New insights into Andean evolution: An introduction to contributions from the 6th ISAG symposium (Barcelona, 2005)

1. Introduction

The 6th International Symposium on Andean Geodynamics (ISAG) was held in Barcelona on September 12–14, 2005. Like previous editions, it provided an opportunity to some 250 scientists from most fields of the Earth Sciences to expose and discuss recent progresses in the understanding of the Andes. As much as 218 communications were presented (abstracts can be downloaded from http://irdal.ird.fr/article.php3?id_article=1575), manifesting that the interest of the geoscientific community for this outstanding orogen is not only high, but also growing. This interest has been furthermore illustrated by the recent publication of three books, respectively edited by Kay and Ramos (2006), Oncken et al. (2007), and Moreno and Gibbons (2007), that bring together many noteworthy contributions.

The prefaces to the *Tectonophysics* special issues published as sequels to the Oxford 1993 ISAG (Dewey and Lamb, 1996), Göttingen 1999 ISAG (Jaillard et al., 2002), and Toulouse 2002 ISAG (Gerbault and Hérail, 2005), reviewed the concepts, models and issues prevailing about the Andes. In the last years some views have started to evolve under the growing perception and concern that current models do not always reflect reality satisfactorily, especially in the Central Andes (e.g., Sempere, 2000; Wörner and Seyfried, 2001; Sempere et al., 2004; Hartley, 2005; Garzione et al., 2006; Sempere and Jacay, 2006, 2007). This situation somehow makes that some shifts in the dominant paradigm, which for this region is best illustrated by Isacks's (1988) landmark paper, have to be expected in the forthcoming years. We therefore take the opportunity to briefly review here some selected issues that are emerging—cautioning that topics dealt with and references mentioned below are by no way exhaustive, much more being available in the papers and books cited herein.

2. Gradients and segmentation along the Andes

The Andes have long been viewed as a prime example of an orogen produced by subduction of an oceanic plate beneath a continental margin. However, the tectonic and magmatic evolution of this outstanding mountain belt is still rather far from being satisfactorily understood. In particular, it has become increasingly evident that the Andes actually consist of contrasted segments that have been shaped by specific sets of processes and have therefore undergone dissimilar evolutions (e.g., Jordan et al., 1983; Kley et al., 1999). The reasons for these differences are only partially identified at the present time.

The first-order genetic distinction separates the Northern Andes, which basically extend north of the Gulf of Guayaquil, from the rest of the Cordillera (Fig. 1). Unlike the Central and Southern Andes, which were built along the active continental margin of western South America since at least Early Mesozoic times, the evolution of the Northern Andes has been dominated since the Late Cretaceous by the major, protracted, collisional to transpressional deformation produced by the eastward motion of the Caribbean–Colombian Oceanic Plateau (CCOP) against northwestern South America. The evolution of the Central and Southern Andes, in contrast, has apparently developed in a “pure” ocean–continent subduction setting, i.e. without collisional processes involving either another continental mass or an oceanic plateau. Although some authors described the Central Andean mountain build-up as the result of an “ocean–continent collision” (e.g., Russo and Silver, 1996; Yuan et al., 2000), this concept is arguable and debated by Andean geoscientists, in particular because the crustal thickness and buoyancy of the downgoing oceanic plate is significantly lower than that of a continental mass or oceanic plateau.

An intriguing and long-known paradox is that the crustal thickness in the Central Andes, in spite of an orogenic evolution devoid of collisions, reaches values similar to those in the Himalayas–Tibet, which were indeed achieved through continent–continent collision. This apparent similarity in maximum crustal thickness has affected for several decades our scientific thinking about the processes that result in crustal thickening, tending to emphasize observations and interpretations common to the Andes and Himalayas (see examples in Gerbault and Hérail, 2005), and thus to envisage the Andes as another, albeit particular, case of “collisional” orogen. A further apparent paradox, however, may be discerned in the fact that the orogenic volume produced by ~80 Myr-long collision-related tectonics in the Northern Andes is intriguingly much lower than that in the Central Andes although these were not affected by true collisional processes (Fig. 1). Furthermore, the Andean orogenic volume exhibits significant longitudinal gradients, primarily opposing the Central Andes, where orogenic volume is highest, and the Southern Andes, where it is considerably lower, although these segments evolved in a similar subduction context and grade into each other (Fig. 1). The Central Andes are in turn segmented by other gradients (see below).

The papers included in this Special Issue relate to the Northern Andes (3 papers) and Central Andes (10 papers), and we detail hereafter specific aspects relative to these two segments.

3. Collisional interaction of an oceanic plateau along a continental margin: the Northern Andes

The Northern Andes have been built along and “around” the northwestern margin of South America. Their complex structure have been dealt with by many studies and shown to be related to the interaction of the Caribbean plate with northwest South America (e.g., Pindell and Barrett, 1990; Erlich and Barrett, 1990; Freymueller et al., 1993; Meschede and Frisch, 1998; Lallemant and Sisson, 2005; Montes et al., 2005). Before orogeny the eastern and western regions of northwest South America had underwent contrasted evolutions, as the former first evolved as a passive margin created in the Jurassic by the separation of North and South America (in fact, West Gondwana), whereas the latter had functioned as an active margin maintained by the subduction of Panthalassan (Pacific) oceanic plates beneath the continent since at least the Paleozoic (e.g., Jaillard et al., 1990, 1995; Pindell et al., 2005, 2006).

The Caribbean–Colombian oceanic plateau (CCOP) was built on a Panthalassan oceanic plate starting ~91 Ma, presumably by activity of the Galapagos hot spot (e.g., Kerr et al., 2002, 2003; Kerr and Tarney, 2005; Luzieux et al., 2006). Driven by plate motions, collision of the CCOP with the northwestern margin of South America began to develop in the early Campanian (~85–80 Ma: Kerr et al., 2002; Jaillard et al., 2004; Spikings et al., 2005) or in the late Campanian (~75 Ma: Vallejo et al., 2006; ~73 Ma: Luzieux et al., 2006). Significant interchanges of terrestrial biota between North America and South America started in the Campanian, suggesting that collision had built by ~80–75 Ma a land bridge between the two continental masses (Gayet et al., 1992). Collision produced accretions of oceanic and arc terranes along the Ecuador–Colombian margin during the Late Cretaceous and Paleogene (McCourt et al., 1984; Reynaud et al., 1999; Spikings et al., 2001, 2005; Jaillard et al., 2004; Toro-Álava and Jaillard, 2005) and became more tangential as the CCOP moved along the northwest margin of South America. Intense and long-lasting tectonic interaction of the Caribbean plate (i.e., the deforming CCOP) along adjacent northwest South America has produced a variety of deformational features (e.g., Dengo and Covey, 1993; Taboada et al., 2000; Guillier et al., 2001; Gómez et al., 2005), including compressional to transpressional inversion of ancient extensional structures (e.g., Kellogg and Vega, 1995; Colletta et al., 1997; Acosta et al., 2004; Mora et al., 2006). The resulting orogenic units, however, exhibit markedly lower volumes and average elevation than the Central Andes.

