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Preface

Andean geodynamics: main issues and contributions from the 4th ISAG, Göttingen

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1. Introduction

The Andean chain is the paradigm of a continental orogeny resulting from the subduction of an oceanic plate beneath a continental plate. Despite an apparent simple general geometry, the tectonic–magmatic evolution of the Andes is not entirely resolved, and some of the basic processes responsible for the growth of this orogeny remain poorly understood. A number of striking questions is only partly resolved and need much more investigations. We briefly summarize here some of these questions.

(1) What is the present-day deep structure of the mountain belt? What are the relative parts of tectonic shortening, slab geometry, tectonic and magmatic underplating in the volume and composition of the Andean crust? What are the possible consequences of such geometry on the thermal structure of the Andean belt?

(2) What are the external forces and processes which cause the horizontal shortening of the continental plate, and what is the role of the lithospheric mantle during crustal shortening?

(3) What are the timing, nature and location of crucial deformational events that contributed to the building up of the mountain belt? Are they typical "orogenic phases" occurring during restricted period of times, which can be correlated along major portions of the chain? What may explain the temporal and spatial variations in such tectonic events?

(4) What are the gravitational and climatic responses to crustal thickening and uplift? What are the feedback controls of such modifications on erosion and uplift?

Addressing these questions ideally requires, at least, interactions between structural geologists, sedimentologists, geophysicists and geochemists, as well as geochronologists, geomorphologists, paleobotanists and modelling specialists. The 4th International Symposium on Andean Geodynamics (ISAG) meeting held in Göttingen on October 4–6 1999 aimed to assemble some 250 scientists from these fields of earth science to present and discuss recent progress in the understanding of this prominent mountain belt.

This meeting occurred at a transition period. During the late 1990s, several reviews have been published on Andean geodynamics and more papers synthesizing various aspects of Andean evolution are coming out. The state of our knowledge in the geological, tectonic and geodynamic evolution for the Andes was presented by Salfity (1994), Reutter et al.

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(1994), Tankard et al. (1995a,b), Dewey and Lamb (1996), Allmendinger et al. (1997), ISAG (1999), Cordani et al. (2000a,b) and Miller and Hervé (2000), many of which are based on the earlier landmark paper by Isacks (1988). The climate and faunal response to uplift and its timing were reviewed by Gregory-Wodzicki (2000). More geophysical work introducing new techniques to the Andes (e.g., seismic reflection) was combined with the geological evidence for Andean-wide lithosphere imaging (Beck et al., 1996; Dorbath et al., 1996; ANCORP Working Group, 1999; Beaumont et al., 1999; Reutter, 1999), including the offshore regions and marine geology (CINCA Study Group, 1996; Von Huene et al., 1999; Bourgois et al., 2000).

The application to the Andes of other new methods, such as GPS measurements (e.g., Kellog and Vega, 1995; Norabuena et al., 1998; Kendrick et al., 1999; Weber et al., 2001), cosmogenic isotope datings (e.g., Siame et al., 1997) or tomographic imaging (e.g., Dorbath and Granet, 1996; Engdahl et al., 1998; Beaumont et al., 1999; Graeber and Asch, 1999; Masson et al., 2000) and refinements of existing techniques, like isotopic mapping (e.g., Bock et al., 2000; Wörner et al., 2000) have all advanced our basic understanding of this unique mountain range.

Accordingly, the 4th ISAG meeting brought together many of the scientists involved in these studies. Many papers presented at the conference and some of the papers included in this volume are also directly related to the current UNESCO International Geological Correlation Program (IGCP 436) "Tectonic evolution of the Pacific Gondwana margin structure, assembly and break-up events." IGCP, ILP, the University of Göttingen and other organizations and companies (DFG, IRD, State of Niedersachsen, ER-Mapper and the Wintershall) provided financial support for the meeting, which is greatly acknowledged.

2. ISAG 99 contributions

The papers from this symposium gathered in this section/volume of Tectonophysics address several of the questions raised above (Fig. 1) and can be grouped into three main themes. After broadly presenting these topics, we intend to outline the respective contributions of the presented papers.

2.1. Processes of crustal thickening in the Andes

2.1.1. Central Andes (≈ 5°S–45°S)

The Central Andes is characterized by a high mean elevation resulting from a strongly thickened crust and a well-developed thin-skinned fold and thrust belt to the east (sub-Andean zone). No exotic terranes have been accreted since the Jurassic. Crustal thickening processes have been studied mainly in the Altiplano (≈ 15°S–25°S), the mean elevation of which is about 3900 m (Isacks, 1988) and the crustal thickness of which reaches 70–75 km (Lyon-Caen et al., 1985; Fukao et al., 1989; Beck et al., 1996; Schmitz et al., 1999; Yuan et al., 2000) (Fig. 2).

