Tectonic history of the Andes and Subandean zones: Implications for the development of the Amazon drainage basin

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Tectonic history of the Andes and sub-Andean zones: implications for the development of the Amazon drainage basin

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Abstract

The Andes and the Amazon River have been neighbouring geographical and geological features for at least the past 10 million years. However, the nature of the interactions between them remains unclear. The western margin of South America has been convergent since ~100 Ma, but only during the last 30 million years has there been an adjacent subduction orogen of the extent observed today. Instead, the configuration of the Amazon River evolved from 11 Ma and has remained largely unchanged at least for the past 6 million years. In this chapter we review the available data on the history of deformation, palaeoelevation and exhumation of the northern Central Andes, Northern Andes and adjacent sub-Andean basins in order to compare these data sets with the evolution of the Amazon drainage basin. The available data are far too scarce to propose definitive patterns, but do allow us to pose testable hypotheses on the interaction between the Andes, evolution of sub-Andean zones and the Amazon River. Deformation in the Andes began prior to the establishment of the modern Amazon drainage network and patterns. Although the modern Amazon is very young it appears to be closely related to the development of the Andes. This interrelated history of Amazon River and Andes is inferred from the acceleration in the denudation rates of the Eastern Cordillera, which coincides with the moment that Andean palaeoelevations became significant and began to constitute an orographic barrier and trap to moisture-bearing winds. However, this acceleration could also be related to the development of a denser drainage network in the Andean headwaters. All these factors, together with the presence of orogen-perpendicular basement highs, may have prompted a greater and more focused water and sediment influx towards the Amazon lowlands, producing a river directed to its present-day delta plains in the Atlantic Ocean. As previously proposed, the synchronous development of intense deformation in the sub-Andean basins appears to be related to changing mechanical conditions in the foreland sedimentary wedge that prompted deformation to migrate to the lowlands.

Amazonia, Landscape and Species Evolution: A Look into the Past, 1st edition. Edited by C. Hoorn and F.P. Wesselingh.

Introduction

The Amazon drainage basin occupies an area of about 7 million km², including portions of the territories of Colombia, Ecuador, Peru, Bolivia and Brazil. The Amazon drainage basin is the largest on Earth (Fig. 4.1) and is located adjacent to the Andean orogen, one of the world's largest mountain chains. Most of the rivers feeding the Amazon are sourced in the Andes (see Fig. 4.1). It is therefore expected that the Andes and Amazon interact spatially and temporally in some way (see Fig. 4.1).

The relationship between Andean orogenesis and development of the modern Amazon drainage system is poorly understood. Dobson *et al.* (2001) argue that the modern transcontinental Amazon drainage was not established until the Late Miocene. This raises the question why, in the light of the extended deformation history in the Andes, this major river system was established so recently. In reply, Hoorn (2006) proposed that the rise of the Andes mountains is the critical factor for the origin of the present-day Amazon River.

Recently new data were generated on the orogenic processes in the Andean segments most closely related with the evolution of the Amazon sedimentary and hydrographic basin, the northern Central Andes (Oncken *et al.* 2006) and Northern Andes (Pindell *et al.* 1998; Cediel *et al.* 2003). However, there is no published comprehensive study on the relation between these data and the development of the Amazon drainage basin. Therefore, in this chapter we compare the timing of compressional deformation, surface uplift and exhumation in the Andean orogen with the timing of evolution of the Amazon drainage system. We find that the timing of the onset of compressional deformation in the Andes is only locally understood, with numerous limitations in the resolution of the existing chronology. Data that evaluate the timing and rates of exhumation in the Andes (e.g. Fig. 4.2) are also scarce. In addition, the assessment of palaeoelevation of the Andes – and its implications – is difficult, in particular because palaeoelevation is reconstructed from data that are both scarce and variably interpreted. Therefore, rather than a final result, we here propose a working model of the relationships between the Andes and the

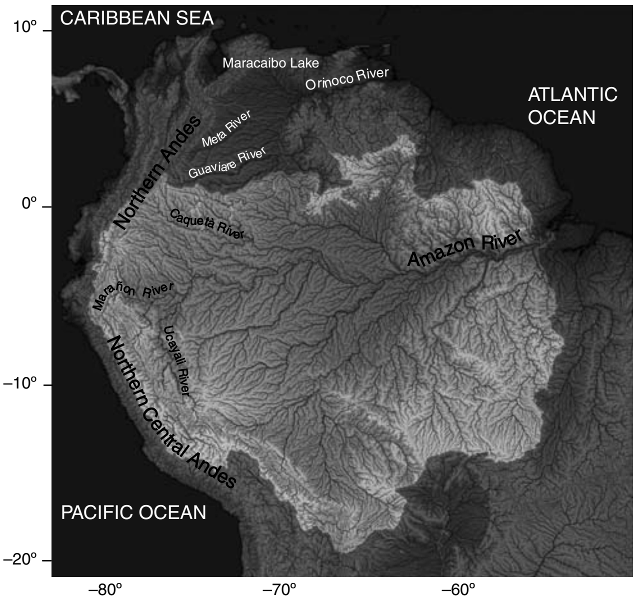

Fig. 4.1 Topographic map of northern South America, with the hydrographic network, including the Amazon (lighter coloured area) and Orinoco hydrographic basins. The Andean orogen is also present with two distinct geological segments – Northern Andes and Central Andes.

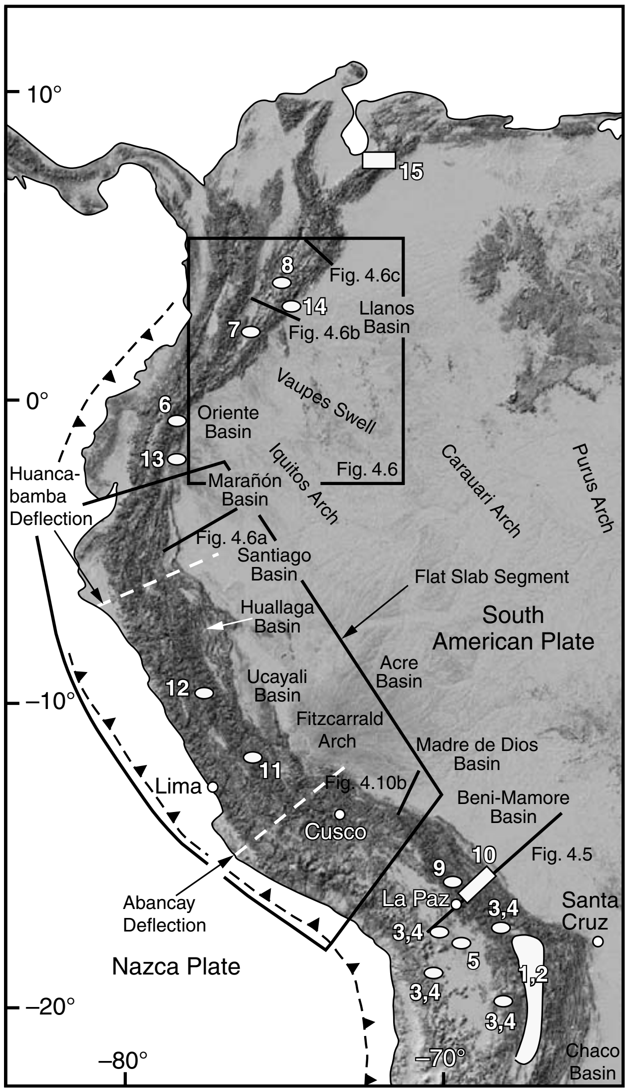

Fig. 4.2 Shaded relief map of the Central and Northern Andes. Numbers in the figure denote different studies related to palaeoelevation: 1. Kennan *et al.* (1997); 2. Barke & Lamb (2006); 3. Gregory-Wodzicki *et al.* (1998); 4. Graham *et al.* (2001); 5. Garzione *et al.* (2006); 6. Burgos (2006); 7. Fields (1959); 8. Wijninga (1996). Studies relating to exhumation: 9. Benjamin *et al.* (1987); 10. Barnes *et al.* (2006); 11. Laubacher & Naeser (1994); 12. Garver *et al.* (2005); 13. Spikings *et al.* (2000); 14. Mora *et al.* (2008); 15. Kohn *et al.* (1984). Notice also the different names of the most important sub-Andean basins.

Amazon River, one that must be verified with more data from the Amazon drainage basin and Amazon headwaters.

This chapter also includes data from areas that were previously connected with the Amazon drainage basin, but are not part of that basin today (see Fig. 4.2). For instance, the Colombian Eastern Cordillera is partly adjacent to the Llanos foreland and partly adjacent to the North Amazonian foreland. However, the Llanos foreland was part of the palaeo-Amazon foreland until at least the Late Miocene (Hoorn *et al.* 1995; Hoorn 2006; Albert *et al.* 2006; see also Chapters 8 & 26) and therefore many of the processes that control the presence of a major river basin east of the Andes can be potentially understood in this area.

In this compilation, first we review the available data on deformation, surface uplift and exhumation of the Central and Northern Andes (see Fig. 4.2). Then we explain in detail the location, role and evolution of a basement arch, the Vaupés Swell, which we consider as an instrumental feature that bounds the present-day Amazon drainage basin to the north and probably also controlled in part the Late Miocene deviation of the Amazon River towards the Atlantic. We then integrate and discuss the data sets to propose that intense precipitation in sediment source areas, focused by an extensive, high-elevation mountain range that intercepts moisture-laden air masses, is the reason why so many drainage basins coincide in a single river system. We consider that intense exhumation via erosion is the main factor leading to an overfilled Amazon drainage basin with most of the rivers flowing to the east. We also highlight the role of basement arches in bounding the Amazon drainage basin. The partial temporal correlation between these variables and the craton-directed growth of the sub-Andean ranges is also explored.

As in the rest of this volume, this chapter is intended for all scientists with an interest in the development of Amazonia trough time. However, it is written by geologists using very specific geological terms. Thus, to make this manuscript understandable for a wider spectrum of readers, in Table 4.1 and Fig. 4.3 we define the most important geological concepts used here.

The history of the Amazon River in a regional geological context

Hoorn *et al.* (1995) proposed that around 16 Ma there was no transcontinental Amazon River system with connection to the Atlantic Ocean. Instead, multiple lines of evidence suggest that during the Middle Miocene the present Amazon drainage basin constituted an extensive lacustrine basin, sporadically invaded by short-lived marine incursions. Within this wetland, there is a record of rivers flowing from the Andes since at least the establishment of this extensive lacustrine regime during the late Early to Middle Miocene (Hoorn *et al.* 1995; Hoorn 2006; see also Chapter 8). Evidence for this is gleaned from the lateral distribution of depositional systems (see Chapter 5), the presence of Andean schist fragments in the Middle Miocene sediments and pollen grains typical of mountainous areas (Hoorn *et al.* 1995), and the absence of Andean clastic sediments along the Atlantic margin (Dobson *et al.* 1997, 2001). Analysis of growth bands from mollusc shells indicates that a precipitation regime similar to Present prevailed in the Amazon foreland basin (Kaandorp *et al.* 2005). Based on pollen analysis by Hoorn *et al.* (1995), it was noted that at that time biodiversity in the Amazon wetland was similar to, or even greater than today (Hooghiemstra & Van der Hammen 1998; Van der Hammen & Hooghiemstra 2000). Greater biodiversity may be an indicator of higher or similar temperatures and precipitation at that time. However, temporal and spatial resolution of palaeoprecipitation data is not yet enough to assess the presence or absence of orographic rain shadows in relation to the Andes. Damuth & Kumar (1975) hypothesize that the Late Miocene onset of deposition in the Amazon Fan is due to the origination of the transcontinental palaeo-Amazon River, although it was unclear whether at that time the river was similar in size and drainage network to the river we observe today. Data from offshore Brazil show that sediments unambiguously derived from the Andes began accumulating on the Brazilian coast by 10 Ma (Dobson *et al.* 1997, 2001). The transcontinental palaeo-Amazon river thus initiated in the Late Miocene, as suggested by Damuth & Kumar (1975), but the modern Amazonian network, namely an unambiguously documented similar distribution of depositional systems and flow patterns, is no older than the Pliocene (Roddaz *et al.* 2005; Espurt *et al.* 2007; Figueiredo *et al.* 2009).

