (Fig. 9). However, the distance the slab has to cover before to acquire this angle varies according to the obliquity β of the convergence between the two involved plates. The length of the bend $D\beta$ is much longer in case of oblique convergence than in case of orthogonal convergence (D_0 , Fig. 9). Thus, for a same drift velocity, the more oblique the convergence, the lower the deformation rate of the oceanic slab. Therefore, if the velocities of the oceanic and continental plates remain constant, the change from an oblique to a more orthogonal convergence direction will produce a significant increase of the deformation rate of the subduction slab. I propose that this phenomenon has significant geological consequences.

5.1. Change in the deformation rate

It has been long outlined that changes in the convergence direction along active margins induce variations in the orthogonal convergence vector and in strain partitioning (e.g. Jarrard, 1986; Dewey and Lamb, 1992; Platt, 1993; Dominguez et al., 1998; Chemenda et al., 2000). Because D_β varies with β , the change of the convergence direction is also associated with a change in the bending rate (i.e. the deformation rate) of the subducting plate during its entry into the subduction zone (Fig. 9).

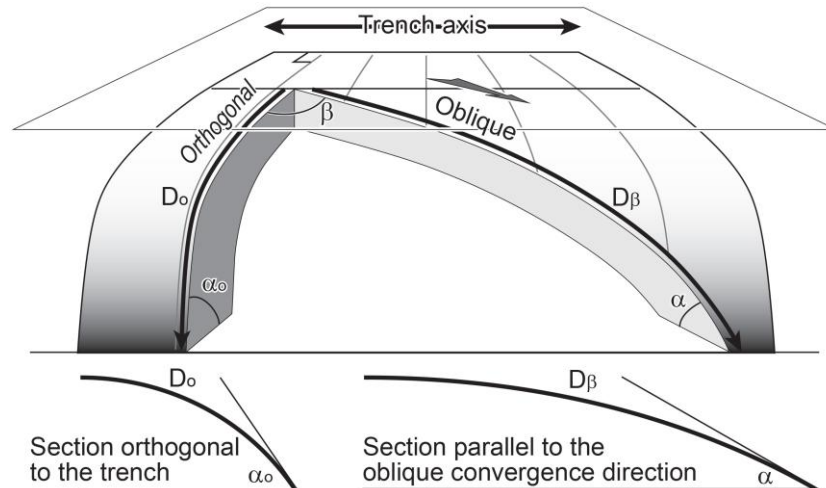


Fig. 9: Sketch depicting the oblique convergence in a subduction zone.

Geometry of the subduction slab being defined by its dip α_0 (measured orthogonally to the trench). An obliquely subducting oceanic slab has to deform more slowly than in case of an orthogonal subduction ($D_\alpha > D_0$). Therefore, for a same absolute drift velocity, a change from oblique to normal subduction implies a higher deformation rate of the oceanic slab.

Let consider the distance D_β that an arbitrary point of the subducting plate has to cover before to acquire the dip α of the subduction zone (Fig. 9). D_β is the portion of an arc ellipse, the length of which is a function of $\text{tg}\beta$. Therefore, for a same decrease of the angle β , the variation of D_β is much higher if the initial angle β is high. As a consequence, the more oblique the convergence, the stronger the acceleration of the deformation rate in case of change toward a more orthogonal convergence. For the same reason, this acceleration is low, or may even be negligible, for a change from a moderately oblique ($\beta \leq 30^\circ$) to a normal convergence.

5.2. Expected geological consequences

Oceanic lithosphere may be considered as an elastic body, but a viscoelastic model is more consistent with geological observations (e.g., Chemenda, 1994). Although detailed mechanical study is behind the scope of this paper, it is commonly admitted that for viscoelastic bodies, an increase of the deformation rate induces an increase of the elastic thickness, so that they will deform less easily and may fail at higher strain than at lower deformation rate (e.g., Park, 1989; Keary and Vine, 1990; reviews in Zhang and Zhao, 2014; Yin et al., 2021). Additionally, oceanic lithosphere is significantly stronger than continental lithosphere, because of its lithological composition and behaviour with depth (e.g. Molnar, 1988; Jackson, 2002; Burov, 2011). Furthermore, oceanic lithosphere usually does not bear inherited structures, except transform faults, while inherited structures of a continental plate may introduce zones of weakness. As a consequence, since they increase the strain rate, changes toward a more normal convergence direction, will induce a higher strength of the

oceanic slab, which will be reluctant to bend to enter the subduction zone and the increased coupling will more probably result in the deformation of the upper continental plate.