For early authors, addition of melt volume at depth due to intense magmatic arc activity was thought to be responsible for the crustal thickening and correlative relief increase of the Central Andes (e.g., James, 1971; Thorpe et al., 1981). Recent estimates show that magma addition would explain only a few percentages of the observed crustal thickness (Francis and Hawkesworth, 1994; Allmendinger et al., 1997; Giese et al., 1999). Nevertheless, magmatism has an acknowledged influence on crustal rheology (Shemenda, 1994; Allmendinger et al., 1997) and might locally contribute to the thickening of active arc zones (Lamb et al., 1997).

Further studies of the deformation of the upper crust and overlying sediments demonstrated that compressional shortening of the upper plate and thrusting of the Andean chain upon the Brazilian Shield are the driving mechanisms of crustal thickening (e.g., Allmendinger et al., 1983; Jordan et al., 1983; Isacks, 1988; Roeder, 1988; Sempere et al., 1990; Sheffels, 1990). Recent works, however, point out that the observed tectonic shortening rates do not account for the observed crustal thickening, especially beneath the arc and forearc zones (Schmitz, 1994; Giese et al., 1999; Ramos and Alemán, 2000; Yuan et al., 2000) and that material derived from the tectonic erosion of the upper plate edge (Rutland, 1971) may have been underplated beneath the western parts of the chain (e.g., Schmitz, 1994; Baby et al., 1997; Kley and Monaldi, 1998; Rochat et al., 1999) (Fig. 2).

Shortening of the ductile lower crust weakened by the magmatic arc activity and the undergoing distributed deformation may have contributed signifi-

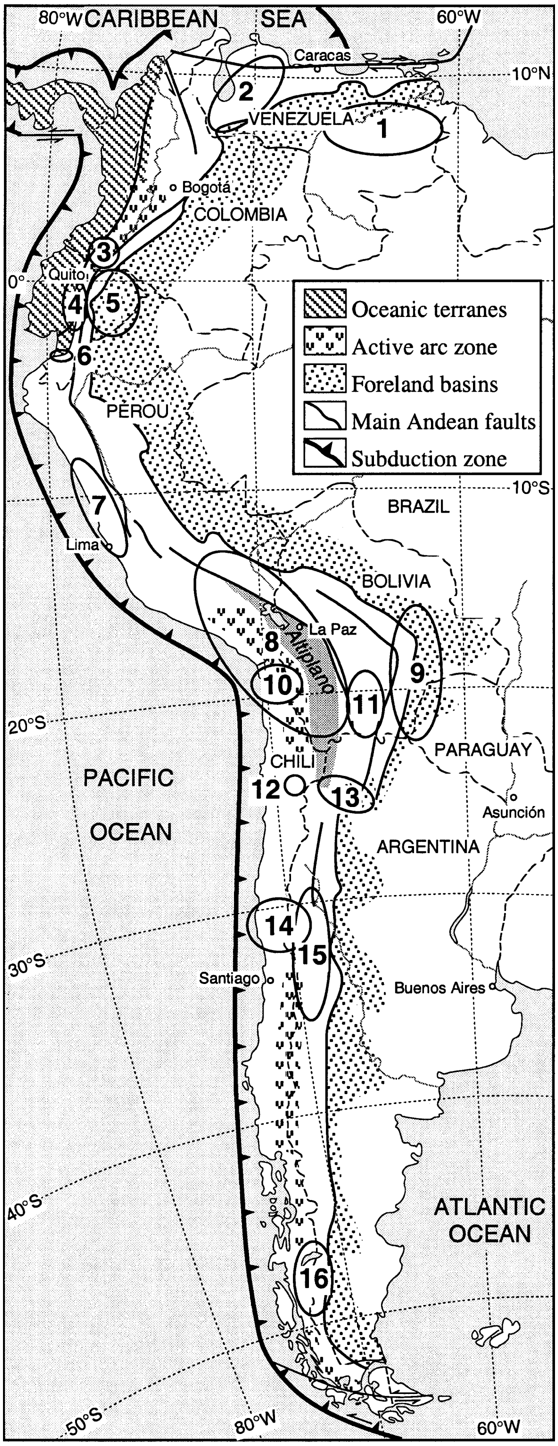

Fig. 1. Geological sketch of the Andes of South America and the location of the areas addressed by the presented papers. 1: Schmitz et al.; 2: Audemard and Audemard; 3: Weber et al.; 4: Hughes and Pilatasig; 5: Christophoul et al.; 6: Bosch et al.; 7: Haederle and Atherton; 8: Sempere et al.; 9: Husson and Moretti; 10: Wörner et al.; 11: Jacobshagen et al.; 12: Campos et al.; 13: Matteini et al.; 14: Pardo et al.; 15: Thomas et al.; and 16: Wenzens.

cantly to the thickening of the Altiplano crust before the activation of the present-day sub-Andean zone (Isacks, 1988; Gubbels et al., 1993; Allmendinger et al., 1997; Lamb, 2000). An alternate explanation is that the geophysical Moho corresponds to the base of the hydrated mantle of the forearc zones rather than to the crust–mantle boundary (Giese et al., 1999; Yuan et al., 2000). Therefore, the estimation of crustal thickness should be changed to lower values.