In this issue Schmitz et al. document the variations in crustal thickness in Venezuela on the basis of deep seismic observations, and thus provide a major advance in our knowledge of the Venezuelan segment of the Northern Andes. They show that crustal thickness overall steadily decreases from ~45 km in the vicinity of the Guyana Shield to ~35 km on the coast, as expected from the regional geological history. In addition Schmitz et al.'s (2008-this issue) study highlights the existence of two anomalous regions: the eastern part of the Eastern Venezuela Basin is characterized by a crustal thickness up to 50 km as well as velocity anomalies in the local lower crust, both resulting from the tectonic interaction of the Caribbean plate with the continent; in contrast, the crustal thickness beneath the Falcón Basin of western Venezuela remarkably thins from 35 to 27 km, and this anomaly extends eastwards into the Bonaire Basin.

Jácome et al. (2008-this issue) complement Schmitz et al.'s study by providing an integrated seismic, flexural and gravimetric modelling of the north-central region of Venezuela (Coastal Cordillera thrust belt and Guárico Basin). They conclude that loading of the South American lithosphere by the Coastal Cordillera generated the subsidence observed in the Guárico Basin. Shortening in the Coastal Cordillera decreases from ~44 km in the west to ~10 km in the east, whereas the Moho is >35 km-deep there and shallows to ~30 km near the Caribbean Sea.

The Mérida Andes of western Venezuela are primarily a compressional to transpressional basement uplift, the altitude of which is locally >5000 m. This mountain belt has grown since the Neogene in response to the ongoing displacement of the Maracaibo block relatively to the South American craton, which in turn is produced by the westward motion of the Caribbean plate (Hervouët et al., 2001, 2005; Audemard and Audemard, 2002; Audemard, 2003; Chacia et al., 2005; Backé et al., 2006; Duerto et al., 2006). The NE-striking Boconó fault is a major wrench structure that occurs in the axis of the Mérida Andes, and its new trench investigation by Audemard et al. (2008-this issue; Fig. 1) complements previous sismological and paleosismological studies in Venezuela (Audemard et al., 1999, 2005; Audemard, 2005). In particular it confirms the Holocene activity of the Boconó fault at the Apartaderos pull-apart basin, which is bounded by a northern strand of the fault where earthquakes recur every 1200–1350 yr and a southern strand where they occur every 400–450 yr. An older activity is documented by slumping, rotational sliding, and numerous earthquake-triggered liquefaction features.

Fig. 1. Location of study areas of the papers included in this Special Issue. The contrasts in topography along the orogen are believed to reflect dissimilar geological evolutions driven by different combinations of tectonic and magmatic processes of varied ages and intensities.

4. Collision-free evolution of a major continental arc: the Central Andes

4.1. Continental arcs

Continental arcs are mountain belts (orogens) typically produced by the subduction of an oceanic plate beneath a continental margin. Their build-up evidently involves tectonic *and* magmatic processes, but the detailed ranges, interactions and consequences of these two categories of processes remain a matter of debate. Continental arcs being one particular class of orogens, a short comparative overview of orogenic types and processes may be useful to highlight the main issues typical of the Central Andes.

The characteristic high topography of continental mountains and plateaux may result from crustal thickening, which isostatically increases surface altitudes, and/or from mantle upwelling, which dynamically uplifts the crust (but does not thicken it). Because most orogens present a thickened crust, a major current issue in geology concerns the processes involved in crustal thickening. In the case of the Andes, high altitudes are believed to be a primary consequence of a thickened crust, but it has been proposed that lithospheric thinning (by delamination) and related asthenospheric upwelling have played a significant role in uplift (Isacks, 1988; Kay and Kay, 1993; Kay et al., 1994; McQuarrie et al., 2005; Garzione et al., 2006; Molnar and Garzione, 2007).

The classic type of crustal thickening is illustrated by collisional orogens, which are produced during continental collision through the superposition (= “continental subduction”) and/or imbrication of two crustal masses. Crustal growth and thickening, however, is also known to develop by magmatic addition above ocean–ocean and ocean–continent subduction zones, resulting in arc orogens (oceanic and continental arcs, respectively; Tatsumi and Eggins, 1995; Dimalanta et al., 2002; Turner et al., 2003; Busby, 2004; Tatsumi, 2005; Tatsumi and Stern, 2006; Kodaira et al., 2007; Lee et al., 2007).

Because many major orogens (such as the Himalayas, Alps, Appalachians, Caledonides, Pyrenees, New Zealand Southern Alps, etc.) have resulted from frontal (Himalayan-type orogens) to tangential (New Zealand-type orogens) continental collisions and related tectonic shortening, the currently prevailing view is that collision-like processes can be generalized to arc orogens. Thus, arc orogens where some tectonic shortening is observed are sometimes described as resulting from “ocean–continent collision” (e.g., Russo and Silver, 1996; Yuan et al., 2000, in the case of the Central Andes).

However, magmatism is abundant in arc orogens but only minor during continental collisions, implying that arc and collisional orogens should be formally distinguished on this basis, as they are in terms of metamorphic processes (Ernst, 2005). In the early years of the plate tectonics paradigm it was proposed that arc orogens are primarily formed through subduction-related magmatic accretion, and since then this idea has been abundantly confirmed in island arc contexts (e.g., Busby, 2004; Tatsumi and Stern, 2006). Crustal growth by magmatic accretion has also been illustrated at continental arcs (e.g., Lee et al., 2007). Although this type of interpretation was initially applied to the Central Andes (e.g., James, 1971a,b; Thorpe et al., 1981), it has found limited support since the mid-1980s, mainly because most authors have concentrated on the existence of tectonic shortening in this orogen (see below). However, the hypothesis of Andean crustal growth by magmatic accretion has not been convincingly discarded and does remain an interpretative option as well as a matter of study (e.g., Kono et al., 1989; Rogers and Hawkesworth, 1989; Sandeman et al., 1995; James and Sacks, 1999; Haschke et al., 2002a; Haschke and Günther, 2003). Furthermore, the deformational processes at depth are known to be decoupled from those observed in the upper crust, the latter passively following the former (England and Molnar, 1991). In this light, the possible role of ductile lower crustal flow in modulating Andean crustal thickness has attracted some attention since a few years (e.g., Husson and Sempere, 2003; Yang et al., 2003). The genetic relationships between subduction and non-collisional crustal thickening thus remain imprecise and controverted; more work, and in particular more interdisciplinary integration, are needed to advance this issue.