In case of change from a very oblique to a less oblique convergence direction ($\beta_2 < \beta_1$), the subducting slab, which had initially to bend until a low angle α_1 (Fig. 10, above), must either (1) undergo a more rapid bend to fit the geometric dip of the subduction zone determined by the new convergence direction (Fig. 10, case A), or (2) deform the upper plate in order to minimize its proper deformation, so that the new slab dip will remain close to its initial bend, with $\alpha_2 < \alpha_0$ (Fig. 10, case B). In the first case, it can be expected that the upper plate will be uplifted and compressively deformed, due to the increased coupling along the interface between the two plates (Fig. 10, case A). In the second case, the upper plate will be tectonically eroded, in order to allow the oceanic plate to bend according to an angle α_2 , close to the initial slab dip α_1 (Fig. 10, case B). Since oceanic crust is mechanically stronger than the continental one (Burov, 2011), it can be expected that the second case is more frequent.

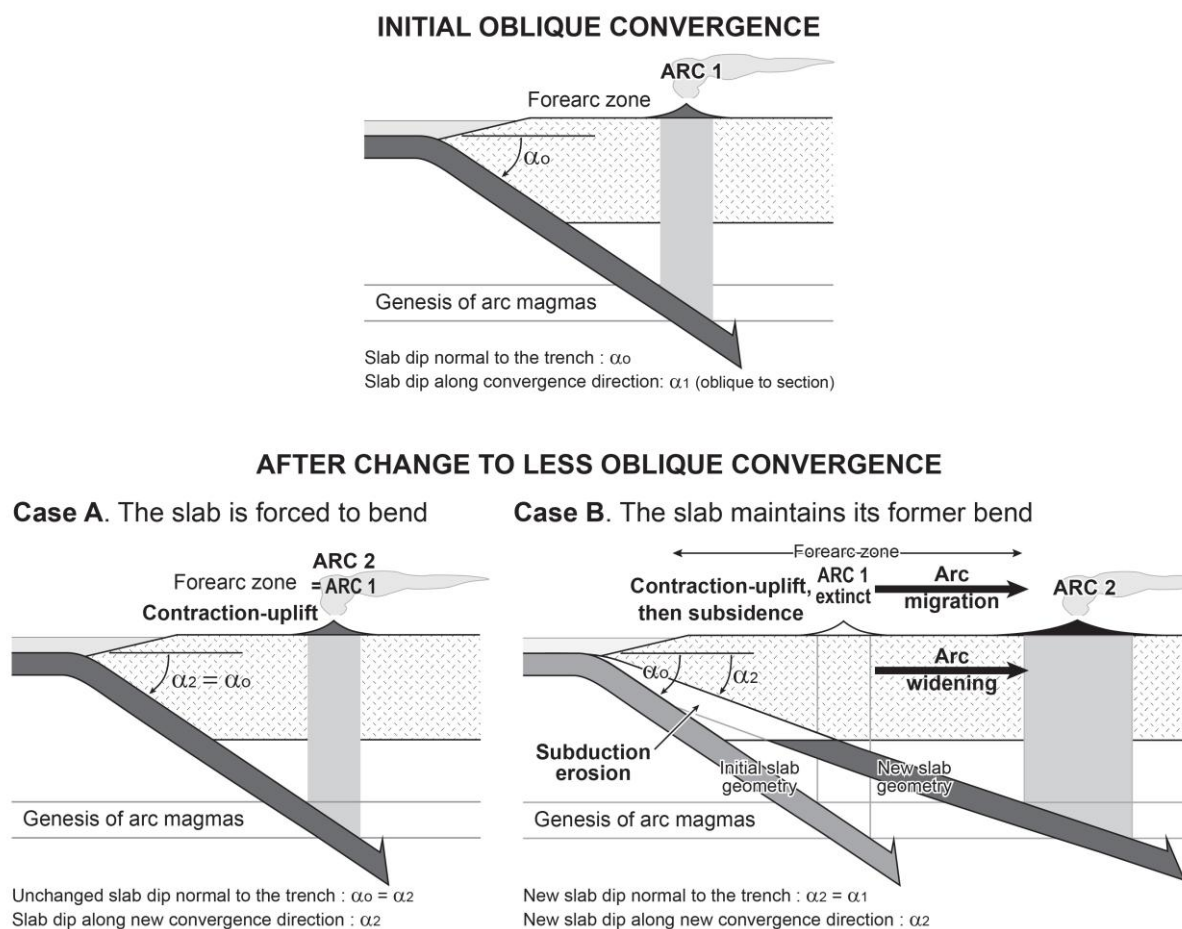


Fig. 10: Sketches showing the geological consequences of a change to a less oblique ocean-continent convergence.

Above: Initial oblique convergence. The subduction zone dips with an angle α_0 , but the bend of the oblique oceanic slab along its displacement vector is α_1 , with $\alpha_1 < \alpha_0$.

Below: After convergence became orthogonal.

A. The oceanic slab is forced to bend according to the initial subduction geometry. Its final bend α_2 is close to α_0 , but contractional deformation and uplift are expected in the arc and forearc zones.

B. The oceanic slab resists to the increase of deformation rate. Its final bend α_2 is close to α_1 , thus triggering subduction erosion of the edge of the upper plate. The latter provokes the subsidence of the forearc zone, and the landward shift and widening of the magmatic arc.