On the other hand, the role of transcurrent movements, which add crustal forearc slivers or partition on the deformation within the chain, has been recently emphasized (e.g., Beck, 1988; Abels and Bischoff, 1999; Coutand et al., 1999; Roperch et al., 2000; Lamb, 2000).

Between 15°S and 25°S, we have now relatively good constraints on when, where and how much tectonic shortening took place, what the rates of tectonic erosion at the leading upper plate edge are, and what the magmatic addition rates should be and where it took place. From their analysis, Wörner et al. (10, Fig. 1) conclude that the volume observed in the crustal balance occurred mainly in Upper Oligocene times and can be located between the present arc zone and the central Altiplano.

Similar mechanisms (tectonic shortening, underplating of material removed from tectonic erosion and magmatic addition) have been invoked to explain the relief of the Andes of Peru (e.g., Mégard, 1987; Vicente, 1989). Using gravimetric data, which complete former extensive mapping and petrologic analysis (Pitcher et al., 1985), Haederle and Atherton (7, Fig. 1) state that the huge Cretaceous Coastal Batholith of Peru is 2–3 km thick by 30–60 km wide, and therefore, contributes only to a minor part of the total crustal thickness of this arc zone. These new geometrical data bring new constraints on the Batholith emplacement.

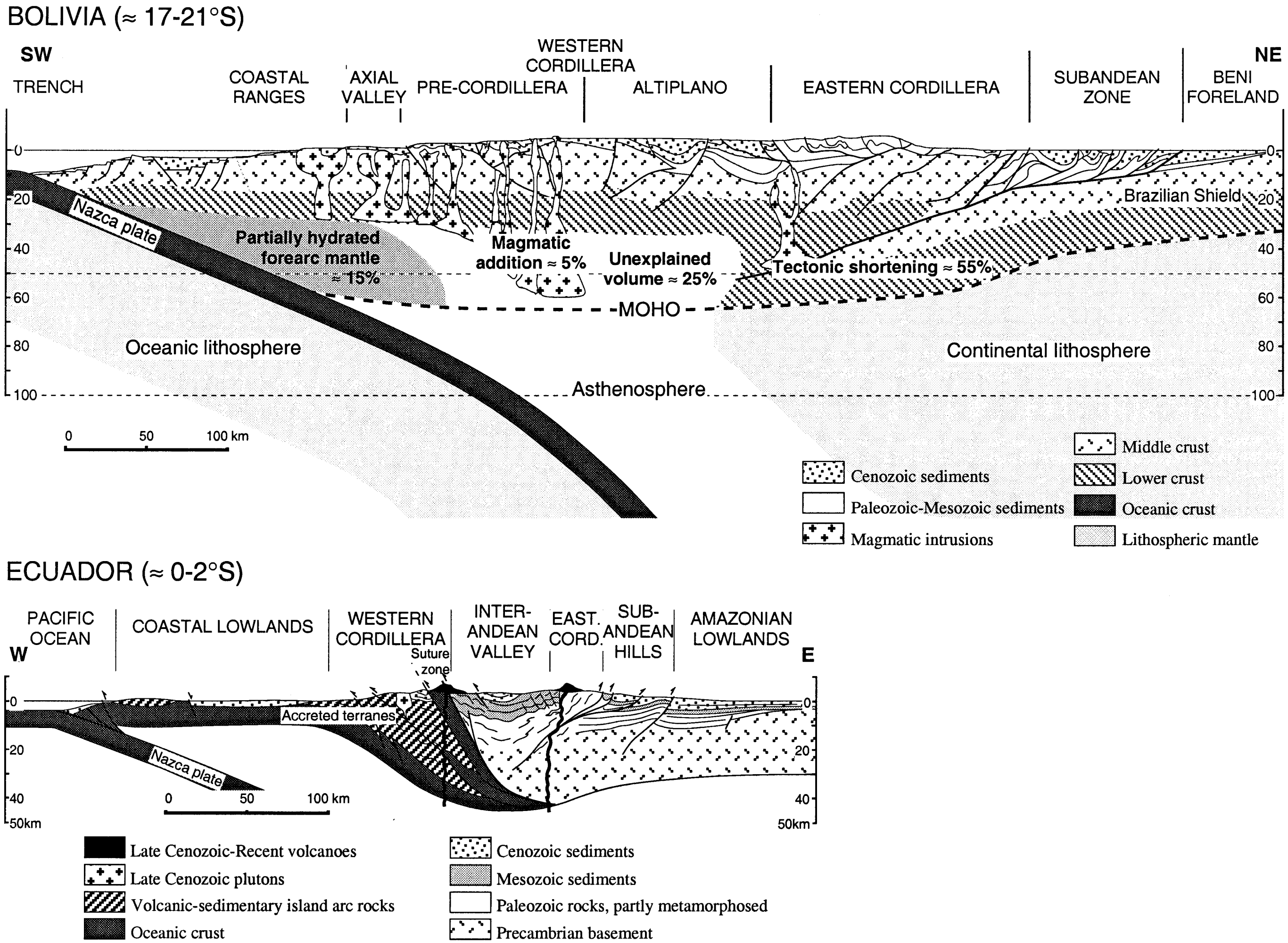

Fig. 2. Compared crustal-scale section of the Andes of Bolivia (compiled from Allmendinger et al., 1997; Scheuber and Giese, 1999; Giese et al., 1999; Rochat, 2000) and Ecuador (from Mégard, 1989). The "unexplained volume" below the Arc–Altiplano in Bolivia may be partially represented by the underplated material proceeding from the subduction erosion of the continental edge (see text).

2.1.2. Northern Andes (≈ 5°S–11°N)

The Northern Andes differ from the Central Andes in that (1) the present forearc and arc zones are made up of exotic oceanic terranes accreted during the Mesozoic and Tertiary (Gansser, 1973; Mégard, 1987; Kerr et al., 1998; Reynaud et al., 1999; Taboada et al., 2000); (2) the belt shows minor width and average elevation; and (3) the eastern foreland zone is dominantly of thick-skinned style and has a relatively restricted extension (Colletta et al., 1990; Balkwill et al., 1995; Rivadeneira et al., 1999) (Fig. 2). The northern Andean orogeny is considered to result mainly from compression related to the accretions of oceanic terranes (Feininger and Bristow, 1980; McCourt et al., 1984; Lebrat et al., 1987; Toussaint and Restrepo, 1994; Litherland et al., 1994; Spikings et al., 2001; Guillier et al., in press) and from right lateral movements (Winter and Lavenu, 1989; Freymüller et al., 1993; Kellog and Vega, 1995).