The Andes and their relationship with the plate tectonic context

The western margin of South America (see Fig. 4.2) has been characterized by subduction (i.e. the downward movement of the oceanic plate below the South American plate) for the past 100 Ma (Sdrolias & Müller 2006; Cobbold *et al.* 2007). The most notable change in the subduction regime during the Cenozoic was the break-up of the Farallon Plate into the Nazca and Cocos plates at ~23.4 Ma (Sdrolias & Müller 2006; Cobbold *et al.* 2007). Given this simple tectonic setting, an unresolved question is why significant orogenesis occurred during the late Cenozoic. It has been suggested that variations in the convergence rate between the Nazca oceanic plate and South American plates could be the main factor (see Fig. 4.2). Pindell *et al.* (1998) and more recently Oncken *et al.* (2006) and Babeyko *et al.* (2006) present convincing arguments about the major role played by the velocity of the westward drift of South America along the Northern (Pindell *et al.* 1998), Central and Southern Andes (Oncken *et al.* 2006) (see Fig. 4.1). In this context, Oncken *et al.* (2006) found a positive correlation in the Central Andes between the onset of shortening in the western margin of the upper plate (at ~48 Ma), and a positive difference between South American drift and the subducting slab rollback rate (recorded also since ~48 Ma). They define the onset of such a positive difference as the shortening threshold in the upper South American plate. A similar idea had been proposed by Silver *et al.* (1998), based on qualitative assumptions. Pindell *et al.* (1998) suggested that in the Northern Andes an additional factor is the arrival of the Panamá Arch to the western South American margin.

However, it has been recognized that only after that onset of shortening, at ~30 Ma, can generalized mountain building with similar geographical extension as the present-day Andes

Table 4.1 Definitions of some key geological terms.

Term	Definition
Stresses	Applied forces that deform rock units constituting the Earth's crust (uppermost portion of the Earth). They are in the simplest case horizontal and perpendicular to the anisotropies (earth discontinuities) in the rock units affected by stresses
Tensional stresses	Stresses that tend to expand particles or rock units
Compressional stresses	Stresses that tend to shorten the rock units (Marrett & Peacock 1999)
Transpressional stresses	Stresses oblique to the anisotropies in the rock units or earth discontinuities are termed transpressional when the main component is compressive (Sanderson & Marchini 1984)
Transtensional stresses	Stresses oblique to the anisotropies in the rock units or earth discontinuities are termed transtensional when the main component is tensile (Sanderson & Marchini 1984)
Contractional deformation or shortening	A permanent state of rock units prompted by the horizontal component of movement of those units and caused by compressional far-field or local stresses. For example, in Fig. 4.3 it can be seen that a rock mass with stratified units in an undeformed state (above) can have a horizontal component of movement (below), which caused the rear part of the rock units to be displaced horizontally to the left side of the cartoon. This component is dominant in those parts where deformation occurs along sub-horizontal faults
Thin-skinned tectonics	A style of deformation in contractional thrustbelts, and in some cases used in extensional settings, where rock units from the sedimentary cover are detached along a basal décollement from the underlying older and more rigid basement units
Thick-skinned tectonics	A style of deformation in contractional thrustbelts and in some cases used in extensional settings, where rock units from the older crystalline basement and the overlying sedimentary cover are involved
Rock uplift	Vertical movement of rocks with respect to the geoid (England & Molnar 1990)
Surface uplift	Vertical movement of the surface with respect to the geoid (England & Molnar 1990; see also Fig. 4.3)
Exhumation	Vertical movement of rocks with respect to the surface. Erosional denudation means, in most cases, exhumation (England & Molnar 1990), except in the case of tectonic denudation in normal faults
Orogenesis	All the aforementioned processes and forces can, under certain conditions, create mountains and therefore, if there is a widespread process of mountain building, the term orogenesis is used to define all the group of processes involved
Stream power	The energy that a river or water flow needs to be able to transport sediments; it depends on water discharge and surface slope (Whipple & Tucker 1999). The main factor controlling water discharge in tropical areas is rainfall
Orographic rainfall	In the Amazon basin most of the rainfall at its western side is caused by the orographic effect created by the Andes Mountains (Bookhagen & Strecker 2008). In this case, above certain elevations, the eastern side of the Andes focuses rainfall by retention of moisture-bearing winds on the eastern slopes of the mountain range (e.g. Mora *et al.* 2008)

be clearly documented, both in the Central Andes (see Fig. 4.1) (Allmendinger *et al.* 1997; Oncken *et al.* 2006) and in the Northern Andes (see Fig. 4.1; Gómez *et al.* 2001, 2005; Toro *et al.* 2004; Parra *et al.* 2005, 2009b; Bayona *et al.* 2008). The coincidence in the onset of deformation for the Central and Northern Andes is significant but is also expected. If a subduction margin is present west of the South American plate and this factor controls mountain building, a synchronous onset of deformation along the entire margin is likely. Allmendinger *et al.* (1997) proposed that the reorganization of Pacific plates at ~25 Ma may have caused generalized mountain building in the Central Andes. Babeyko *et al.* (2006) suggested that the velocity of the westward drift of South America was sufficiently high by ~25 Ma to trigger extensive deformation of the strong Andean lithosphere. For the Northern Andes, Pindell *et al.* (1998) even suggest that from the deduced flow lines of convergence between the Caribbean and South American plates, the relative convergence rate between the two plates doubled by ~25 Ma. Based on this Pindell *et al.* (1998) predicted that, compared with previous periods, the Andean orogeny should have been particularly intense in the Northern Andes during the Late Oligocene.

The previous discussion does not imply that structural styles do not need to change along strike. They are also due to other features such as subduction angle or different palaeogeographic and crustal features in the upper plate. For instance, there are two flat slab segments in the subducting Nazca Plate (see Fig. 4.2). One of them in the southern Central Andes (Jordan *et al.* 1997), and the other in the northern Central Andes, the latter aged at ~10 Ma (see Fig. 4.2; Gutscher *et al.* 1999; Hampel 2002). Both of them have been proposed to influence the deformation of the Andean

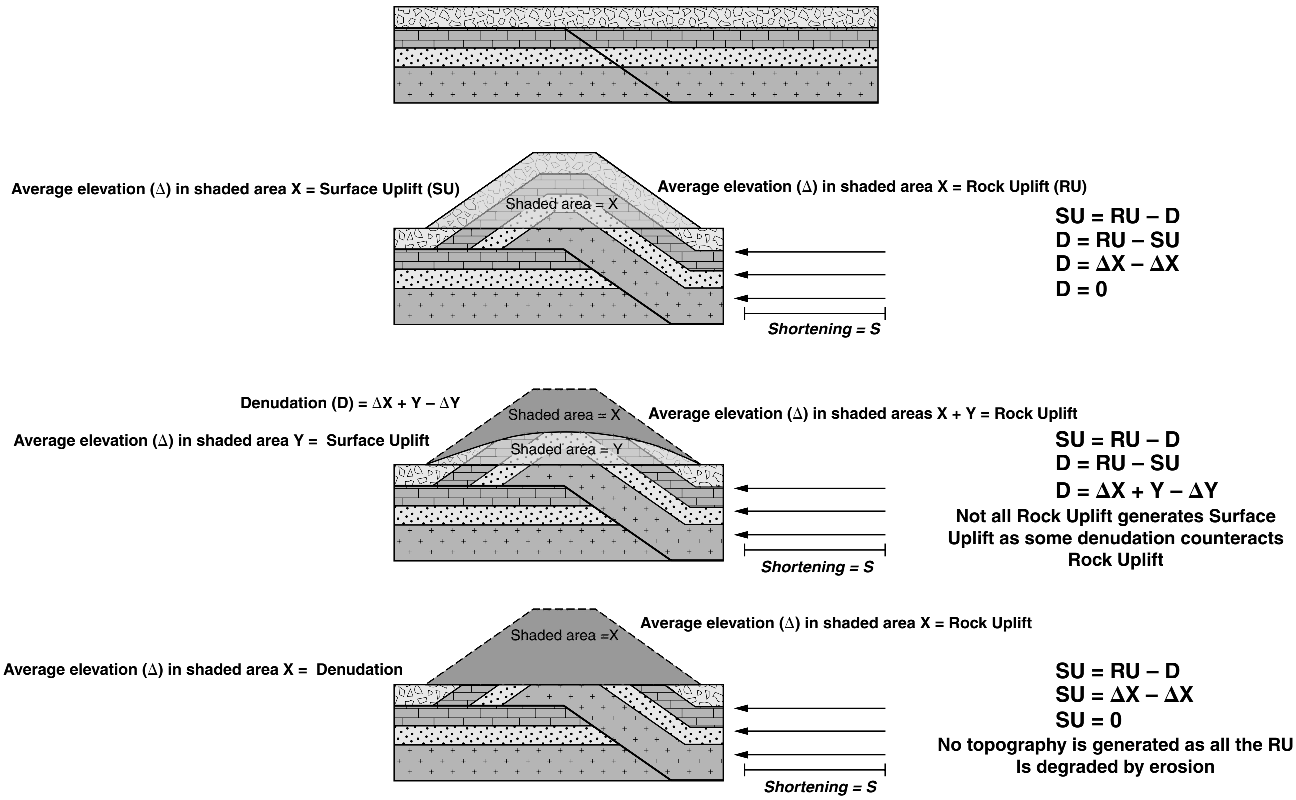

Fig. 4.3 Definition of the main geological concepts related to the deformation and erosion of the materials constituting the Earth's crust.

foreland fold and thrust belts. However, as demonstrated by Kley *et al.* (1999) there is only a partial correlation between the segmentation of the Andean foreland fold and thrust belt and the geometry of the subducted Nazca Plate as initially proposed by Jordan *et al.* (1983). More convincing data on the coincidence in time and space between the subduction of the Nazca ridge and the Pliocene uplift of the Fitzcarrald Arch in the Peruvian sub-Andean zone (see Fig. 4.2) has been shown by Espurt *et al.* (2007). This broad low-elevation basement arch segments the Peruvian sub-Andean zone (see Fig. 4.2).