In real cases, it is likely that both processes occur, and that contractional deformation and uplift of the forearc zones (Fig. 10, case A) are associated with subduction erosion, in turn responsible for subsidence of the forearc zones (Von Huene and Scholl, 1991) (Fig. 10, case B). Subduction erosion of the upper plate will allow the new subduction zone to dip more gently than the former one ($\alpha_2 < \alpha_0$). As the depth of arc magma generation (≈ 80 -100 km) will not change, this would induce the landward migration of the magmatic front and the widening of the magmatic arc, since the down-going slab will intersect the depth zone of arc magma generation along a wider length (Fig. 10, case B). Moreover, flattening of the subduction zone is likely to increase the length of the contact zone submitted to coupling. This, together with the greater normal component of convergence, would increase the compressional stress in the upper plate, and therefore, its compressional deformation (e.g. Horton, 2018).

Conversely, a progressive change of the convergence direction in a more oblique direction is expected to induce an extensional regime in the forearc zones, provoking the subsidence of the latter, the steepening of the subduction zone and maybe the roll-back of the slab. Such examples, however, are not known in the Andes for the studied period.

6. Discussion

The Late Cretaceous-Oligocene period (≈ 85 -25 Ma) is marked (1) in the forearc zones, by the alternation of compressional deformation and subsidence pulses creating belts of forearc basins (Fig. 4, 5), (2) by the eastward shift (≈ 75 , 35 Ma) and widening (≈ 55 , 35 Ma) of the magmatic arc (Fig. 2, 4), and (3) by the eastward shift of compressional deformation, triggering the progressive emergence of the backarc areas and deposition of widespread clastic sheets (Fig. 2, 4).

The coincidence between compressional deformation in the Andean margin and changes in the convergence rate or direction has been already outlined (Jaillard and Soler, 1996). These apparently independent phenomena may be related to the change to a less oblique convergence direction of the oceanic plate with respect to South America, which occurred between 85 and 25 Ma.