Based on the part of a programme of extensive mapping of the Western Cordillera of Ecuador (e.g., Boland et al., 2000), Hughes and Pilatasig (4, Fig. 1) have shown that the Western Cordillera of Central Ecuador consists of juxtaposed terranes. A first terrane made up of oceanic magmatic rocks showing MORB or oceanic plateau affinities accreted during the Late Cretaceous (Campanian). A second terrane made up of magmatic and sedimentary rocks of island arc origin accreted during the Middle to Late Eocene. These terranes are presently separated by a major right lateral fault, probably active since the Late Eocene–Oligocene.

Few authors have studied the deep structure of this part of the Andes and attempted to explain the isostatic balance of its relief. Some authors, however, suggested that accretions of oceanic terranes can lead to the underplating of oceanic material, which contributes significantly to the crustal thickening of this type of accretion orogeny (Mégard, 1989; Arculus et al., 1999; Guillier et al., in press) (Fig. 2). The study of recent xenoliths and Cretaceous eclogites provides insights on the composition of the deep crustal parts of the (present and past) northern Andes.

Weber et al. studied such xenoliths sampled by a recent volcano of the arc zone of southern Colombia (3, Fig. 1). The chemical composition and metamorphic facies and evolution of these xenoliths indicate that they proceed from deep levels of the continental crust and from deeply buried fragments of accreted oceanic terranes.

In the forearc zone of southern Ecuador, mainly basic High-Pressure (HP) metamorphic rocks exhumed 130–120 Ma ago are surrounded by HT metamorphic rocks representing deep levels of the continental crust (Aspden et al., 1995). According to Bosch et al. (6, Fig. 1), the HP rocks include material of sedimentary origin derived from the erosion of a continental crust as well as mafic and ultramafic eclogites of oceanic plateau affinity, suggesting that the oceanic plateau jammed the subduction and dragged down a sedimentary accretionary prism. They are associated with the remnants of a normal oceanic crust metamorphosed in greenschist facies.

2.2. Role of inherited features in Andean structure and evolution

Crustal heterogeneities or particular lithologic domains pre-existing to the Andean deformation have long been recognized as major features controlling the location, geometry and style of compressional deformation in orogenic belts (e.g., Jackson, 1980; Hayward and Graham, 1989; Lowell, 1995). Similarly, it has been abundantly documented that Andean deformations are influenced by the pre-Andean crustal history (Dalmayrac et al., 1980; Götze et al., 1994; Tankard et al., 1995a,b), the Mesozoic paleogeography (Mitouard et al., 1992; Uliana et al., 1995; Welsink et al., 1995; Rivadeneira et al., 1999) or the presence of décollement layers in the sedimentary pile (Allmendinger et al., 1983, 1997; Baby et al., 1989). Thus, the pre-Andean history may be a key to the understanding of the local expression of the Andean tectonic shortening and build-up.