Cenozoic shortening history in the Central and Northern Andes

Northern Bolivian and Southern Peruvian Andes

Although compressional deformation events, starting in the Paleocene, have been documented elsewhere in the Andes, we focus on events that are synchronous with the Neogene origin of the Amazon River (Hoorn *et al.* 1995; Hoorn 2006). A recent synthesis of published deformation data suggests that in the Central Andes of Bolivia and southern Peru, deformation began at ~47 Ma in the pre-Cordillera (the Andean ranges closer to the Pacific coast; Fig. 4.4, Table 4.2). Deformation propagated eastwards, reaching the Eastern Cordillera first at ~40 Ma (Fig. 4.5; see also Figs 4.2 & 4.4) and virtually the entire Altiplano and Eastern Cordillera at ~30 Ma, with higher shortening rates in the Bolivian Eastern Cordillera (Oncken *et al.* 2006; see Figs 4.2, 4.4 & 4.5). Interestingly, deformation only reached the sub-Andean zones (see Fig. 4.2) at ~10 Ma (Kley 1996; Echavarria *et al.* 2003; Horton 2005). Baby *et al.* (1997) and Rochat *et al.* (1999) calculate total shortening of the Eastern Cordillera and the Bolivian sub-Andean zone from crustal balanced cross-sections (see Fig. 4.5). Total shortening varies from 180 to 230 km, and the Altiplano is interpreted as a piggyback basin carried on the crustal duplex of the Eastern Cordillera, linked with the underthrusting of the Brazilian Shield (see Fig. 4.5).

Northern Peruvian Andes and Ecuadorian Andes

In the Peruvian Andes, pre-Eocene deformation events have been grouped under the term 'Peruvian phase'. Although the spatial distribution of this deformation phase is unclear, this shortening event affected the Peruvian Eastern Cordillera but not the sub-Andean zone (Megard 1984; see Figs 4.2 & 4.4). An important exception is the Peruvian sub-Andean Santiago Basin (see Figs 4.2 & 4.4); seismic reflection sections show that this basin (Figs 4.6a & 4.7; see also Figs 4.2 & 4.4) was deformed by transpressive tectonic inversion of an Upper Triassic–Lower Jurassic rift, starting in the Late Cretaceous with low-amplitude/low-elevation structures rarely creating relief or substantial shortening (Navarro

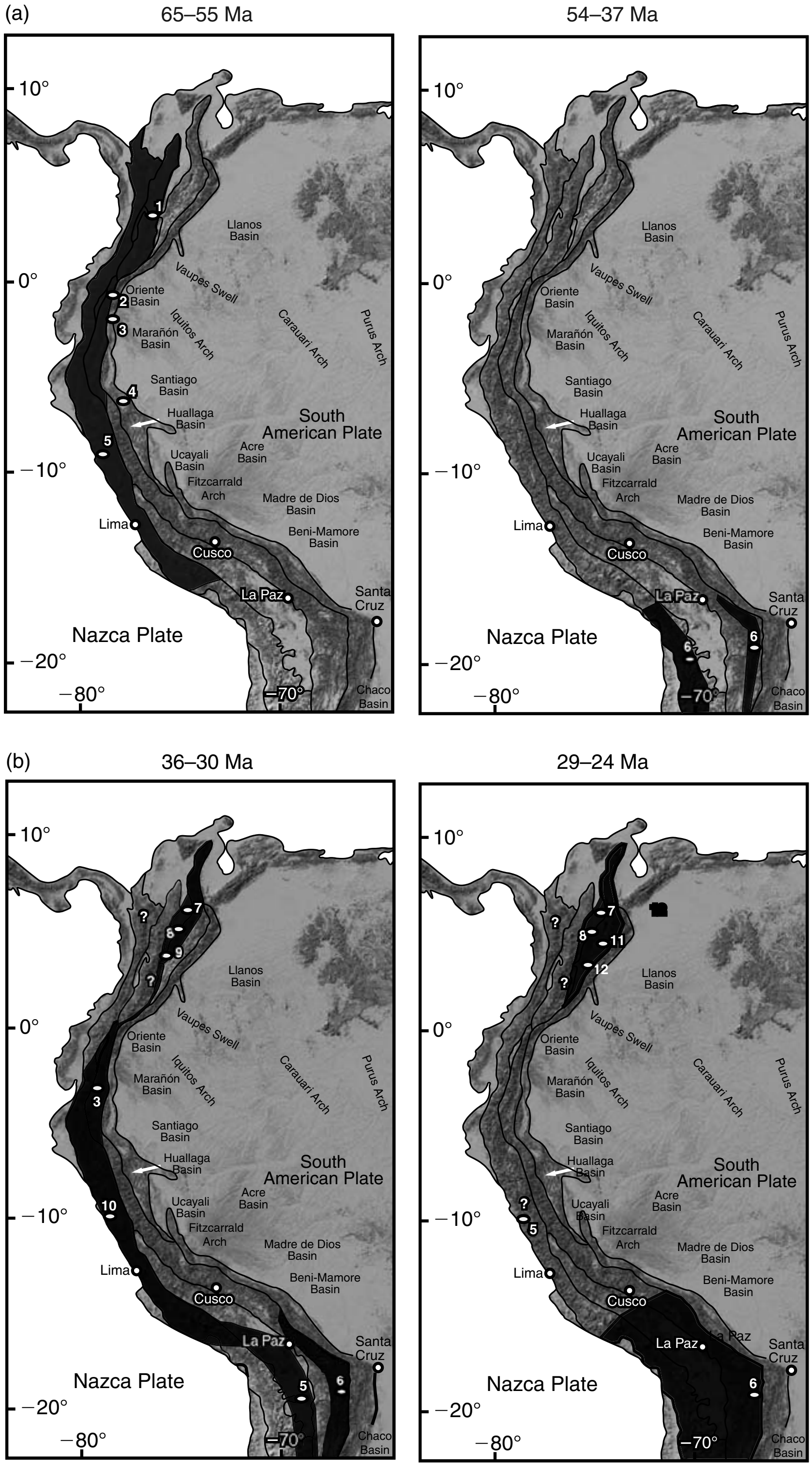

Fig. 4.4 (a–c) Summary of the different stages of deformation, exhumation and increased palaeoelevation in the Andes based on Oncken *et al.* (2006), the sources mentioned in the text and the sources shown in the maps. The black lines separate from west to east the Andean Western Cordilleras and pre-Cordilleras; the Altiplano-Puna Plateau domain and Andean Eastern Cordilleras, only separated in Colombia into the Central and Eastern Cordillera. The easternmost domain surrounded by black lines coincides with the sub-Andean zones. Notice that the deformation domains do not always coincide with the different divisions of the Andes. White dots are locations with documented deformation events. Black dots are areas with palaeoelevation data. Black dots with a white envelope are areas with documented exhumation rates from thermochronolgy. Each number is a different source. See a summary of data sources in Table 4.2 with numbers corresponding to the numbers used in this figure.

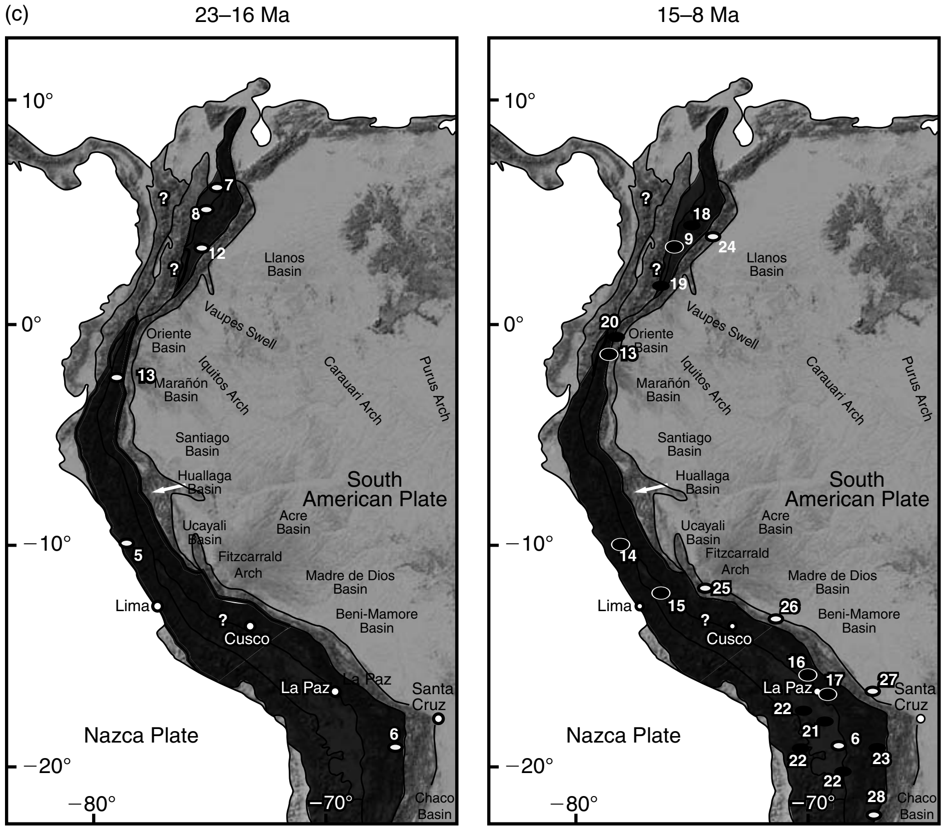

Fig. 4.4 *Continued.*

et al. 2005). This Late Cretaceous deformation is also well known in the Oriente Basin of Ecuador (see Figs 4.2 & 4.4), where it is associated with alkalic igneous activity (Barragan *et al.* 2005). The Early Eocene period in Peru is considered to be a time of tectonic quiescence and erosion of the Andes to the west of the sub-Andean zone (Hermoza *et al.* 2005; see Chapter 5). The Middle to Late Eocene Incaic phase has been regarded as the main Andean phase in Peru, at least in the Western Cordillera (see Figs 4.2 & 4.4) and adjacent Altiplano (Mégard 1984). In the Western Cordillera of Peru, pre-Late Eocene structures are eroded and unconformably overlain by ~40 million year-old clastic deposits (Noble *et al.* 1974, 1979), but structures of this age were not documented in the Eastern Cordillera (see Figs 4.2 & 4.4). Christophoul *et al.* (2002) describe Middle Eocene low-elevation/low-amplitude contractional structures (tectonic inversion) in the Oriente Basin (see Figs 4.2 & 4.4) of Ecuador.

The Quechua phases would have affected the Peruvian Western Cordillera, Altiplano and Eastern Cordillera (see Figs 4.2 & 4.4) during the Late Oligocene and Miocene (Mégard 1984). However, it is unclear if these corresponded to a mostly continuous deformation, affecting different segments of the Peruvian Andes (like in the Northern Andes of Colombia, see below), or whether they can really be separated into discrete periods. Recent studies (Hermoza *et al.* 2005; Roddaz *et al.* 2005; Espurt *et al.* 2007) show a continuous Cenozoic eastward propagation of the Peruvian retro-foreland basin system from the Western Cordillera to the present sub-Andean zone (see Fig. 4.4). Laubacher & Naeser (1994) argue that the first signal of deformation, as reflected in AFT (apatite fission tracks) cooling ages from granitic rocks in the Peruvian Eastern Cordillera, occurred during the Early Miocene (see Fig. 4.4). However, this claim needs to be judged with caution as it is based on only three samples and no track length data. In contrast, a conformable deposition in the sub-Andean basins of Peru at that time (Early Miocene?) shows no evidence of Middle Eocene to Middle Miocene shortening affecting this structural province of the Peruvian Andes (Gil 2001; Hermoza *et al.* 2005). As will be explained below, deformation only reached the sub-Andean zone during the Late Miocene.