The creation of at least one forearc basin (late Celica-Lancones-Yunguilla Basin) and the beginning of landward arc migration on the Peruvian margin around 75-80 Ma are coeval with significant, although poorly constrained, changes of the convergence (Seton et al., 2012; Wright et al., 2016). From then on, deformation of the forearc zone is marked by the alternation of (a) periods of compressional deformation and uplift (75-80, 68, 58, 40, 26-23 Ma), grossly coeval with changes in the convergence direction, and (b) periods of creation of subsiding extensional forearc basins (≈ 78 , 68, 45, 25 Ma). In the present model, stage (a) would be due to the change in the kinematic regime related to change of convergence direction, leading to contraction, uplift and subduction erosion as explained above, whereas stage (b) would correlate with periods of new geometric and kinematic equilibria, and release of the compressional stress, allowing extensional subsidence of the forearc zones subsequent to subduction erosion (Jaillard and Soler, 1996). This suggests that subduction erosion is favoured during periods of change to a more orthogonal convergence direction (Herbozo et al., 2020).

Note that in Ecuador, accretionary events by subduction jam of buoyant oceanic terranes occurred around 80-75 Ma, 68 Ma and 56 Ma (Spikings et al., 2010; 2024; Vallejo et al., 2019; Jaillard, 2022). Since convergence was almost parallel to the Ecuadorian margin before 80 Ma (Fig. 7, 8), change in the convergence direction at 80-75 Ma may have pushed the oceanic material, close to the margin at that time, into the subduction zone, thus provoking its

jam (see Feininger and Bristow, 1980; Spikings et al., 2001; 2024; Jaillard et al., 2009). Since oblique convergence continued, the oceanic material may have been split into strips, each one still migrating N-ward, parallel to the Ecuadorian margin. Subsequent changes in the convergence direction may have provoked the introduction of strips of oceanic material beneath the already accreted terranes, or the further sinking of already underplated material, thus building and thickening a crustal-scale accretionary wedge (Jaillard, 2022; 2024). This would explain why contractional deformation recorded both in the Peruvian and Ecuadorian margins are grossly coeval, although accretion is only known in the latter. As a matter of fact, the consequences of changes in the convergence direction as exposed above would have comparable consequences on the upper plate, whatever the material involved. If so, deformation phases in Ecuador would be primarily caused by changes of the convergence to a more orthogonal convergence direction, rather than by the accretion of oceanic material.

From Eocene (≈ 50 -40 Ma) onwards, the area affected by compressional deformation widened and migrated eastward, the maximum deformation affecting thereafter the arc and western part of the backarc zones. Flattening of the subduction zone due to subduction erosion increased the surface submitted to interplate coupling (Fig. 10, case B). This, together with the increased normal convergence component, would be responsible for the increase of the interplate coupling and, therefore, for the eastward migration of the stress applied to the incipient Andes. From 25 Ma onwards, no significant changes are recorded in the convergence direction, which remained roughly orthogonal to the Andean margin, and the Andean margin was submitted to frequent, repeated compressional phases that may appear as continuous (Sempere, 1991; Marocco et al., 1995; Noblet et al., 1995).

The drastic widening of the magmatic arc in the Eocene (≈ 50 -40 Ma) has been interpreted as the result of the flattening of the subduction zone (Soler and Bonhomme, 1990; see also Scheuber et al., 1994; Capaldi et al., 2021). Density variation of the subducting slab may explain part of this change, since the relative age of the subducting oceanic slab slightly diminished after 40 or 30 Ma (Soler and Bonhomme, 1990; Wright et al., 2016). Nevertheless, the inferred progressive flattening of the subduction zone may be also a geometric consequence of the subduction erosion of the upper plate, subsequent to the change of convergence direction (Fig. 10, case B), and one cannot exclude that both processes may have acted together.