2.2.1. Pre-Andean evolution (> 220 Ma)

Pre-Andean, stable South America is made of pre-Cambrian cratons (or shields), pre-Cambrian to Paleozoic-accreted continental terranes and Paleozoic orogenic belts mainly located on the western margin of Gondwana (e.g., Rapela et al., 1998; Bahlburg et al., 2000; Cordani et al., 2000a,b; de Almeida et al., 2000) and subsequent sedimentary accumulations. In the Northern Andes, few are known about the Paleozoic tectonic evolution (Ramos and Alemán, 2000).

In southern Venezuela, the 3- to 4-Ga-old Guayana shield reaches unusual elevations (1200–3000 m). Using gravimetric and seismic refraction data, Schmitz et al. explore the deep structure of this craton (1, Fig. 1) and determine an overthickened crustal thickness of 45–50 km.

South of the Arica Elbow (≈ 20°S), the Late Precambrian and Paleozoic evolution of what was the Western Gondwana margin is better understood. This evolution and possible crustal additions through time by arc magmatism have been studied on the basis of geochemical data on igneous and sedimentary rocks (Kay et al., 1999; Bock et al., 2000).

Jacobshagen et al. studied deformed Early Paleozoic rocks of southern Bolivia (11, Fig. 1). Illite crystallinity studies indicate an anchi- to epizonal metamorphism, while new radiometric data yielded a Late Hercynian age (320–290 Ma) for the deformation. Synthesizing other recent data, they propose that this deformation resulted from the eastward propagation of deformations during the late stages of the Hercynian orogeny known farther west.

Thomas et al. consider three distinct models for the collision of a continental terrane to the Gondwanan margin of south America (Precordillera, Argentina) (15, Fig. 1) in the Ordovician. At that time, Gondwana is assumed to be separated from the supercontinent Laurentia (North America and Greenland) by the Iapetus Ocean. After critical discussion of the syn- and post-orogenic sedimentation and paleogeography, they conclude that the accreted terrane was a microcontinent independent from Laurentia.

2.2.2. Extensional structures and their Tertiary tectonic inversion

The Late Paleozoic–Cretaceous evolution of South America has been marked by various episodes of extensional deformation and magmatism, most of which being coeval with active oceanic subduction. In the Andean domain, the areas affected by these extensional features seem to shift eastward through time. In Permian–Early Jurassic times, the break-up of Gondwana and Tethyan rifting were associated with local thermal anomalies leading to crustal melting and locally extensive volcanism (e.g., Kontak et al., 1985; Pichowiak et al., 1990; Litherland et al., 1994; Marzoli et al., 1999; Pankhurst et al., 2000) and widespread sedimentation controlled by extensional faulting (e.g., Jaillard et al., 1990; Flint et al., 1993; Parnaud et al., 1995; Ramos and Alemán, 2000). Rifting of the South Atlantic in the Late Jurassic–Early Cretaceous was then responsible for extensional features and magmatic manifestations (e.g., Uliana and Biddle, 1988; Turner et al., 1994; Ardill et al., 1998; Viramonte et al., 1999), which continued locally until early Late Cretaceous times (e.g., Soler and Sempéré, 1993; Jaillard et al., 2000) although transpressional deformations already began in the arc and forearc zones.

Sempere et al. (8, Fig. 1) reassess and update the extension-related, tectonic and magmatic manifestations which affected the Andean margin of Peru and Bolivia between Permian and Jurassic times. They propose that this "rift" structure controlled the location and nature of subsequent compressional deformations and allowed the individualization of crustal blocks characterized by their distinct tectonic behaviour and evolution, such as the Bolivian Altiplano.

As outlined by Myers (1975) and illustrated by Haederle and Atherton, the emplacement of the Coastal Batholith of Peru was controlled to the east by a regional major vertical fault system (Tapacocha Axis).

2.3. Tertiary Andean uplift and crustal thickening

The along-strike structure of the Andean chain is classically divided into: (1) a forearc zone, which corresponds to the pacific slope and offshore areas; (2) an arc zone mainly represented by the present chain (and Altiplano); and (3) a back-arc area, which includes the Eastern Cordilleras and Amazonian slope and foothills and the eastern lowlands underlain by the foreland basin. Since the Tertiary, each zone is dominated by distinctive deformation styles.