Northern Andes of Colombia

Pre-Eocene deformation episodes in the Northern Andes of Colombia have been documented in the Central Cordillera (Fig. 4.8; Gómez *et al.* 2005), and even older deformation was recorded in the Western Cordillera (see Figs 4.4 & 4.8; Cediel *et al.* 2003). These events have also affected the area of the present-day Colombian Eastern Cordillera (see Figs 4.4 & 4.8; Bayona *et al.* 2006b), but

Table 4.2 Summary of data sources used in Fig. 4.4; numbers in the left-hand column correspond to the numbers used in the figure.

Number	Reference	Location	Type of data	Interpreted process	Timing
Late Cretaceous-Early Tertiary					
1	Gómez *et al.* (2005)	Central Cordillera (Colombia)	Thermochron/basin modelling	Deformation/exhumation	Late Cretaceous
2	Barragan *et al.* (2005)	Oriente Basin (Ecuador)	Documentation of contractional tectonics	Deformation	Late Cretaceous
3	Spikings *et al.* (2000)	Cordillera Real (Ecuador)	Low-temperature thermochron	Exhumation	Late Cretaceous/Paleocene
4	Navarro *et al.* (2005)	Santiago Basin (Peru)	Documentation of contractional tectonics	Deformation	Late Cretaceous
5	Megard (1984)	Eastern Cordillera (Peru)	Documentation of contractional tectonics	Deformation	Pre-Eocene
Late Eocene					
6	Oncken *et al.* (2006)	Pre- and Eastern Cordilleras (Bolivia)	Documentation of contractional tectonics	Deformation	Eocene (starting at ~47 Ma)
7	Gómez *et al.* (2001)	Western flank of the Eastern Cordillera (Colombia)	Documentation of contractional tectonics	Deformation	Late Eocene
8	Bayona *et al.* (2008)	Western flank of the Eastern Cordillera (Colombia)	Basin modelling	Deformation	Late Eocene
9	Parra *et al.* (2009b)	Central part of the Eastern Cordillera (Colombia)	Basin modelling	Deformation/exhumation	Late Eocene
10	Noble *et al.* (1974, 1979)	Western Cordillera (Perú)	Documentation of contractional tectonics	Deformation	Late Eocene
Oligocene					
11	Toro *et al.* (2004)	Eastern Cordillera (Colombia)	Documentation of contractional tectonics	Deformation	Oligocene
6	Oncken *et al.* (2006)	Central Andes	Documentation of contractional tectonics	Deformation	Oligocene
12	Parra *et al.* (2009b)	Eastern Cordillera (Colombia)	Thermochron/basin modelling	Deformation/exhumation	Late Oligocene
5	Megard (1984)	Western Cordillera, Altiplano and Eastern Cordillera (Peru)	Documentation of contractional tectonics	Deformation	Late Oligocene
Early Middle Miocene					
7	Gómez *et al.* (2001)	Western flank of the Eastern Cordillera (Colombia)	Thermochron/basin modelling	Deformation/exhumation	Early Miocene
8	Bayona *et al.* (2008)	Eastern Cordillera (Colombia)	Basin modelling	Deformation	Early Miocene

(*Continued*)

Table 4.2 Continued.

Number	Reference	Location	Type of data	Interpreted process	Timing
9	Parra *et al.* (2009b)	Eastern flank of the Eastern Cordillera (Colombia)	Thermochron/basin modelling	Deformation/exhumation	Early Miocene
13	Spikings *et al.* (2000)	Northern Cordillera Real (Ecuador)	Low-temperature thermochron	Exhumation	Middle Miocene
5	Megard (1984)	Western Cordillera, Altiplano and Eastern Cordillero (Peru)	Documentation of contractional tectonics	Deformation	Miocene
6	Oncken *et al.* (2006)	Central Andes	Documentation of contractional tectonics	Deformation	Miocene
Late Miocene					
3	Spikings *et al.* (2000)	Cordillera Real (Ecuador)	Low-temperature thermochron	Deformation/exhumation	Late Miocene
14	Garver *et al.* (2005)	Peruvian Cordillera Blanca	Low-temperature thermochron	Deformation/exhumation	Late Miocene
15	Laubacher & Naeser (1994)	Peruvian Eastern Cordillera	Low-temperature thermochron	Deformation/exhumation	Late Miocene
16	Benjamin *et al.* (1987)	Bolivian Eastern Cordillera	Low-temperature thermochron	Deformation/exhumation	Late Miocene
17	Barnes *et al.* (2006)	Bolivian Eastern Cordillera	Low-temperature thermochron	Deformation/exhumation	Late Miocene
Late Miocene or Late Miocene/Pliocene					
18	Wijninga (1996)	Colombian Eastern Cordillera	Palaeobotany	Increased palaeoelevation	Late Miocene/Pliocene
19	Fields (1959)	Colombian Eastern Cordillera	Palaeobotany	Increased paleoelevation	Late Miocene/Pliocene
20	Burgos (2006)	Cordillera Real (Ecuador)	Palaeosurfaces	Increased palaeoelevation	Pliocene
21	Garzione *et al.* (2006)	Bolivian Altiplano	Stable isotope palaeoaltimetry	Increased palaeoelevation	Late Miocene
22	Graham *et al.* (2001); Gregory-Wodzicki *et al.* (1998)	Bolivian Altiplano	Palaeobotany	Increased palaeoelevation	Late Miocene
23	Barke & Lamb (2006); Kennan *et al*. (1997)	Bolivian Altiplano	Palaeosurfaces	Increased palaeoelevation	Late Miocene
Late Miocene/Pliocene					
24	Mora *et al.* (2008)	Colombian sub-Andes	Low-temperature thermochron	Deformation/exhumation	Late Miocene/Pliocene
25	Espurt *et al.* (2008)	Peruvian sub-Andes	Low-temperature thermochron	Deformation/exhumation	Late Miocene/Pliocene
26	Mora *et al.* (unpublished work)	Peruvian sub-Andes	Low-temperature thermochron	Deformation/exhumation	Late Miocene/Pliocene
27	Baby *et al.* (1997)	Bolivian sub-Andes	Documentation of contractional tectonics	Deformation	Late Miocene/Pliocene
28	Echavarria *et al.* (2003)	Bolivian sub-Andes	Documentation of contractional tectonics	Deformation	Late Miocene/Pliocene

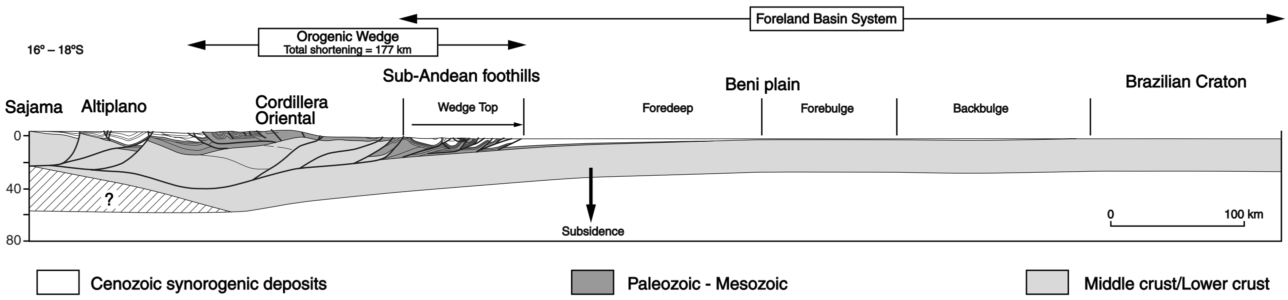

Fig. 4.5 Regional cross-section at the latitude of the Central Andes south of the Bolivian orocline (~16–18°S), showing the main elements constituting the Andean orogen and adjacent foreland (modified after Baby *et al*. 1997). See its location in Fig. 4.2.

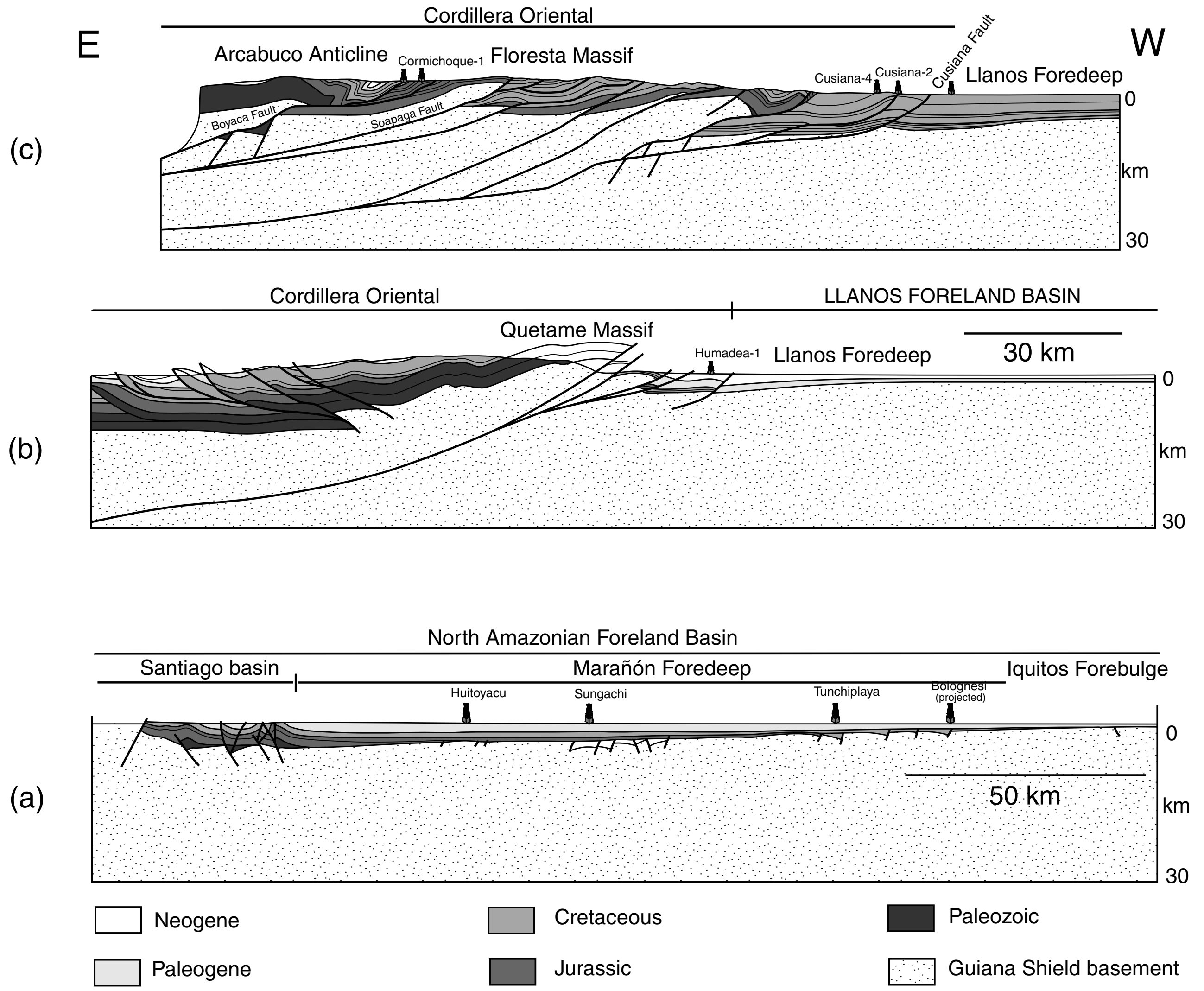

Fig. 4.6 Balanced cross-sections from the Northern Andes, Eastern Cordilleras and sub-Andean basins. **(a)** Geological cross-section along the Santiago Basin (see location in Fig. 4.2; after Gil *et al.* 2001). **(b)** Geological cross-section through the Central Eastern Cordillera of Colombia (see location in Fig. 4.2; modified after Mora *et al.* 2008). **(c)** Geological cross-section through the northern Eastern Cordillera of Colombia (see location in Fig. 4.2; after Toro *et al.* 2004).