Several authors proposed that the more rapid the ocean-continent convergence, the higher the intensity of upper plate deformation (e.g., Pardo-Casas and Molnar, 1987; Wolf and Huisman, 2019, and references therein). The increase of convergence rate from latest Cretaceous to Eocene times (e.g., Soler and Bonhomme, 1990; Seton et al., 2012; Wright et al., 2016; Fig. 8) actually correlates with the onset of contractional deformation in the Andes of northern Peru and Ecuador. However, the period of highest contraction and uplift of the Andes (Late Oligocene-Miocene) is coeval with a continuous decrease of convergence velocity (Soler and Bonhomme, 1990; Seton et al., 2012; Wright et al., 2016; Fig. 8). Therefore, even though convergence absolute velocity may play a role in the tectonic regime of the upper plate, the role of changes on the convergence conditions seem to be more important regarding the deformation of the continental plate (e.g., Jaillard and Soler, 1996; Jaillard et al., 2000; Horton, 2018).

Finally, an abrupt increase of the convergence rate should have the same consequences as the change of convergence direction to a more orthogonal direction, since the latter must also provoke a sharp increase of the deformation rate of the subducting slab. Actually, a change of convergence direction from oblique to normal mechanically provokes an increase of the normal component of convergence. The latter process may also act independently; for instance, the significant increase of normal convergence rate recorded in middle to late

Eocene and Miocene times (Fig. 8) is probably responsible for the contractional deformation and exhumation of the Andean domain, since no significant change in the convergence direction is recorded during these periods (Fig. 8). However, tectonic deformation occurring during periods of decrease in the convergence rate (late Campanian?, Paleocene, late Miocene) remains to be explained.

7. Summary and conclusions

During Late Cretaceous-Paleogene times (≈ 80 -25 Ma), the convergence between the motion of the Farallón relative to the South American plates was marked by a step by step evolution from an oblique to an orthogonal convergence direction. Combined with other subduction processes, this phenomenon may have contributed to the contractional tectonic regime that built the Andes of Ecuador and northern Peru.

Change from a very oblique subduction system to a more orthogonal direction implies a significant increase of the deformation rate of the oceanic slab. Since an acceleration of its deformation rate makes more difficult the deformation of the slab, this change in the kinematic conditions would provoke either the forced bending of the subducting plate, or the subduction erosion of the upper plate, in order to minimize the deformation of the subducting slab. Because oceanic lithosphere is mechanically stronger than the continental lithosphere, the second case is more likely to occur, and would induce the flattening of the subduction zone. This model is consistent with major features of the Late Cretaceous-Paleogene evolution of the Peruvian-Ecuadorian margin.

The forearc zones are marked by an alternation of (a) compressional deformation, which roughly coincide with periods of change of oceanic slab drift to an orthogonal direction, and (b) periods of forearc subsidence, likely related to subduction erosion, which would correspond to periods of releasing stress and steady-state convergence direction.

The eastward shift and widening of the magmatic arc observed between ≈ 75 and 40 Ma would result from the progressive flattening of the subduction zone, subsequent to subduction erosion, which would have enlarged the zone of magma production. The flattening of the subducting slab, together with the increased normal component of convergence, necessarily increased the surface submitted to inter-plate coupling. This may have contributed to the increase of contractional deformation and to the backarc-ward migration of the areas submitted to contractional deformations, from the Miocene onwards.

This model would also explain why the progressive change toward an orthogonal ocean-continent convergence between 80 and 25 Ma originated well defined discrete contractional phases, whereas the Neogene evolution of the Andes (≈ 25 -0 Ma), marked by a nearly constant convergent direction is sometimes interpreted as a compressional "continuum".

Additionally, this process may explain why Andean margins with, or without accretion, recorded coeval deformation phases. In fact, changes in the convergence direction may explain both the underplating beneath the Andean margin of buoyant oceanic material located close to the trench, and the correlative deformation of the continental plate.

Finally, although pre-Late Cretaceous convergence directions are poorly constrained, this interpretation might also explain why the Andean margin behaved as an extension-dominated margin during Jurassic-Early Cretaceous times, recorded the first compressional deformation in Albian times, when South America began to drift west-ward, accelerating convergence rate (Frutos, 1981), and initiated the contractional building of the Andes in Late Cretaceous times, when the ocean-continent convergence began to evolve to an orthogonal direction.

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