Deformations in the forearc zones are complex and still poorly known (Scheuber and González, 1999). Western submarine forearc areas are commonly marked by extensional sedimentary basins (Moberly et al., 1982; Hartley et al., 2000), the subsidence of which is related to the tectonic erosion of the continental margin (Von Huene and Scholl, 1991; Jaillard et al., 2000). To the east, compressional deformations (Vicente, 1989; Muñoz and Charrier, 1996; García et al., 1996; Hungerbühler et al., in press), strike-slip faults (Daly, 1989; Cembrano et al., 1996; Reutter et al., 1996) and vertical uplift (Sébrier et al., 1988; Macharé and Ortlieb, 1992; Steinmann et al., 1999; Hartley et al., 2000) are observed (Fig. 3). This uplift seems to have occurred mainly after about 10 Ma (Sébrier et al., 1988; Gregory-Wodzicki, 2000). Between 15°S and 25°S, it has been interpreted as the result of lower crustal flow and regional tilting rather than of tectonic shortening (Isacks, 1988; Allmendinger et al., 1997; Lamb et al., 1997). Due to the uplift and subsequently increasing topographic gradient, gravitational landslides are common from small scales to giant offshore slumps (e.g., Moberly et al., 1982; Bourgois et al., 1988; Duperret et al., 1995; Von Huene et al., 1999).

In the forearc zone of northernmost Chile, Wörner et al. (10, Fig. 1) interpret a large rotated block as a huge gravitational slump of Late Miocene age, relayed by a secondary landslide, both resulting from Miocene uplift, increasing aridity and subsequently decreasing erosion. The evolution of this area is characterized since the Miocene by (1) uplift, (2) erosion–sedimentation and (3) large volumes of felsic volcanism. Uplift would be related to flat subduction periods inducing strong coupling, whereas large volcanic eruptions would correspond to periods of steeper subduction, allowing asthenospheric heat flow and crustal melting.

Deformational structures in the arc and forearc areas are often obscured by widespread and thick volcanic covers. Where visible, they are usually dominated by compressional folds and reverse faults, the vergence of which varies according to the localities (Mégard, 1987; Muñoz and Charrier, 1996; Rochat et al., 1999) or to the authors (e.g., Bourgois et al., 1987; Kerr et al., 1998). Uplift usually began by Eocene times and became very rapid since the Late Miocene throughout the Andes (Benjamin et al., 1987; Sébrier et al., 1988; Laubacher and Naeser, 1994; Hoorn et al., 1995; Spikings et al., 2000; Gregory-Wodzicki, 2000). However, the rate of deformation appears to be too small and the timing is inconsistent with simple crustal shortening to explain the observed uplift and crustal thickening.

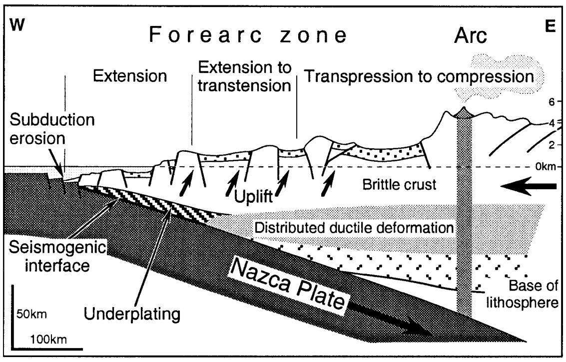

Fig. 3. Geodynamic setting and deformation styles of the forearc zone of northern Chile (after Hartley et al., 2000).

Pardo et al. (14, Fig. 1) study the forces acting upon the downgoing slab below the southern Central Andes (30°S–32°S). They revise the location and focal mechanism of the seismicity, making it possible to determine the stress distribution along the slab. Shallow events located at the interface zone between the oceanic and continental plates are mainly compressional; intermediate earthquakes (60–100 km) are mostly correlated with the bending of the slab which becomes horizontal at depths around 100 km; and deeper events (>100 km) are extensional due to slab pull. They also focused their study upon the causes of damage generated by the 1997 Punitaqui earthquake (central Chile).

The connection between volumes and compositions of volcanic rocks has been proven to be a powerful tool to constrain the deep structure and dynamic evolution of the crust, and the geochemical study of igneous rocks can help to identify pre-Andean crustal domains (e.g., De Silva, 1989; Wörner et al., 1992, 2000; Coira et al., 1993; Aitcheson et al., 1995; Tosdal, 1996; Carlier and Lorand, 1997; James and Sacks, 1999; Guivel et al., 1999; Bock et al., 2000; Lagabrielle et al., 2000). Reviewing geochemical data, Kay et al. (1999) linked the geometry of the subducting slab and the thickness of the Andean crust through time, as well as the location and rate of igneous activity, to the trace element composition of the erupted magmas.