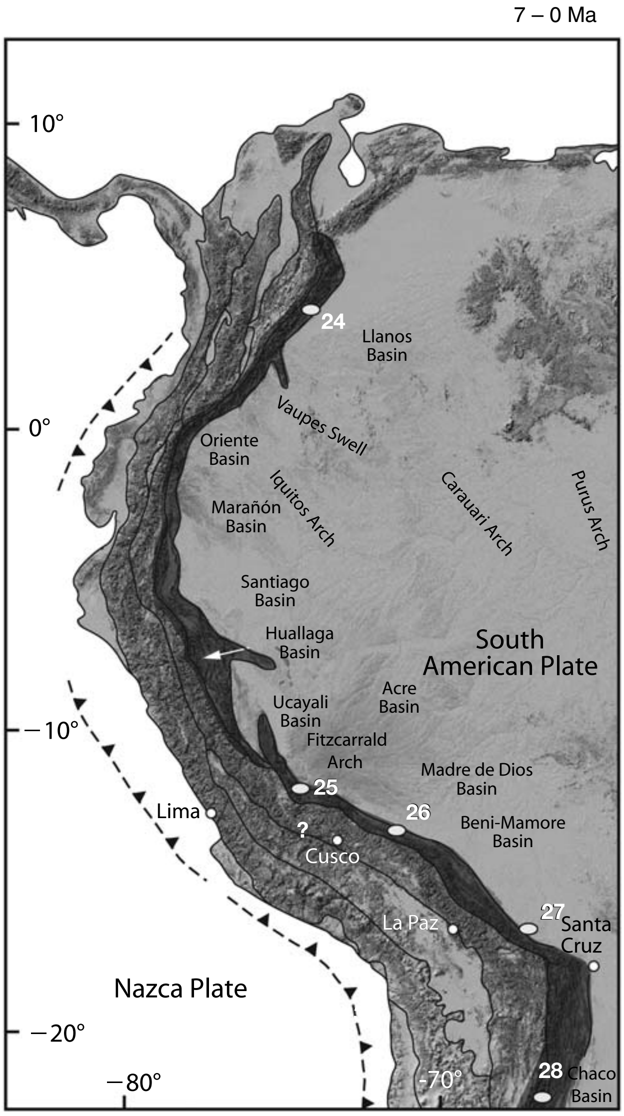

Fig. 4.7 Shaded relief showing the sub-Andean zones (darker grey areas) and the data points supporting a Late Miocene and younger deformation in those areas.

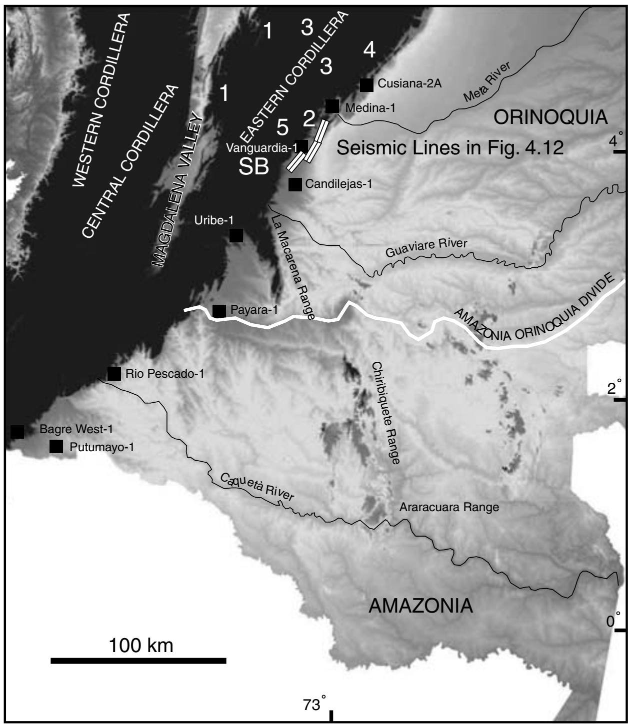

Fig. 4.8 Topographic map (see location in Fig. 4.2) including the main geographical features around the Vaupés Swell, which in fact acts as the drainage divide between the Orinoquía and the Amazonia. Notice the Chiribiquete, Araracuara and Macarena ranges and the position of the Orinoco-Amazon drainage divide. All these elements are defined by the Vaupés Swell low-elevation arch.The numbers indicate the location of different recent geological studies in the Eastern Cordillera: 1. Gómez *et al.* (2003, 2005); 2. Parra *et al.* (2009a, 2009b); 3. Bayona *et al.* (2008); 4. Toro *et al.* (2004); 5. Mora *et al.* (2008). Wells used in Fig. 4.11 are also shown (black squares) and seismic profiles used to make the composite seismic line in Fig. 4.12 (white boxes). SB, indicates the location of the high plain of Bogotá.

deformation rates were so low that in most of the area sedimentation was faster than the vertical rock uplift (Mora *et al.* 2006). Deformation and associated erosional unconformities reached the present western foothills of the Colombian Eastern Cordillera and the Magdalena Valley (see Figs 4.4 & 4.8) before Early–Middle Eocene times (Restrepo-Pace *et al.* 2004), whereas the Central Cordillera was inactive at this time (Gómez *et al.* 2005). However, similar to the Central Andes, only by ~25 Ma was deformation ubiquitously present throughout the Eastern Cordillera. This was documented by growth strata in both the western (Gómez *et al.* 2001, 2005) and eastern foothills of the Eastern Cordillera (see Fig. 4.6; Parra *et al.* 2005, 2009b; Martínez 2006; Bayona *et al.* 2008) and by thermochronology (Toro *et al.* 2004; Parra *et al.* 2009b). Therefore, the Eastern Cordillera has been uplifted as an enormous Laramide-type basement block (e.g. Erslev 1986) since the Late Oligocene (see Figs 4.4, 4.6 & 4.8). Balanced cross-sections show that shortening rates increased dramatically in the Northern Andes by Late Miocene times (Mora *et al.* 2008). In contrast, the very narrow Colombian sub-Andean ranges (Fig. 4.7) underwent most of their shortening since Late Miocene times (Mora *et al.* 2008).

Altogether, a strikingly evident synchronism along the Eastern Cordilleras of the Central and Northern Andes becomes apparent. For instance, the onset of deformation in the Eastern Cordilleras of both Colombia and Bolivia is dated to the Late Eocene. However, there are insufficient data to evaluate the timing of onset of deformation in the Peruvian Eastern Cordillera (see Fig. 4.4). The generalized Late Oligocene to Early Miocene deformation in the Andes, especially along the Andean Eastern Cordilleras, is also remarkable (see Fig. 4.4). Nevertheless, the lack of evidence of significant shortening in the sub-Andean zones prior to Late

Miocene times is also clear (see Figs 4.4 & 4.7). It is worth noting that prior to the Late Miocene there is abundant evidence for deformation, but no evidence for high palaeoelevations along the Central and Northern Andes (see Fig. 4.4).

Surface uplift data in the Andes

Palaeoelevation data from the Central and Northern Andes are essential to understand the evolution of the Amazon drainage system. This is because the amount and timing of surface uplift for a given region directly relate to the stream power on slope and topography and subsequently the palaeo-depositional settings and fluvial networks (e.g. Flemings & Jordan 1989; see Table 4.1). Equally, the presence of laterally extensive terrains with some degree of topographic relief may favour the presence of larger drainage areas, which in turn causes higher erosion rates. However, palaeoelevation data in the Andes are remarkably scarce (e.g. Gregory-Wodzicki 2000), and there is a need for additional palaeoelevation data from throughout the Andes.

In the Central Andes, palaeoelevation was initially inferred based on analysis of palaeo-surfaces (Kennan *et al.* 1997; see also Figs 4.2 & 4.4). These authors proposed a topographic growth of ~2 km since the Late Miocene in the Bolivian Eastern Cordillera, where present-day elevations range from 2000 to 4000 m. More recently, Gregory-Wodzicki *et al.* (1998) and Graham *et al.* (2001) used palaeoflora and leaf physiognomy in the Bolivian Eastern Cordillera and Altiplano to suggest that from one-third to one-half of the surface uplift of the Eastern Cordillera and Altiplano had already occurred by the beginning of the Pliocene (see Fig. 4.2).

In contrast, Garzione *et al.* (2006, 2008) used oxygen isotopic compositions of carbonates to suggest that the Altiplano reached its present-day elevation somewhere between ~10.3 and 6.8 ± 0.4 Ma (see Figs 4.2 & 4.4). Surface uplift was estimated at ~2.5–3.5 km during that time, which is much faster than the surface uplift inferred from previous work. The authors conclude that such an enormous amount of surface uplift in such a short time can only occur if there is removal of dense eclogitic lower crust and mantle lithosphere.

However, as noted by Oncken *et al.* (2006), the data of Garzione *et al.* come from a single section in the Corque syncline (see Fig. 4.2), which is located in the hanging wall of a very precise structure out of many that were involved in the uplift of the Altiplano. Nevertheless, the regional geological context (deformation rates) favours a progressive surface uplift during the Cenozoic deformation (e.g. Baby *et al.* 1997; Allmendinger *et al.* 1997; Horton 1998, 2005; Rochat *et al.* 1999; Victor *et al.* 2004; Oncken *et al.* 2006), rather than a very fast Late Miocene surface uplift, which would require either delamination of lithospheric mantle, or faster shortening rates. Indeed, there is no evidence for peak shortening rates during that specific period of time (e.g. Oncken *et al.* 2006).

Barke & Lamb (2006) (see Figs 4.2 & 4.4) used the morphology of well-preserved regional palaeo-surfaces in the Bolivian Eastern Cordillera to propose that this range was uplifted ~2 km during the last 12 My, reaching the present-day average elevations of 3250 m.

Garver *et al.* (2005) (see Figs 4.2 & 4.4) review the palaeoelevations inferred from palaeo-surfaces northwards, in the Peruvian Eastern Cordillera, and state that it is likely that much of the final surface uplift of the Eastern Andes in northern Peru happened in the last 5–6 My. However, previous stages of surface uplift as old as 15 Ma are also documented.

In Ecuador, Burgos (2006) deduced that much of the surface uplift of the Cordillera Real (see Figs 4.2 & 4.4) has occurred during the last 4 My, based on the comparison of the original and present-day position of the Lower Pliocene Pisiyambo surface.

Northwards, in the Magdalena inter-Andean valley of Colombia (see Figs 4.2, 4.4 & 4.8), west of the Eastern Cordillera (see Fig. 4.8), the so called La Venta fossils (Fields 1959) record mangroves and other fauna typical of very humid climates in the ~15 My-old La Venta Formation. It is likely that this desert, now one of the few arid regions in the Northern Andes, is the result of an orographic rain shadow that prevented moisture-bearing winds from reaching the inter-Andean Magdalena Valley (see Fig. 4.8). This is a similar palaeo-environmental response as was reported by Kleinert & Strecker (2001) in the Argentinian Andes. In other areas in the Andes, elevations of at least 2000 m are required to have an effective orographic barrier limiting the propagation of moisture-bearing winds (Sobel & Strecker 2003; Blisniuk *et al.* 2005). Therefore, we can conclude that at ~15 Ma there was insufficient relief in the adjacent Eastern Cordillera to form such a barrier.