Arc orogens present two interesting and complementary aspects. First, they provide a framework to estimate how and how much extraction and transfer of material from the mantle, as demonstrated by geochemical studies (e.g., Faure, 2001), participate in thickening the crust above subduction zones (e.g., Reymer and Schubert, 1984; Tatsumi and Stern, 2006), and what is the bearing of this mass transfer on the evolution of the orogen. Second, because Himalayan-type collisions are also driven by subduction of an oceanic plate, a continental arc has generally developed before collision along the margin of one of the colliding continents, resulting in distinct crustal evolutions and structures of the two continents. Thus a Himalayan-type orogen is generally preceded by an Andean-type orogen along the overriding margin of a subduction-driven system that ultimately results in collision.

A puzzling character of active continental arcs is that they present a variety of morphological, tectonic, and magmatic features, most notably along the rim of the Pacific Ocean. The Central Andes, in particular, profoundly contrast with all other Neogene ocean–continent subduction orogens on Earth in that their crustal thickness reaches values comparable to those in the Himalayas–Tibet. The evident and pronounced dissimilarity between subduction orogens east and west of the Pacific Ocean has been related to large-scale mantle dynamics and flow (e.g., Doglioni et al., 1999, 2007) and subducting slab width (Schellart et al., 2007). This intriguing variety of subduction orogens worldwide makes more crucial the need to better understand what are the processes and parameters that control crustal growth and deformation in continental arcs and, in particular, the Central Andes.

4.2. Anatomy of the Central Andes

As underlined above, the Andes show considerable longitudinal variations. Its largest and most voluminous segment is formed by the Central Andes, which transitionally grade to the south into the Southern Andes, and more rapidly to the north into the Northern Andes (Fig. 1). It is widely agreed that the Central Andes provide an extreme recent case of arc orogen, as it is recognized that they have only been built by tectonic and magmatic processes produced by the subduction of an oceanic plate beneath western South America. The Central Andes are characterized by the dominance of protracted magmatic activity in the west (widely distributed around the Western Cordillera, i.e. the arc) and of tectonic shortening in the east (in the Eastern Cordillera and adjacent areas), with the foreland extending east of the latter, and the forearc west of the former. As a whole, the Central Andes form a major segment of continental arc and display extraordinary characteristics and noteworthy internal longitudinal gradients.

Variations in orogenic volume make that the ~4000 km-long Central Andes are in turn segmented into the northern Central Andes (NCA, 5°30'S–~13°S; entirely located in Peru), the Central Andean Orocline (CAO, ~13°S–28°S; over southern Peru, Bolivia, northern Chile, northwestern Argentina), and the southern Central Andes (SCA, 28°S–37°S; over central Chile and west-central Argentina). The CAO covers an area of ~1,300,000 km^2 and its orogenic volume is by far the largest of the entire Andes. Orogenic volume strikingly decreases north and south of the CAO (Fig. 1). The transition between the CAO and NCA is formed by the Abancay deflection, a peculiar sub-segment where the Andean structural strike exhibits a significant rotation (Roperch et al., 2006).

The Central Andean Orocline was initially named “Bolivian Orocline” due to its oceanward concavity and geographic location, but it largely extends outside Bolivia and is best characterized by its considerable crustal thickness. In some areas the width of the CAO, between the subduction trench and the sub-Andean front, is >850 km, and its crustal thickness is >70 km (Lyon-Caen et al., 1985; Kono et al., 1989; Beck et al., 1996; Schmitz et al., 1999; Yuan et al., 2000, 2002). The origin and apparent persistence of such a crustal thickness in the CAO are intriguing given that the orogen does not result from continent–continent collision and displays significant tectonic shortening mainly along its eastern half. The CAO thus provides the most extreme Neogene case of crustal thickening among the varied and contrasted segments of arc orogens known along the Pacific Ocean margins.

4.3. The Central Andes under the weight of a paradigm

Scientific activity and production take place under the light of paradigms (Kuhn, 1962) and, evidently, the geosciences make no exception. Paradigms orient research at all scales and are notoriously long-lived. Plate tectonics—the paradigm that currently governs our large-scale understanding of the Earth—was formalized and accepted by a majority of geoscientists, more than 50 years after Alfred Wegener's (1915) seminal observations and interpretations, because it provided a powerful framework to understand and investigate how our planet has been and is physically evolving. In the case of the Central Andes, nearly all geoscientific studies conducted since the late 1980s have admitted, explicitly or not, that crustal thickening there has been primarily achieved through tectonic shortening of the South American margin, and that magmatic additions to the crust have been only minor (see references in Dewey and Lamb, 1996; Jaillard et al., 2002; Gerbault and Hérail, 2005). This dominant large-scale interpretative framework can be termed "the Isacksian paradigm" as it was first highlighted and formalized in Isacks's (1988) landmark paper.

Within this paradigm, a particular idea operating in the Central Andes, albeit less distinctly, consists in the belief that significant parts of the Andean tectonic history was imposed by subduction of specific features of the subducted plate. In this view prominent aspects of deformation of the South American margin were somewhat passive reactions to mechanical strain imposed on it by the subduction of large objects present on the subducted plate. In the last decade, for instance, a number of papers have favored the subduction of the Nazca Ridge as the cause for a variety of Central Andean geological features thought to be younger than 10 Ma (e.g., Hampel, 2002; Rousse et al., 2003; Rosenbaum et al., 2005; Espurt et al., 2007). Other features of the subducting plates have been invoked to explain other aspects of the Central Andean evolution (e.g., Soler et al., 1989; Yáñez et al., 2001; van Hunen et al., 2002; Sdrolias and Müller, 2006). The idea that subduction of large topographic and/or thermal peculiarities carried by the subducting plate must have imposed specific deformation along portions of the Andean margin is evidently acceptable, but the magnitude of the mechanical effects actually produced by such processes is not always easy to assess.

Scientists have long been searching for relations between features related to the oceanic plate on one hand (direction and velocity of convergence, slab dip), and orogenic volume on the other, in the Central Andes (e.g., Jordan et al., 1983; Reutter et al., 1994; Sandeman et al., 1995; Lamb et al., 1997; James and Sacks, 1999; Ramos and Aleman, 2000) and elsewhere. The effects of subduction, however, do not depend only on characteristics of the subducting plate, but also on the rheological properties of the mantle on each side of the slab (e.g., Russo and Silver, 1996; Heuret et al., 2007; Schellart et al., 2007), on the heterogeneities of the continental margin, and on the amount and evolution of the magmas generated in the mantle wedge.