Following these lines, Matteini et al. (13, Fig. 1) studied Late Miocene volcanic suites located in the arc to back-arc areas at 24°S. They correlate trace elements in Miocene volcanic rocks with crustal thickening between the Late Miocene and the present times. In addition, a significant regional difference in crustal composition between the arc and back-arc domains is proposed.

Campos et al. (12, Fig. 1) present data on fluid and glass inclusions from a newly opened Cu–Porphyry copper deposit in Northern Chile. Their data indicate that melts were included into the host quartz crystals at rather high temperature and had very high magmatic copper contents. Based on their observation, they argue for the "orthomagmatic" model of metal enrichment in Andean Cu porphyries.

Due to thrusting of the chain upon the stable south American plate (Lyon-Caen et al., 1985; Giese et al., 1999), back-arc areas are generally characterized by active compressional shortening materialized by east-verging fold and thrust belts (sub-Andean zones) (e.g., Allmendinger et al., 1983; Mégard, 1987; Dengo and Covey, 1993; Roeder and Chamberlain, 1995) although the shortening amount and deformation style are highly variable along the strike (Gil et al., 1999) (Fig. 4).

Husson and Moretti (9, Fig. 1) present a critical review of the factors controlling the heat flow of orogenic belts and their foredeep and compare them to the results of numerical modeling. Then, after presenting the results of the heat flow measurements on the Bolivian sub-Andean zone—which is an important petroleum prospect—they propose that the western high heat flow is due to thickening and internal radiogenic heat, while the eastern low heat flow is related to erosion and fluid circulations within the deformed sedimentary wedge.

Farther east, the evolution of foreland basins is firstly governed by the tectonic development of the orogen, which controls the flexural subsidence (Beaumont, 1981; Jordan, 1981; Cant and Stockmal, 1993; Miall, 1995) and the sediment supply provided by the erosion of the chain (Heller et al., 1988; Flemings and Jordan, 1989). Other factors are large-scale thermal processes (Mitrovica et al., 1989; Pysklywec and Mitrovica, 1998), climate which governs relief dissection, sediment transport and drainage patterns (e.g., Masek et al., 1994), eustatic sea-level changes (Jordan and Flemings, 1991), deformation styles (Houston et al., 2000) or combinations of these factors (Catuneanu et al., 1999). The nature, composition and distribution of sediments of a foreland basin provide information about the nature and location of the source areas and the erosion and transport processes (e.g., De Celles and Hertel, 1989; Potter, 1994). In the same way, the age, thickness and stratal pattern of the sedimentary successions give useful constraints about the deformation affecting the basin and its neighbouring areas (Flemings and Jordan, 1989; Leturmy et al., 2000).

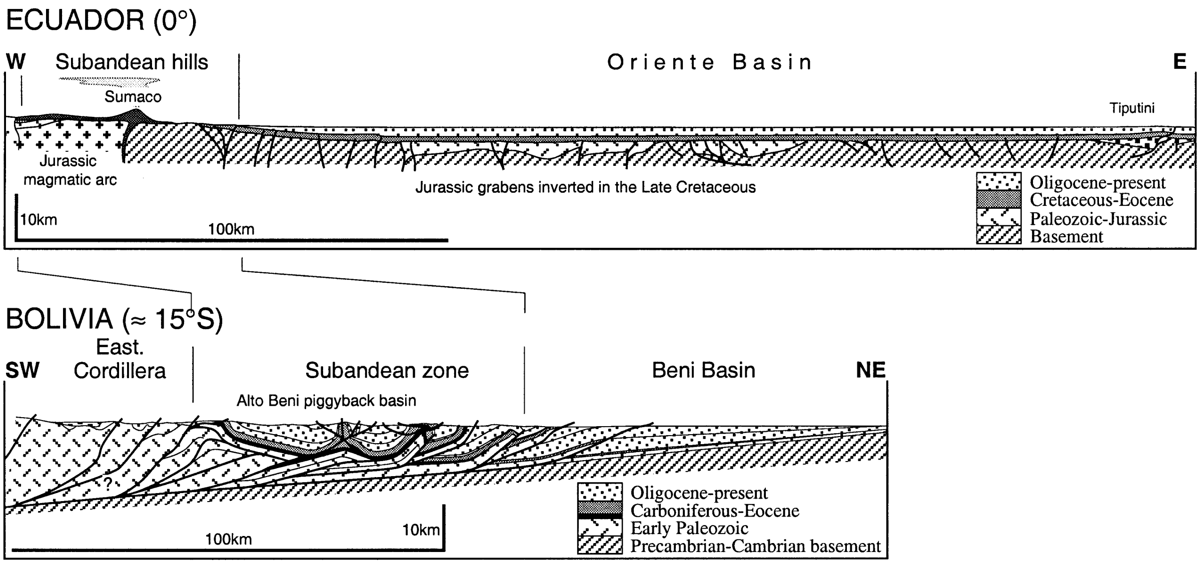

Fig. 4. Simplified cross-sections of the sub-Andean zones and Eastern Basin of Ecuador (after Rivadeneira et al., 1999) and Northern Bolivia (after Baby et al., 1995).