Finally, in the Eastern Cordillera of Colombia, pollen data suggest that the Central High Plain of the Sabana de Bogotà (see Figs 4.2, 4.4 & 4.8) reached its present elevation progressively from ~15 to 3 Ma, with maximum surface uplift rates between 6 and 3 Ma (Wijninga 1996; Van der Hammen *et al.* 1973; Hooghiemstra *et al.* 2006).

In general, the evidence shows that although deformation started as early as 30 Ma and earlier in the Andean Eastern Cordilleras, the available palaeoelevation data allow us to infer that present-day surface elevations over large areas were only reached during or after the Late Miocene.

Exhumation histories

In the Central Andes, one of the first studies on the exhumation of the Bolivian Eastern Cordillera was done by Benjamin *et al.* (1987; see Figs 4.2 & 4.4). Masek *et al.* (1994) and Anders *et al.* (2002) pointed out that the data from Benjamin *et al.* (1987) show increased denudation rates around 10–15 Ma. More recently the AFT data from Barnes *et al.* (2006) confirm accelerated erosion since ~15 Ma, based on individual AFT ages and on modelling of track lengths (see Figs 4.2 & 4.4).

For the Peruvian Eastern Cordillera, to the north, Laubacher & Naeser (1994; see Figs 4.2 & 4.4) deduced two main cooling episodes based on three AFT samples, but without length data. The first occurred during the Late Oligocene to the earliest Miocene, implying erosion of ~2 km of rock. The second cooling episode occurred from the Late Miocene to Present, with denudation of 3–4 km and rates of 0.33 mm/year. In the Peruvian Cordillera Blanca, Garver *et al.* (2005; see Fig. 4.2) produced ZFT (zircon fission tracks) and zircon uranium-thorium/helium (zircon [U-Th]/He;

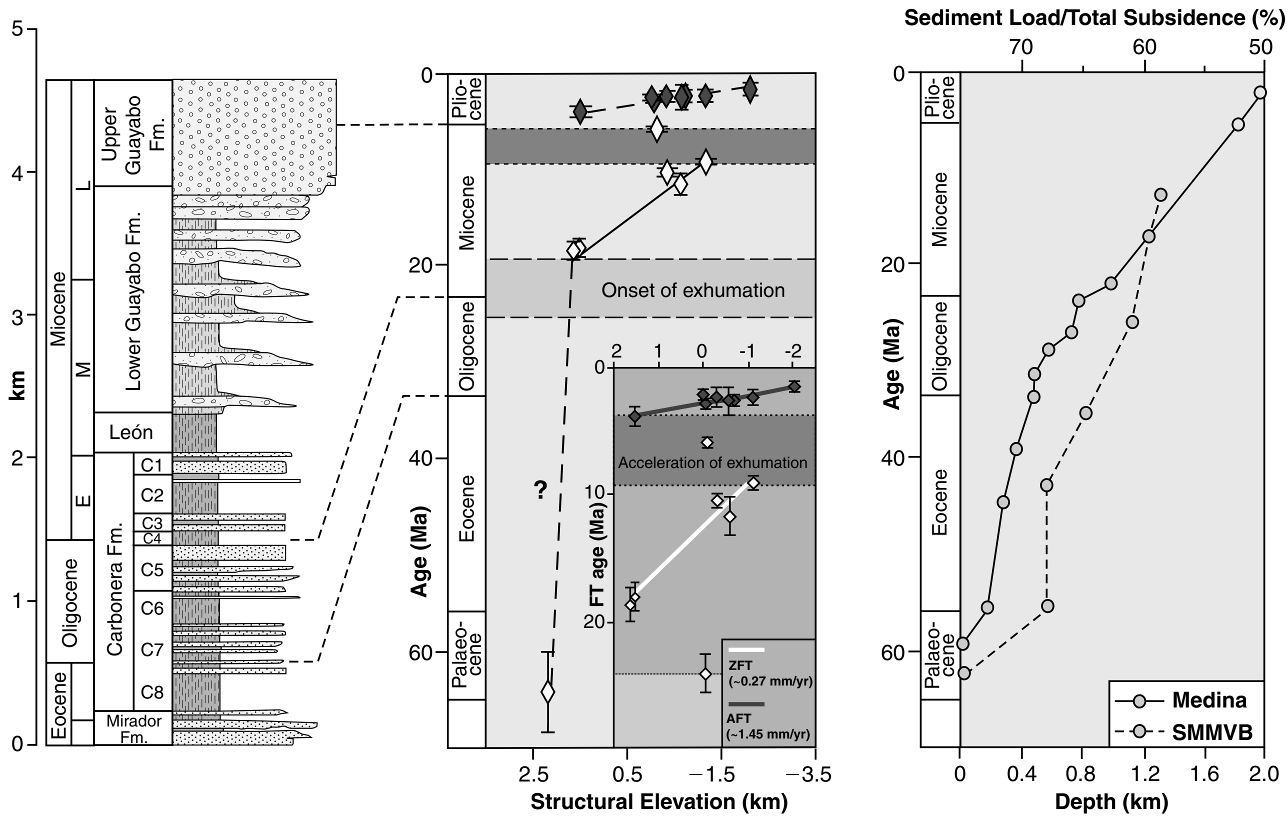

Fig. 4.9 Comparison between exhumation data in a basement massif along the eastern foothills in the Colombian Eastern Cordillera (data after Mora *et al.* 2008; Parra *et al.* 2009b) – see location in Fig. 4.8 in the area number 2 and number 5 – and the subsidence and granulometric changes in the Neogene record of the adjacent foredeep along the eastern foothills (modified after Parra *et al.* 2009a). Exhumation is shown plotted by age versus structural position in order to derive exhumation rates based on zircon fission track data and apatite fission track data from the same basement massif. Notice that the ZFT data show a first period of lower exhumation rates, whereas the AFT data show a period with higher exhumation rates. The foredeep depozone is located in the surroundings of the Medina-1 well (see Medina-1 location in Fig. 4.8 and its regional correlation with other wells in Fig. 4.11). AFT, apatite fission tracks; Fm, Formation; FT, fission tracks; SMMVB, Southern Middle Magdalena Valley Basin; ZFT, zircon fission tracks.

or zircon uranium thorium-helium, ZHe) data suggesting that since ~6.2 Ma nearly 5 km of unroofing occurred along the Cordillera Blanca normal fault.

For the Cordillera Real of the Ecuadorian Andes, Spikings *et al.* (2000; see Figs 4.2 & 4.4) presented ZFT, AFT individual ages and modelling data suggesting more rapid exhumation rates between 10 and 0 Ma compared with the previous 15 Ma.

For the Colombian Eastern Cordillera, Parra *et al.* (2009b; see Figs 4.2, 4.4 & 4.8) demonstrated that denudation along the eastern foothills started at ~22 Ma. Calculated denudation rates during the interval 22 to ~10 Ma average 0.3 mm/year, with a maximum of 0.5 mm/year in a few locations (Parra *et al.* 2009b; Fig. 4.9). However, during the Late Miocene, the data from the Eastern Cordillera show a progressive acceleration until ~4 Ma, when exhumation rates reach values between 1.5 and 2 mm/year (Mora *et al.* 2008) (see Fig. 4.9). Interestingly, Parra *et al.* (2009b) and Mora *et al.* (2008) found a positive correlation in the Colombian sub-Andes between increasing granulometry upsection, and more proximal depositional settings in the foredeep, synchronous with increasing denudation rates in the adjacent Eastern Cordillera (see Fig. 4.9). In contrast, Parra *et al.* (2009b) found that faster subsidence rates, at ~30 Ma in the foredeep, correspond with the onset of deformation in the hinterland, whereas faster exhumation rates at ~6 Ma have no expression in faster subsidence rates (see Fig. 4.9).

In the northernmost Venezuelan Andes, Kohn *et al.* (1984) obtained AFT ages of between 8 and 2 Ma; a few locations yielded Late Oligocene ages. The absence of profiles located in well-defined structures, and problems with inferred upwarping of isotherms make it difficult to estimate denudation rates, or the onset of denudation in this area. However, Pliocene AFT ages in the Mérida Andes partially coincide with Late Miocene-Pliocene acceleration of denudation rates in the Colombian Eastern Cordillera, as reported by Mora *et al.* (2008).

In general, the data above show a Late Miocene acceleration of denudation rates in the eastern face of the Eastern Cordilleras of the Central and Northern Andes. However, it should be noted that there are significantly fewer data from the Peruvian Eastern Cordilleras than in other areas of the Northern Central and Northern Andes. Particularly interesting is that the timing of