Indeed, aside from the tectonic, i.e. mechanical, interaction of the converging plates, the other first-order characteristic feature of subduction zones is the production of a generally abundant arc magmatism along the margin of the overriding plate. Tectonic and magmatic processes should therefore be viewed as two related aspects of one same system. The idea that arc orogens are formed through magmatic accretion forced by subduction is widely admitted in island arc contexts (e.g., Tatsumi and Stern, 2006), but has only received minor attention in the case of the Central Andes (James, 1971a,b; Thorpe et al., 1981; Kono et al., 1989; Rogers and Hawkesworth, 1989; James and Sacks, 1999; Haschke et al., 2002a; Haschke and Günther, 2003)—a situation largely due to the adoption of the current paradigm in the late 1980s. It is a matter of fact that, since Isacks (1988), many researchers in the Central Andes have concentrated on tectonic shortening.

This orientation has led to a number of dissimilar, purportedly "balanced" crustal-scale cross-sections obtained by graphic construction (e.g., Roeder, 1988; Sheffels, 1990; Schmitz, 1994; Baby et al., 1997; Kley et al., 1999; McQuarrie, 2002). In the CAO shortening is indeed evident in the Eastern Cordillera and sub-Andean belt (e.g., Roeder, 1988; Sempere et al., 1990; Sheffels, 1990; Roeder and Chamberlain, 1995), but, at least in southern Peru, it is very limited or even absent in the western Altiplano, arc and forearc (James, 1971b; Myers, 1975; Kono et al., 1989; James and Sacks, 1999; Sempere and Jacay, 2006, 2007). The observed crustal thickness cannot be accounted for by the available tectonic shortening estimates, especially in the arc and forearc (Schmitz, 1994; Kley and Monaldi, 1998; Giese et al., 1999; Ramos and Aleman, 2000; Yuan et al., 2000). This lack of surface evidence for significant shortening in the western half of the Andes was accommodated in graphic constructions by supposing blind crustal duplexes (e.g., McQuarrie, 2002) or insertion, at the base of the crust, of crustal slices tectonically displaced from the western margin (e.g., Baby et al., 1997; an idea set forth by Rutland, 1971), but no evidence has been obtained yet for any of such large-scale and dramatic phenomena. The latter hypothesis is furthermore contradicted by the occurrence of Early Paleozoic arc rocks all along the coast of southern Peru (e.g., Mukasa and Henry, 1990; Loewy et al., 2004), and by the well-known limited migrations of the arc during Mesozoic and Cenozoic times.

The limited Andean-age shortening observed at the surface of the forearc, arc, and inner backarc of southern Peru provides a major counterexample against the assumption of an orogenic build-up mostly driven by "ocean–continent collision", and has led a handful of authors to propose that crustal thicknening in the Western Cordillera was essentially achieved by magmatic additions (e.g., James, 1971b; Kono et al., 1989; James and Sacks, 1999), representing a net crustal growth at the arc. This idea is supported by the fact that the isotopic characteristics of most Andean magmas unambiguously indicate that they largely consist of material extracted from the mantle (e.g., McNutt et al., 1975; Harmon et al., 1981; Boily et al., 1989, 1990; Soler and Rotach-Toulhoat, 1990; Parada et al., 1999; Faure, 2001; Kelemen et al., 2004). Furthermore, I-type magmatism, a typical feature of Andean arc batholiths (Pitcher et al., 1985), is now understood to result from the reworking of crustal materials by mantle-derived magmas, and is even viewed to drive the coupled growth and differentiation of continental crust (Kemp et al., 2007). Crustal growth rates at arcs are now known to be at least 40–95 km^3/km Myr, i.e. twice the rates estimated by Reymer and Schubert (1984), who nevertheless mentioned a few cases with arc crustal growth rates as high as ~300 km^3/km Myr. Estimated volumes of volcanic rocks erupted at the surface were invoked to discard magmatic addition as a significant cause of crustal thickening (e.g., Francis and Hawkesworth, 1994), but updated estimates are much higher (de Silva and Gosnold, 2007); besides, no secure constraints are available on the ratio of volcanic volumes to total magmatic volumes, and this ratio might well be anomalously low in the case of thick crusts. Moreover, in this case, crustal thickening and surface uplift may be enhanced by further metamorphic processes that decrease the overall density of the lower crust (e.g., Le Pichon et al., 1997).

At least in the northwestern portion of the CAO, a number of geophysical and geological data and observations provide further counterexamples against the idea of crustal thickening by major horizontal shortening in the Central Andes. In central Peru at 11°–12°S, the eastern boundary of the Eastern Cordillera is a seismically active subvertical fault zone that cuts through the lithosphere down to at least 30 km depth (Dorbath et al., 1986). Seismic tomography also detects a subvertical lithospheric-scale boundary in the eastern Altiplano of Bolivia (Dorbath et al., 1993) and in its prolongation, i.e. along the southwestern edge of the Eastern Cordillera of southern Peru, the distribution of magmatic rocks (Sempere et al., 2004) and the isotopic geochemistry of mantle-derived rocks (Carlier et al., 2005) also map a subvertical lithospheric boundary, which coincides at the surface with a major fault system separating two contrasting orogenic domains (Sempere and Jacay, 2006, 2007).

In all fields of science, paradigms are known to deeply influence interpretations and even observations (Kuhn, 1962). The belief that the Central Andes originated by tectonic shortening has commonly biased cartography in this orogen, for instance by forcing high-angle or poorly-exposed faults to be mapped as reverse faults and thrusts. Eloquently enough, some same areas have been mapped in dramatically different ways by geologists who favored distinct models (see cases in Sempere, 2000; Wörner and Seyfried, 2001). Extensional structures have often been overlooked, because they were thought to be irrelevant in the investigation of Andean orogenic issues. However, observations and models from a variety of undoubtedly extensional settings in Europe and Africa have shown that some structural geometries previously thought to be typical of contractional processes, as in the Central Andes, in fact also occur in extensional contexts, in particular where normal faults were initiated as flexure-forming blind faults (e.g., Finch et al., 2004). At least in southwestern Peru, identification and correction of such biases result in major revisions of structural mapping, and the forearc, arc, and southwest Altiplano in fact appear to have been dominated by transcurrence (including transpressional deformation) and extension since ~30 Ma (Sempere et al., 2004; Sempere and Jacay, 2006, 2007), in contrast with the northeast Altiplano, Eastern Cordillera, and sub-Andean belt, where shortening has been indeed significant. Besides, the Pacific Andean escarpment is the locus of oceanward reverse faulting, suggesting incipient oceanward gravitational collapse of the Western Cordillera (Wörner and Seyfried, 2001; Wörner et al., 2002; Sempere and Jacay, 2006, 2007). Transpressional structures in the forearc, such as the Cordillera de Domeyko in Northern Chile (e.g., Reutter et al., 1991; Arriagada et al., 2000, 2003), can only account for relatively minor shortening and crustal thickening.