Christophoul et al. present an update of the sedimentary evolution of the Oriente Basin of Ecuador since the Eocene (5, Fig. 1). They correlate the main earliest Eocene unconformity with an Andean tectonic event inducing renewed erosion and subsidence, whereas the Middle Eocene–Miocene evolution would be controlled by isostatic adjustments due to erosion and unroofing of the Andean chain. This evolution leads them to suggest that the Oriente Basin did not behave as a typical foreland basin (Fig. 4).

Farther north, the northern Colombian Andes and their transition to the easterly Caribbean ranges of Venezuela (Fig. 1) are characterized by a complex interaction between the grossly E-ward motions of the Pacific and Caribbean oceanic plates, the W-ward motion of the South American plate and the NNE-ward escape of the continental slices of northwestern South America (e.g., Pindell and Barrett, 1990; Freymüller et al., 1993; Taboada et al., 2000). Based on previous works and on the present-day tectonic activity, Audemard and Audemard propose that the Mérida Andes of Western Venezuela (2, Fig. 1) would result from both dextral transpressional motions and from the build-up of a tectonic wedge related to a NW-ward-dipping continental subduction. The latter interpretation thus challenges previous interpretations which assumed a SE-ward continent subduction (De Toni and Kellogg, 1993; Colletta et al., 1997; Hervouët et al., 2001).

Wenzens (16, Fig. 1) reports the interpretations of ages and extension of glacial deposits from the Patagonian Ice Sheet in order to derive information on the existence and timing of the last major glacial advances of the Last Glacial Maximum (LGM). There are conflicting results and interpretations, and the present contribution represents one view in this dispute. Wenzens maintains that the glacial outflow is much controlled by Neogene tectonic processes, which created the underlying morphology. He also argues that LGM is related to higher precipitation as the result of the effect of the westerlies and that the final period of ice advance lasted until 9500 ^{14}C years BP.

3. Conclusions

The Altiplano in the Central Andes, second highest and largest high plateau in the world, is a first-order feature and received much attention in the last 20 years, while the Colombian and Patagonian Andes have been much less studied (e.g., Fig. 1). Available data suggest, however, that the structure and evolution of the Andean belt are rather variable in time and space although the chain is morphologically continuous from northern Colombia to southern Chile. This volume combines papers presented during the 4th ISAG meeting, which deal with various parts of the Andes and bring new insights on two sections of the chain and allow a more general perspective and comparison.

In the northern Andes, tectonic shortening is limited. Present-day data suggest that the tectonic evolution and crustal thickening are dominated by the accretion and underplating of oceanic material (Bosch et al., Hughes and Pilatasig and Weber et al.) and by transpressional tectonics (Audemard and Audemard) locally guided by former sutures (Hughes and Pilatasig). Therefore, as reflected by the Tertiary evolution of the Oriente Basin (Christophoul et al.), the relief of this part of the chain may result from the isostatic reaction subsequent to the underplating of oceanic material and to the growth of giant flower structures rather than to normal trench shortening.

Such shortening of the continental crust is in fact dominating the build-up of the Andes of Peru, Bolivia and northern Chile although the underplating of continental material dragged down by subduction erosion and the thickening of the ductile lower crust or lithospheric mantle may have played significant roles. The pre-Andean state of the heterogeneous continental crust is being progressively revealed by studying the geochemical signature of the arc magmatism (e.g., Matteini et al.) and the Paleozoic tectonic evolution (Thomas et al.) that allows us to distinguish better the respective parts of the Paleozoic and Andean deformations (Jacobshagen et al.). Additionally, this complex history is shown to noticeably influence the Mesozoic evolution of the margin (Haederle and Atherton and Sempere et al.) and finally its Andean deformation (Sempere et al.). The topographic chain is presently submitted to active compression (Pardo et al.) and subduction erosion to the west. This implies a steepening of the western slope, which would induces gravitational deformations (Wörner et al.), and an active eastward thrusting of the thickened belt inducing thermal anomalies in the foreland basin (Husson and Moretti).

Significant advances have been made recently in our knowledge of the Andes build-up through the exploration of the present-day structure and inferred processes by means of the modelling of "indirect" geophysical, geochemical or petrological data. Validation of these conclusions, however, requires their comparison with well-controlled local and regional studies derived from all fields of geosciences and Andean regions. Our wish is that the present set of contributions, the forthcoming editions of the International Symposium on Andean Geodynamics (ISAG) and other opened scientific meetings will contribute to the exchange and confrontation of complementary results and to the conception of multi-disciplinary projects.

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