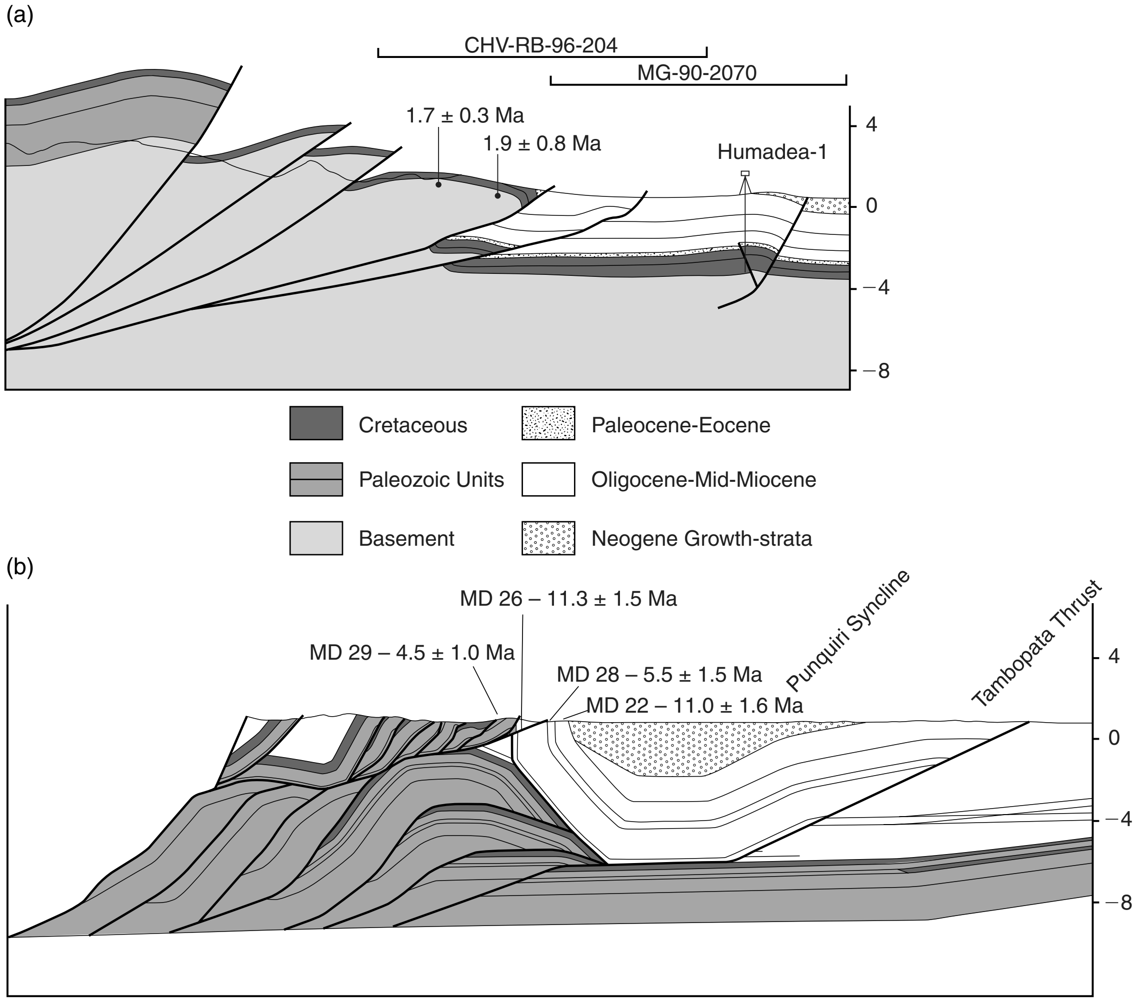

Fig. 4.10 Structural cross-sections from the sub-Andean zones of **(a)** the Llanos Basin (modified after Mora 2007) and **(b)** the Madre de Dios Basin. In the structural cross-section from the Llanos basin, AFT ages near the base of the Cretaceous sequence are shown; vitrinite reflectance data and the much younger AFT ages compared with the stratigraphic age of the units show that those ages are totally reset ages. Mora (2007) considers these ages as indicators of the age of initial thrusting. In (b), the section from the Madre de Dios area, the antiformal stack has totally reset ages of of 11 ± 1.6 Ma close to well-documented growth strata. This age is taken as a proxy for the onset of thrust-induced denudation and close to the age of the antiformal stack. See text for discussion and Fig. 4.2 for cross-section location.

higher denudation rates roughly coincides with coarser grained sedimentation in the sub-Andes (e.g. rivers with higher stream power) and faster sedimentation rates in the sub-Andean foredeeps, if these data are compared with those shown by Roddaz *et al.* in Chapter 5.

Structural styles and timing of the structures in the sub-Andean zones

In Figs 4.5, 4.6 & 4.10 we present cross-sections along the thin-skinned sub-Andean belts from the Benì area northwards through the Madre de Dios, Marañón and Santiago basins (after Baby *et al.* 1997, 1999; Gil *et al.* 2001; Toro *et al.* 2004; Hermoza *et al.* 2005; Roddaz *et al.* 2005; Mora *et al.* 2008) towards the Colombian Llanos foothills to show the structural styles of all these sub-Andean basins (see Figs 4.2 & 4.4).

Along strike structural changes are partially preconditioned by changes in the palaeogeography and thickness distribution of the deformed strata, as well as by the deformation regime and relative orientation of the stress vector with respect to the deformed belt, as noted by Kley *et al.* (1999), Gil *et al.* (2001) and Mora (2007).

The highest reliefs of the Colombian Eastern Cordillera coincide with a wide Early Cretaceous rift zone, and the master faults of this rift system were reactivated during the Cenozoic deformation (see Figs 4.6 & 4.8). In turn, the adjacent narrow sub-Andean belt of northern Colombia is constituted by short-cuts splaying from the master faults and coinciding with an area where apparently much thinner Lower Cretaceous synrift units are present (see Figs 4.6 & 4.8; Mora 2007). The pinch-out of such synrift units preconditions the foreland propagation of the thrust front and the southward extent of the sub-Andean ranges (see Fig. 4.6c). However, when compared with other sub-Andean foothill areas, it appears that the Colombian sub-Andean zone is much narrower and underwent less shortening. In this area, Mora (2007) used borehole breakout data and map patterns to document obliquity between applied stresses and the rift boundary. Based on analogue models, it appears that the obliquity between the inherited rift boundary and the applied stresses (e.g. Marshak *et al.* 1992; Macedo & Marshak, 1999) could be the factor causing such a narrow sub-Andean front.

A close analogue to this situation is the Santiago Basin (see Figs 4.2 & 4.4 for location and Fig. 4.6 for cross-section) in northern Peru, where there is a narrow sub-Andean belt whose width coincides with the eastward pinch-out of evaporites of the Triassic-Jurassic Pucara Formation. In the Santiago Basin (Navarro *et al.* 2005), as in the Colombian Eastern Cordillera and sub-Andes and Oriente Basin of Ecuador (Barragan *et al.* 2005; see Figs 4.2 & 4.4), there is also a strong influence of inversion tectonics (Gil *et al.* 2001).

Notably, southwards in the Huallaga Basin (for location see the central parts of Figs 4.2 & 4.4) the detachment level is still constituted by the evaporites of the Pucara Formation, but shortening in the Huallaga Basin reaches ~40% (Hermoza *et al.* 2005), whereas shortening in the Santiago Basin is only ~13.5%. The difference between the Huallaga (see Fig. 4.2) and Santiago Basins could be related to the width of the area where the Pucara Formation was deposited in both basins. However, Gil *et al.* (2001) proposed that the significant difference also could be due to the transpressional regime north of the Huancabamba deflexion (see Fig. 4.2), where the Santiago Basin is located. In this case, as in the Colombian sub-Andean zone, the obliquity between the inherited rift boundary and the applied stresses is the cause for such a narrow sub-Andean front. Therefore, as proposed by Gil *et al.* (2001), this is a typical feature of the Northern Andes. In comparison, there is more shortening in the Huallaga Basin (see Fig. 4.2) and southwards in the Madre de Dios (see Figs 4.2 & 4.10), where the Central Andes are characterized by their orientation almost perpendicular to the plate convergence vector.

Gil *et al.* (2001) mention the virtual absence to the south of significant tectonic inheritance as an important factor that appears to precondition the degree of basement involvement (i.e. thin-skinned vs thick-skinned deformation). Therefore, following Gil *et al.* (2001), the absence of a Mesozoic synrift sequence and associated normal faults south of the Abancay deflection (see Fig. 4.2), and the presence of a thick sequence of Paleozoic platform sediments, preconditions the Madre de Dios, Beni (location in Fig. 4.2 and cross-section through the Madre de Dios Basin in Fig. 4.10) and Chaco sub-Andean zones (see Fig. 4.2) to be characterized by purely thin-skinned deformation. In turn, the absence of such Paleozoic platform sediments conditions the transition from the Bolivian sub-Andes (see Chaco basin in Figs 4.2 and 4.4) to the thick-skinned deformation of the Pampean ranges (Allmendinger *et al.* 1997; Kley *et al.* 1999) (the Pampean ranges are located south of the Chaco basin; see location of the Chaco basin in Fig. 4.2).

In spite of the marked changes in the structural style conditioned by the stress regime and palaeogeographic conditions, there is a unifying aspect that has been rarely noted in the literature. All the sub-Andean zones from Bolivia to Venezuela share a common feature, which is the youth of their main deformation phase (post-Late Miocene, see Figs 4.4 & 4.7). The main argument (Baby *et al.* 1995; Dunn *et al.* 1995; Casero *et al.* 1997; Gil 2001; Mora *et al.* 2007) that most of the deformation occurred after the Late Miocene, is that Late Miocene and younger rock units are involved in the deformation, whereas older growth stratal relations are scarce and only preserved in some places in Late Miocene or younger sequences. This hypothesis has been confirmed using thermochronology (AFT) in the Northern Andes of Colombia, where the frontal thrusts have ages for the onset of exhumation between 12 and 3 Ma (Mora *et al.* 2008, see one such locality in Fig. 4.10). In fact, Mora *et al.* (2008) were also able to demonstrate that the majority of the shortening in the Northern Andes of Colombia occurred after the Late Miocene. AFT data have also been used to show a Late Miocene minimum age for exhumation prompted by tectonic stacking in duplexes and thrust faults in the Santiago Basin (Baby *et al.* 2005), the Huallaga Sub-basin (Alvarez-Calderon 1999) and the Madre de Dios Basin (Mora, A., Baby, P. & Roddaz, M., unpublished work; Fig. 4.2).

For example, in the Madre de Dios Basin (Figs 4.2 & 4.10), Mora *et al.* (unpublished work) used AFT cooling ages in a triangular zone, located at the backlimb of the Puquiri syncline (see Fig. 4.10b) to propose a minimum age for the deformation in this triangular zone. Cooling ages from those samples passing chi-squared cluster around 11 Ma (samples MD 22 and MD 26) corresponding to ages measured on Tertiary, pre-Middle Miocene units. Sample MD 22 comes from a unit just below the base of documented growth strata (see Fig. 4.10b). However, there are younger populations in outcropping Permian and Cretaceous rock units. Based on these considerations, Mora *et al.* (unpublished work) say that folding in the eastern flank of the Punquiri syncline, and therefore shortening of the underlying duplexes, started around 11 Ma and has probably been active through Late Miocene and Pliocene times as deduced from younger populations in other samples. Late Miocene deformation has also been shown to the south in the sub-Andean ranges adjacent to the Chaco foreland (Moretti *et al.* 1996; Echavarria *et al.* 2003) and inferred in the Beni Basin (Strub 2006) (see Fig. 4.2).

Following the aforementioned exhumation and palaeoelevation data, deformation in the sub-Andean thin-skinned ranges is normally synchronous with a period of faster denudation rates in the adjacent Eastern Cordillera. Even with the limited resolution of the data that we have, when comparing the aforementioned palaeoelevation data with the Late Miocene-Pliocene age of deformation in the sub-Andes, it appears that the sub-Andean zones began deforming close to the time when a certain elevation was reached in the hinterland Andean regions or at least synchronous with the main topographic growth in the Andean Eastern Cordilleras. The upheaval of the low-elevation sub-Andean ranges should have also influenced the river networks and the amount of material being transported from the Andes to the Amazon foreland.