If it is confirmed that in the arc region orogeny was concurrent with extension, and with no or little contraction, it would inevitably further question the dominant paradigm, and instead support the idea that crustal thickening in this region was mainly achieved by magmatic accretion (as advocated by James, 1971b; Kono et al., 1989; James and Sacks, 1999). The hypothesis of Andean crustal growth by magmatic accretion at the arc clearly cannot be discarded, explaining in part the current renewed interest for Andean magmatic processes and products.

4.4. Magmatism as a window into deep crustal and mantle processes

It is well known that magmatic records provide invaluable information on deep processes. Andean magmatism can been used as a probe into the crust and mantle (Kay et al., 1999; Kay and Mpodozis, 2002) and provides means to peer into the processes operating at depth (e.g., Bock et al., 2000; Haschke et al., 2002b). Magmatism is not only one of the two prominent features of the Central Andes—from which the celebrated andesites received their names—, but also happens to conveniently sample the Andean crust (e.g., Wörner et al., 1992; Aitcheson et al., 1995; Mamani et al., 2005).

A zone of low resistivity and seismic velocity indicates that partially molten rocks occur below a large part of the Altiplano (e.g., Schilling et al., 2007), where a 10–4 Ma ignimbritic flare-up is interpreted to have been produced by the emplacement at depth of a giant silicic batholith (de Silva and Gosnold, 2007).

Because magmatism provides significant insights into deep crustal processes and structure, interest for related studies is expected to increase in the future. Accordingly, the Barcelona 2005 ISAG dedicated to this topic more than half a day of its oral sessions. In this special issue, magmatic rocks are used to address tectonic issues in papers by Vásquez and Franz in Chile, and by Jiménez and López-Velásquez in Bolivia. In the southernmost Central Andes, Folguera et al. (2008-this issue) link observations of extensional features with the westward migration of the arc and related extensive basaltic volcanism that have taken place since the Pliocene.

4.5. "The Andes before the Andes" as an insight into "Why the Andes?"

It has been long recognized that ancient features of the Andean lithosphere have deeply influenced actual Andean-age deformation (e.g., Allmendinger et al., 1983; Sempere et al., 2002) and therefore the pre-orogenic evolution of the Andes continues to attract attention. It is particularly intriguing that subduction of Panthalassan oceanic plates beneath the western margin of South America has been active for hundreds of millions of years and yet the Andes have formed only during the last tens. Why apparently did not any orogen of the magnitude of the present-day CAO form during a long period of similar geotectonic setting? So *why* did the Andes form? Answering such questions would certainly provide valuable insights into *how* the Andes formed. A first step toward an answer is to know, as precisely as possible, *when* the Andes started to be built, and what were the chronological steps taken by this build-up. When were high altitudes acquired? (see below). What was the tectonic context when the build-up started? What were these contexts during the times where no significant mountain belt existed? Studying the evolution of the Andean region before the mountain belt formed is indeed crucial, because a better knowledge of the main milestones of the Andean history should shed some light on the processes that built the Andes.

A first caveat should be stated in order to avoid the simplistic vision of a homogeneous margin that was suddenly submitted to orogeny. South America, including its western margin, is highly heterogeneous, consisting of a mosaic of pre-Mesozoic crustal domains, the geometry and rheology of which have generally had some significant bearing on Andean-age deformation, even far away from the Andes proper. This heterogeneity also stems from the fact that the Andean margin was submitted to a number of deformational episodes in the Proterozoic, Paleozoic, and Early Mesozoic (e.g., Mégard, 1978; Kay et al., 1989; Pankhurst and Rapela, 1998). Considerable progress has been recently obtained on the Middle Proterozoic to Early Mesozoic history of Peru, for instance (e.g., Loewy et al., 2004; Mišković et al., 2005; Chew et al., 2007).

Between 22°S and 52°S, in particular, the western margin of what is now South America underwent a complex history. Successive Andean-type continental arcs developed along the Panthalassan margin of western Gondwana, and "normal" to shallow subduction periods can be distinguished based on the contrasted extents of the coeval volcanic arcs (Ramos and Folguera, 2007). The closure of Neoproterozoic to Early Paleozoic backarc basins generated deformational belts along this margin (Astini and Dávila, 2004; Escayola et al., 2007; Mulcahy et al., 2007). From Late Proterozoic to Paleozoic times, a number of collisional belts formed in response to the accretion of exotic and parautochthonous terranes, the basement of which consists of Mesoproterozoic (Pampia, Cuyania, Chilenia) and Late Neoproterozoic–earliest Paleozoic (northern Patagonia) metamorphic rocks (Ramos, 1989, 2004; Rapela et al., 1998; Chernicoff and Zapettini, 2004; Pankhurst et al., 2006). This complex evolution resulted in a rheologically anisotropic basement that underwent mainly extensional to transtensional deformation from the Late Paleozoic to the Cretaceous. Ramos and Kay (1991) and Ramos et al. (2002) evidenced the role of Neoproterozoic–Early Paleozoic sutures in the development of Late Triassic rifts, whereas Schmidt et al. (1995) discussed the influence of such heterogeneities in the formation of Early Cretaceous basins between 27° and 33°S. In the same area, Fernández-Seveso and Tankard (1995) explained the formation of Late Paleozoic transtensional basins by reactivations of those lithospheric discontinuities. Similarly, Mosquera and Ramos (2006) related Late Triassic–Early Jurassic extensional basins to the collapse of the Early Permian Patagonian suture around 39°S (von Gosen, 2003).

Such extensional detachments and other heterogeneities provided potential décollements for most of the Andean fold-and-thrust belts between 25°S and 52°S, largely influencing the post-Early Cretaceous deformation geometry (Cristallini and Ramos, 2000; Kley et al., 2005;

Zapata and Folguera, 2005; Giambiagi et al., 2008-this issue). In particular, Andean shortening has easily propagated into thick Paleozoic sedimentary wedges that had accumulated in coeval foreland basins (e.g., the sub-Andean Belt at 17°–25°S, and the Precordillera system at 27°–32°S; Allmendinger et al., 1997; Cingolani et al., 2003; Astini and Dávila, 2004; Alonso et al., 2005). Conversely, the paleotectonic boundaries of these basins have resisted the propagation of Andean thin-skinned deformation, as is the case in the Santa Bárbara system and northern Sierras Pampeanas around 26°S in Northwest Argentina.

Based on a 450 km-long magnetotelluric profile at ~31°30'S in central Argentina, Favetto et al. (2008-this issue) identify the Cambrian-age suture between the ~2.3–2.1 Ga-old Río de la Plata craton in the east and the younger Pampean terrane in the west, unraveling at this latitude the deep structure of the South American basement east of the Andes. They show that this 500 Ma-old suture coincides with the eastern limit of the Sierra Chica de Córdoba, an Andean-age basement uplift (Ramos et al., 2002) and thus confirm that ancient structures are prone to reactivation, playing prominent roles in concentrating subsequent deformation.

Vásquez and Franz's (2008-this issue) study of the Cobquecura pluton confirms that present-day central Chile was the locus of anorogenic-type (A-type) magmatism in the Triassic, implying a coeval post-orogenic and/or extensional setting (Eby, 1992). The end of orogenic developments is known to involve not only erosion and transcurrent to extensional tectonic regimes (related to gravitational collapse of the overthickened crust and/or delamination of the lithosphere), but also emplacement of voluminous igneous formations (Bonin, 2004). The Triassic anorogenic magmatism documented in central Chile was most likely related to the major Choiyoi magmatism that developed at the end of the regional Late Paleozoic orogeny (Kay et al., 1989).

Jiménez and López-Velásquez (2008-this issue) review the known occurrences of Phanerozoic magmatic rocks in the Huarina belt of Bolivia, a tectono-stratigraphic unit that extends along the western Eastern Cordillera and in the southern Altiplano, and includes numerous tin deposits. Geophysical and isotopic data suggest that the Huarina belt was formed along a major, lithospheric-scale, weakness zone that separated three basement blocks and was probably inherited from the Proterozoic history. Most of the Phanerozoic backarc magmatism in the Bolivian Andes occurred along this belt, reflecting the permanence of this basement heterogeneity, which in particular controlled Mesozoic rifting processes (Sempere et al., 2002). When Cenozoic shortening gradually thickened the lithosphere, delamination of its denser and less viscous root was induced, which in turn produced mantle-derived magmas that generated and interacted with crustal melts and gave rise to the dominantly peraluminous magmas characteristic of the area.

Knowledge of the pre-Andean history thus also provides insights for the understanding of regional to local expressions of Andean-age tectonics. Inherited crustal heterogeneities are major features controlling the location, geometry and style of deformation in the entire Andes. At all scales, deformation was influenced by features ranging from regional paleotectonic and paleogeographic features to the simple presence of potential décollements in a sedimentary pile submitted to deformation (Jaillard et al., 2002). Many examples of compressional reactivation of extensional structures have been illustrated (e.g., Uliana et al., 1995; Kley et al., 2005). In this issue, more cases are documented by Carrera and Muñoz and by Giambiagi et al.

The stratigraphy and deformation of the many foreland basins of the Andes have provided invaluable information on Andean growth. In Peru and Bolivia, stratigraphic studies have long detected that some western relief emerged in Late Cretaceous times (e.g., Mégard, 1978), and the birth of some proto-Andes has been more precisely dated to near the Turonian–Coniacian transition (~90–89 Ma; Jaillard, 1994; Sempere, 1994; Jaillard and Soler, 1996), i.e. to more than 10 Myr after the final separation of South America from Africa. In central Argentina, the Andean foreland basin was created somewhat earlier, at some time during the 120–95 Ma interval, presumably in response to growing proto-Andes in the west (Orts and Ramos, 2006). At 45°–46°S, most contractional structures are unconformably post-dated by 110 Ma-old volcanic rocks (Folguera and Iannizzotto, 2004). At 39°–37°S dikes dated to 105–100 Ma post-date western compressional structures (Zamora-Valcarce et al., 2006). The foreland basin turned markedly continental during the Late Cretaceous, starting before 88 Ma (Corbella et al., 2004). At ~37°S a Late Cretaceous pulse of exhumation was initiated at ~80–70 Ma (Burns et al., 2006; Kay et al., 2006). Between 38° and 39°S, orogenic reliefs in the westernmost part of the basin are post-dated by 75–65 Ma-old volcanic rocks (Franchini et al., 2003).

4.6. Tectonic processes along the eastern side of the Andes: the backarc and foreland

A large part of the abundant literature concerning the backarc and foreland of the Andes was reviewed by Jaillard et al. (2002) and Gerbault and Hérail (2005). From a structural point of view, the backarc usually consists of an east-verging fold-and-thrust-belt (e.g., Allmendinger et al., 1983; Roeder and Chamberlain, 1995), which is generally the case in the sub-Andean zones, but the vergence, amount of shortening and style of deformation commonly vary along strike (e.g., Jordan et al., 1983; Kley et al., 1999). Foreland evolution depends on the flexural subsidence generated by the growing orogen, and on the sediment supply provided by its erosion. Other important factors are deformation styles in the orogen, large-scale thermal processes, climate (governing relief dissection), sediment transport and drainage patterns (e.g., Masek et al., 1994), eustatic sea-level changes, or any combination of these (Jaillard et al., 2002). Stratigraphic characteristics of the foreland infill indirectly inform on the nearby orogenic development.

In the Central Andes, shortening is generally well expressed in the backarc, especially in the Eastern Cordillera and sub-Andean belt (and longitudinal equivalents), but extensional phenomena are also increasingly being described at least in specific areas (e.g., Folguera et al., 2008-this issue). It is common that shortening has propagated by inversion of pre-existent extensional structures. One eloquent case is that of the Eastern Cordillera of central and southern Bolivia, where tectonic vergence is typically to the west in its western half, and to the east in its eastern half (e.g., McQuarrie, 2002), an enigmatic geometry that has been explained by the reactivation of a Mesozoic rift structure located in the axis of the cordillera (Sempere et al., 2002).

In this issue, the eastward propagation of Andean deformation is addressed by Carrera and Muñoz (2008-this issue) through a study of Cenozoic growth strata and unconformities in the southern Cordillera Oriental of northern Argentina (25°30'S). They constrain the ages of the local structures and the timing of deformation propagation, highlighting that thrusting resulted largely from inversion of Cretaceous extensional faults (Carrera et al., 2006). Thrust propagation rates culminated in the Late Miocene–Early Pliocene and Quaternary.

About 1000 km more to the south, the Malargüe fold-and-thrust belt (34°–36°S, Argentina) formed during the Neogene by shortening of part of the Mesozoic Neuquén basin. On the basis of detailed structural data and new $^{39}Ar/^{40}Ar$ datations, Giambiagi et al. (2008-this issue) show that, contrary to prevalent models, deformation propagated from the foreland to the hinterland, and that inversion of normal faults involving the basement was coeval with insertion of shallow detachments and low-angle thrusts. Coeval activities of shallow and deep detachments produced simultaneous thrusting during complex deformation at the thrust front between 15 and 8 Ma.

The shallow subduction that characterized the southernmost Central Andes (35°–37°30'S) from 13 to 5 Ma forced the arc to migrate toward the foreland and produced shortening more than 550 km east of the trench (Ramos and Folguera, 2005). Subduction shifted back to a more classical, steeper geometry at 5 Ma and the arc front re-established in its current western position (Muñoz and Stern, 1988; Kay et al., 2006). Based on detailed mapping and seismic, gravity, and geochronologic data, Folguera et al. (2008-this issue) show that this steepening of the slab had a number of consequences in the local backarc: widespread flows of intra-plate basalts piled up into a plateau in the eastern foreland, and between the latter and the arc front a major tectonic trough formed, controlling the emplacement of crustal and primary mantle-derived melts. The main episode of crustal collapse developed during the 1.7–0.7 Ma interval, and has lingered to the present. A similar evolution is documented at 46–48°S (Lagabrielle et al., 2007).

4.7. Tectonic processes along the western side of the Andes: the forearc

The forearc is a key feature of the entire Andes, but its structure and characteristics vary considerably along strike. It is currently attracting much attention, and more than half a day was dedicated to this topic at the Barcelona 2005 ISAG. The Andean forearc is notoriously dominated by extension, especially in its westernmost parts, where subsidence is classically interpreted to result from tectonic erosion beneath the margin (e.g., von Huene and Scholl, 1991; von Huene et al., 1999; von Huene and Ranero, 2003; Clift and Hartley, 2007). A possible link between subduction seismicity and eustatic changes has been recently proposed (Bourgois et al., 2007).

The variety of forearc structures and deformation along the Andes was reviewed in earlier prefaces (e.g., Jaillard et al., 2002). Deformational structures may be parallel, orthogonal, or oblique to the trench. Aside from extensional features, transpressional and compressional deformations and related basins have been described (e.g., Hartley et al., 2000; González et al., 2003, 2006; Victor et al., 2004; Mpodozis et al., 2005). Regional surface tilting has been suggested (e.g., Isacks, 1988) as well as the occurrence of gravitational tectonics (e.g., von Huene et al., 1999; Wörner and Seyfried, 2001; Wörner et al., 2002).

Climate is notoriously semi-arid to hyperarid along the Central Andean forearc, a common feature of western regions of continents, at these latitudes, that was enhanced by the rain shadow effect imposed by the growing Andes (e.g., Houston and Hartley, 2003; Dunai et al., 2005; Quang et al., 2005). Interaction of climate and orogeny has grown to a major research issue (e.g., Masek et al., 1994; Avouac and Burov, 1996; Pinter and Brandon, 1997; Willet, 1999). The idea that in orogens erosion and sedimentation have a significant bearing on tectonics has been applied to the Andean backarc and foreland (e.g., Montgomery et al., 2001; Thomson et al., 2001; Thomson, 2002), and more western regions (e.g., Sobel et al., 2003). Lamb and Davis (2003) even proposed that climate, through the processes of erosion and sediment deposition, has ultimately controlled the growth of the Central Andes by causing plate boundary stresses to increase where sediment starvation occurs in the trench; this interpretation, however, has been challenged (Hartley, 2005).

The issue of the respective roles of climate and tectonics is addressed by Nalpas et al. (2008-this issue), who use sedimentologic and tectonic observations and geochronologic data to investigate the Atacama Gravels, an extensive blanket of Miocene continental deposits that fill a Neogene paleo-valley system at ~26°30'S in northern Chile. Deposition of this unit began at about the Oligocene–Miocene boundary and ceased in the Late Miocene, and occurred in environments ranging from proximal alluvial fan to playa-lake. Because no synsedimentary deformation is observed, the Miocene change from semi-arid to hyper-arid climatic conditions appears to have been the dominant factor controlling sediment preservation.

Following the M_w = 8.4 earthquake that propagated from NW to SE along the forearc of southern Peru on June 23, 2001, Audin et al. (2008-this issue) map and characterize active faults in the Ilo area (~17°30'–17°45'S). The overall characteristics of the 2001 earthquake, the subsequent seismic events, and the Quaternary activity of the coastal faults suggest that both plates are strongly coupled, that the subduction plane in southern Peru is segmented, and that this segmentation may be imposed by the continental plate structure itself, which thus may have some control on the rupture pattern of major subduction earthquakes along southern Peru.

The geomorphology of the Andean forearc has historically been viewed as an old remnant of a late Miocene planar landscape with no significant active structures accommodating Quaternary deformation. Applying cosmogenic isotope techniques to geomorphologic observations, Hall et al. (2008-this issue) study a well-preserved sequence of planation surfaces and strath terraces developed in the forearc of southern Peru, as well as abrupt changes in topography and drainage incision, and demonstrate that this region has undergone deformation and uplift in the Quaternary, in contrast with previous interpretations favoring that abandonment of these surfaces resulted from Late Miocene uplift. Using in-situ-produced ^{10}Be, they date pediment surfaces between ~1003 and ~119 ka, and estimate incision rates to have ranged between 0.04 and 0.3 mm/yr. Active deformation within the forearc highlights a sharp contrast in deformation style between the eastern and western margins of the Andes.

Contardo et al. (2008-this issue) describe the development of Quaternary forearc marine basins offshore Central Chile (33°30'–36°50'S) through a study of high-resolution seismic-reflection profiles. The middle and upper slope of this accretionary margin is characterized by half-grabens that are filled by three distinctive sequences of Middle Pleistocene to Holocene ages. Their deposition and deformation suggest alternating episodes of compression and/or transpression, relative stability, and extension, that were induced by the subduction context. Large-scale mass transfer processes are also documented. Positive flower structures indicate current transpressional activity associated with tectonic inversion and differential uplift of the accretionary prism. Slope deformation and tilting are likely to be driven by basal accretion of large volumes of underplated sediment underneath the forearc, depending on the climate-controlled volume of sediments available at the trench and on the basal properties of the prism.

4.8. When did the Andes become the Andes? The Andean paleoaltitude issue

Although a large amount of work has been performed in the Central Andes during the last decades, the chronology of crustal thickening and related uplift remains debated. For instance, in the CAO, onset of the main orogeny in the Eastern Cordillera has been estimated at ages as contrasted as ~26 Ma (Sempere et al., 1990) and ~40 Ma (Lamb and Davis, 2003). According to fission-track analyses, uplift in the CAO seems to have slowly started about 46 Ma (Anders et al., 2002), and exhumation rates in the Eastern Cordillera near La Paz, Bolivia, would have increased at ~25±5 Ma (Kennan, 2000) or ~15–10 Ma (Masek et al., 1994). Several sets of paleoaltitudes estimates are now available (e.g., Ghosh et al., 2006; Garzione et al., 2006, 2007; Quade et al., 2007). Andean uplift has thus become an intensely debated issue since a few years, and half a day of the oral sessions of the Barcelona 2005 ISAG was dedicated to this topic only. Although a general picture is somehow emerging, more independent data are needed in order to solve current discrepancies.

Andean uplift is being investigated from different approaches. While estimation and timing of rock uplift and exhumation are addressed by thermochronologic methods, the amount and timing of surface uplift, i.e. the increase in elevation of the local Earth's surface, require paleoaltitude estimates.

Such estimates were initially drawn from the analysis of fossil leaf morphology: according to this method, a locality of the western Bolivian Altiplano, close to southern Peru and now at 3.94 km elevation, was only at 1.16 ± 0.6 km at 10.7 Ma (Gregory-Wodzicki et al., 1998; Gregory-Wodzicki, 2000, 2002). Kowalski (2002) however showed that, at least in Bolivia and East Asia, the current leaf morphology method systematically underestimates altitudes when these are high (as it overestimates MATs, by as much as 15 °C) because applying equations generated from forests in North America to unrelated forests is inaccurate. Low Late Miocene paleoaltitudes estimated by fossil leaf morphology in the CAO are therefore likely to be underestimations.

On the basis of measurements of, respectively, $\delta^{18}O$ and ^{13}C–^{18}O binding rates in paleosol carbonate nodules, the central Altiplano was reconstructed to have been before 10.3 Ma between 0.4 and 2.2 km (Garzione et al., 2007, correcting estimates by Garzione et al., 2006) or, more precisely, between 0.4 and 0.8 km (Quade et al., 2007, correcting estimates by Ghosh et al. 2006); and after 6.8 Ma between 4.0 and 4.7 km (Garzione et al., 2007) or between 2.9 and 3.7 km (Quade et al., 2007), depending on the method used. The accuracy of initial (uncorrected), lower estimates by Ghosh et al. (2006) and Garzione et al. (2006) was questioned, respectively, by Sempere et al. (2006; reply by Eiler et al., 2006) and Hartley et al. (2007; reply by Garzione et al., 2007) because they were in apparent contradiction with the geological record. The few paleoaltimetric results favoring that the Central Andes were at low elevations before ~ 10 Ma thus appear to need confirmation, or, in any case, the upper end of uncertainty intervals might provide preferable estimates.

In fact most studies favor that Andean uplift started in the Eocene or Oligocene, but many do recognize a subsequent resumption of uplift starting in the Late Miocene. Gillis et al. (2006) identified two phases of rapid cooling from 45–40 Ma to 26 Ma and from ~ 11 Ma onward. Also in Bolivia, similar conclusions are reached by Barnes et al. (2006): initial rapid erosion of the plateau margin took place during the ~40–25 Ma interval and has been followed by widespread accelerated erosion since ~ 15 Ma. In the southern Altiplano of Bolivia, Ege et al. (2007) recognized that a major, plateau-wide exhumation developed in the Early Oligocene (33–27 Ma). Based on geomorphologic observations, Sébrier et al. (1988) concluded that the Andes of southern Peru were ≥ 2.0 km-high at ~ 20–17 Ma. In the Pacific Andean slope of southernmost Peru, regional uplift triggered incision of deep valleys into a thick, 23–18 Ma old, ignimbrite blanket shortly after the end of emplacement of these ignimbrites, thus starting ~ 18–17 Ma (Flores and Sempere, 2002). Surfaces in the Atacama Desert, Northern Chile, have been barely affected by erosion since 25 Ma, as documented by cosmogenic ^{21}Ne measurements in exposed clasts, and are the oldest known continuously exposed surfaces on Earth (Dunai et al., 2005); although climate in western South America have been somewhat arid since at least the Late Cretaceous (e.g., Sempere et al., 1997; Quang et al., 2005), hyperarid conditions were required to preserve such surfaces and are likely to have been enhanced at this latitude by a significant uplift of the Andes at that time. A variety of independent methods and studies thus apparently agree to place in the Oligocene–Early Miocene interval the culmination of a first significant uplift period, and thus of one major episode of crustal thickening.

Several works, however, coincide in placing at about 10–9 Ma the onset of a second major episode of uplift (e.g., Ghosh et al., 2006; Garzione et al., 2006; Barke and Lamb, 2006; Schildgen et al., 2007; Thouret et al., 2007). A distal, Atlantic, sedimentary record of Andean erosion identified a significant increase in Andean-derived detrital material at ~ 10–9 Ma (Dobson et al., 2001). In southern Peru, canyons incising the Pacific slope of the Andes had reached their current depth by ~ 4 Ma, were filled by volcanic products by the earliest Quaternary, and were subsequently re-incised (Thouret et al., 2007). Erosion has been dominant in the Pacific slope of southernmost Peru since 2.7 Ma (Flores and Sempere, 2002).

The picture that is thus emerging is one of a mountain building achieved in two steps, the first slowly developing from the mid-Eocene and reaching an acme in the Late Oligocene and Early Miocene, and a later, apparently vigorous step starting at ~ 10–9 Ma and possibly lingering into the Pliocene and/or Present.

5. Concluding remarks

Perhaps one major point to emphasize about the Andes is that their contrasted segmentation reflects that they have been built by a variety of processes which have differed along and across strike in nature, time, and intensity. It is now particularly evident that nearly every Andean geological feature varies along and across strike, and that conclusions reached in one area can rarely be generalized to other regions. Andean studies may therefore have entered a new period, as more modern methods are employed, more specific cases are documented, more interdisciplinary works are conducted, more comparisons are performed with other orogens worldwide, and more discrepancies appear and need to be resolved. As a consequence current concepts can be re-evaluated at all scales, making possible in the future to lift the weight of the current paradigm.

The Central Andes contrast with the Northern Andes in that their history has been devoid of truely collisional phenomena, and with the Southern Andes in that ocean–continent subduction has built considerable orogenic volume in the former, with crustal thickness reaching values currently known only in the Himalayas and Tibet. Because no real consensus exists yet, as indicated by the many ongoing debates, the genesis of the Central Andes remains somewhat enigmatic.

For many decades geoscientific research in the Andes has produced an impressive wealth of data, but, as described by Kuhn (1962) for all sciences, production of knowledge has been often oriented by the dominant paradigms, namely by the concept of “Andean tectonic phases” (Steinmann, 1929) until the mid-1980s, and by the Isacksian paradigm since then. Progress may therefore be expected from the confrontation of counterexamples with relevant aspects of the current paradigm, and by resolving present discrepancies through the obtention of modern, independent data and their intellectual articulation with older reliable data. There is little doubt that interdisciplinary approaches will be increasingly conducted within frameworks where tectonic and magmatic processes are envisioned as two complementary aspects of one orogenic system through time. It is likely that integration of interdisciplinary data and results will lead to original conclusions about processes of crustal thickening and growth in continental arcs, and shall improve our understanding of the lithospheric processes that determine the evolution of basins, deformation, and magmatism observed in the upper crust of these orogens.

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