The Vaupés Swell and the Fitzcarrald Arch

Traditionally, the Mérida Andes (Northern Andes of Venezuela) was considered as the most important barrier in directing the

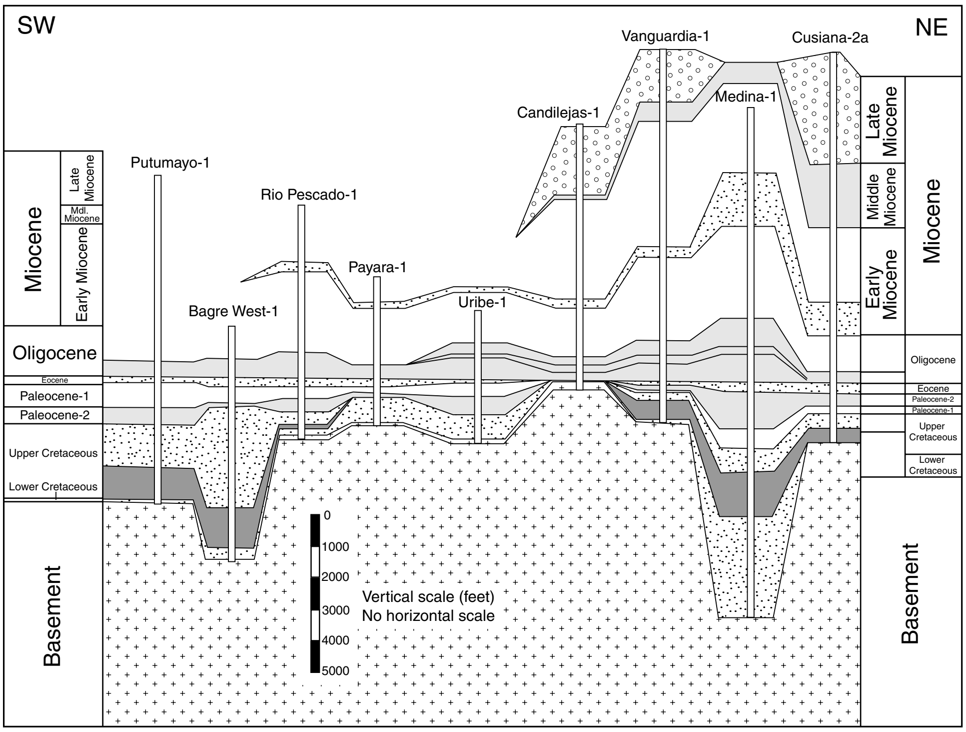

Fig. 4.11 Regional correlation of wells in the foredeep adjacent to the Colombian Eastern Cordillera (modified from Casero *et al.* 2007); the plot has no horizontal scale. The wells included are all those shown in Fig. 4.8. Notice the thicker Cretaceous interval in the Medina-1 well and the Bagre West-1 whereas the same interval is virtually absent in the Candilejas-1 well. This shows the presence of a palaeo-Vaupés Swell during the Cretaceous. A thinning due to post-depositional erosion in the Neogene interval from the Cusiana-2a well towards the Uribe-1 well is also very significant and shows the presence of a mostly post-Late Miocene positive area close to the Uribe-1. This area coincides with the Vaupés Swell – referred to by Casero *et al.* (1997) as Caquetá Nose.

Amazon River towards the Atlantic instead of towards the Caribbean (Hoorn *et al.* 1995). However, the real northern boundary for the Amazon hydrographic basin is the so-called Vaupés Swell (see Fig. 4.8), a western low-elevation promontory where the Guiana Shield is exposed or covered by a thinner sedimentary cover than in the Amazon lowlands to the south, or the Orinoquían savannas northwards (see Fig. 4.8). The Macarena, Chiribiquete and Araracuara low-elevation ranges represent the surface expression of the Vaupés Swell, where the Guiana Shield basement rocks crop out at elevations of up to 1000 m above sea level (see Fig. 4.8).

The origin of this low-elevation Guiana Shield promontory is currently unknown. However, the timing of its deformation can be inferred from many lines of evidence. Regional stratigraphic correlations show (Fig. 4.11) that the Vaupés Swell has been a positive area since the Late Cenomanian, as deduced from the thinning of the units deposited in that region (see Fig. 4.11; Casero *et al.* 1997). Both Paleogene and Cretaceous units are thinner towards the Vaupés Swell, but the most dramatic thickness change is evident in the Miocene rock units (see Fig. 4.11). In addition, seismic lines (Fig. 4.12) demonstrate that the thinning observed in the Miocene units is related to deposition and subsequent erosion of that interval. This is deduced by the absence of Miocene or older growth stratal relationships associated with the dip of the Vaupés Swell in its northwestern face. In contrast, all the Cenozoic units appear to be bent by post-Middle Miocene or even post-Late Miocene movements (see Fig. 4.12). In addition, Wesselingh *et al.* (2006) mention that east of the Araracuara range, around the Apaporis River, there are some patches of Late Miocene sediments (Apaporis sand unit) resting on top of basement. All these observations confirm the Late Miocene-Pliocene origin of the Vaupés Swell. We therefore infer a Late Miocene-Pliocene uplift of the present-day relief of the Vaupés Swell and await thermochronological data from the Macarena, Araracuara and Chiribiquete ranges to confirm our inference.

An analogue transverse feature in the Amazon drainage basin is the Fitzcarrald Arch, separating the Ucayali and Madre de Dios Basins in the Peruvian sub-Andes (see Fig. 4.2). The late Cenozoic origin of this basement arch has been documented by Espurt *et al.* (2007). In that case, a direct relationship has been proposed between the Fitzcarrald Arch and the kinematics of the Nazca ridge due to temporal coincidence between the formation of the former and the subduction of the latter, and the geophysically inferred spatial coincidence of both. In the case of the Vaupés Swell, such a relationship with subducted features has not yet been documented, but is possible.

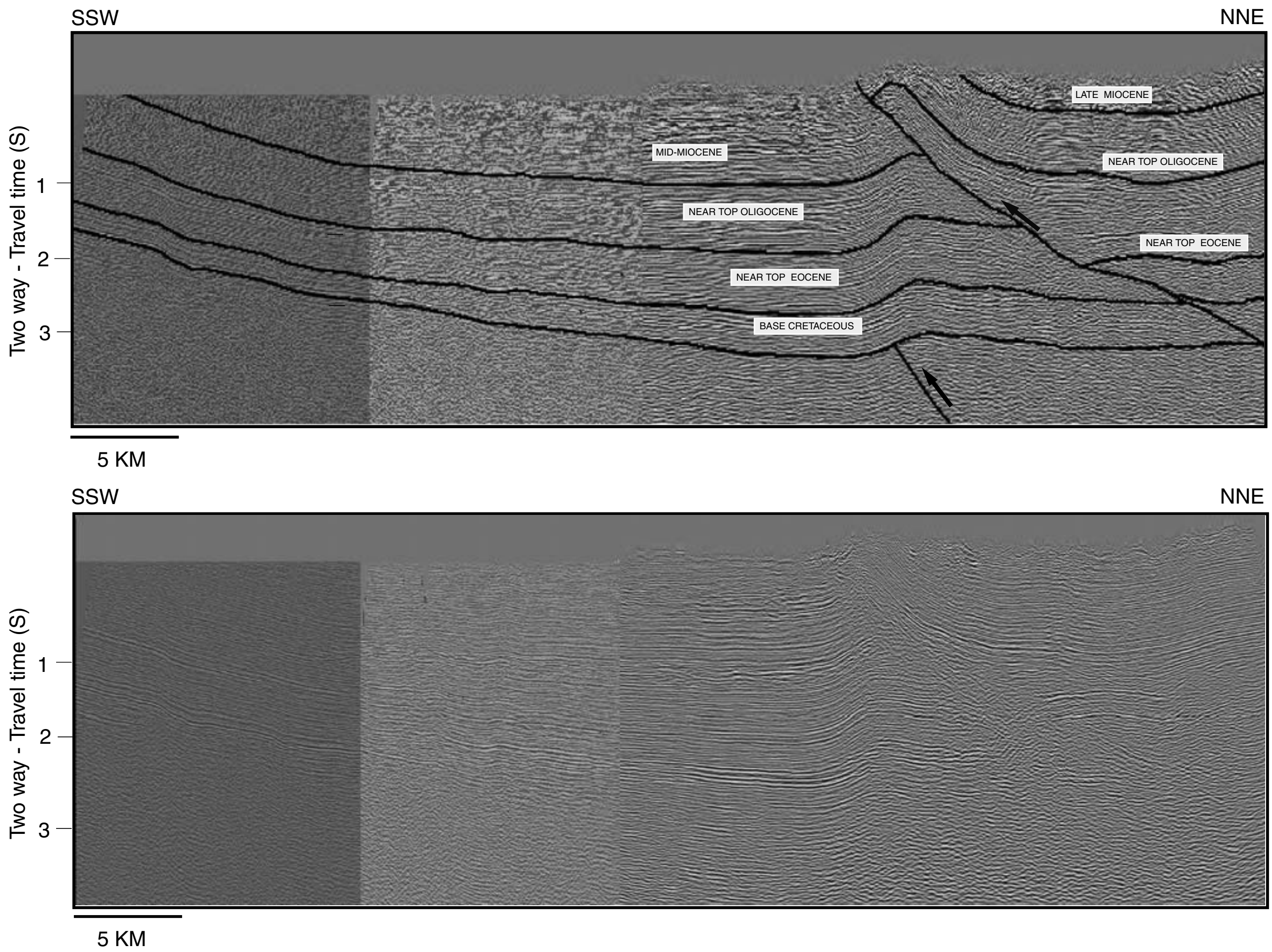

Fig. 4.12 Composite seismic line along the northwestern face of the Vaupés Swell (see location in Fig. 4.8). The frontal thrust of the eastern Colombian foothills (Guaicaramo Fault) is visible to the right. Further south (to the left) there is a progressive slow thinning of the Cretaceous to Miocene units. However, no evident growth stratal relationships can be identified. In contrast, the most significant feature is the post-Middle Miocene erosion that bevels the upper part of the Neogene sequence. This bevelling removes an amount of overburden equivalent to ~ 1 sec two-way travel time (TWT). Using velocities close to 4000 m/s this is ~2000 m of overburden removed. This is due to post-Middle Miocene northward tilting of the basin due to the uplift of the Vaupés Swell. A normal flexural response due to the load of the adjacent thrust would imply lower dip angles in the foreland plate (Bayona *et al.* 2008).

Discussion

The Andes and their relationship to the Amazon foreland and Amazon River dynamics

Based on a summary of the existing data on the history of the Northern and Central Andes and adjacent Amazon we here propose a hypothesis about the mechanisms that link Andean mountain building to reshaping of the continental drainage of northern South America. A key element in this hypothesis is the low-elevation basement arch that forms the present northern boundary of the Amazon drainage basin and separates the Caquetá hydrographic basin, with rivers flowing towards the Amazon, from the Guaviare hydrographic basin, with rivers flowing towards the Orinoco (see Figs 4.1 & 4.8).

Hoorn *et al.* (1995), among other ideas, proposed that the palaeogeographic scenario prior to the Late Miocene included a palaeo-Orinoco, collecting most of the Colombian, Ecuadorian and Peruvian Amazon tributaries and flowing towards the Maracaibo Lake (see this area in Fig. 4.1; for details see fig. 1A in Hoorn *et al.* 1995). Oligocene initial uplift and exhumation of the Colombian Eastern Cordillera, as documented by Parra *et al.* (2005), Gómez *et al.* (2001) and Bayona *et al.* (2008), and Late Oligocene-Early Miocene uplift and exhumation (Kohn *et al.* 1984; Higgs 1993; Colleta *et al.* 1997; Pindell *et al.* 1998) of the Mérida Andes, discard a palaeo-Maracaibo outlet for most of the rivers flowing north at that time. However, sedimentological data from Ecuador to Colombia (e.g., Hoorn *et al.* 1995 and references therein; Christophoul *et al.* 2002; Parra *et al.* 2005) show that a northward flow of most of the sub-Andean fluvial systems collected by a palaeo-Orinoco was likely. The Late Miocene or younger Vaupés Swell would have constituted a barrier from that time on, preventing the northern Amazon tributaries from flowing towards the Caribbean.

We consider the Vaupés Swell to be a key element, but not the only one, causing the majority of the rivers to stop flowing towards the Caribbean, as previously documented by Hoorn *et al.* (1995). Our main argument for additional elements is a pattern in the rivers of the Amazon drainage basin, which is very evident only at some distance from the Vaupés Swell. These rivers almost always flow to the east rather than to the north, with the notable exception of the Ucayali River (see Fig. 4.1). This pattern is reminiscent of the drainage pattern described in overfilled foreland basins, where accommodation space created in the foredeep is not enough to store sediments coming from the adjacent range (Flemings & Jordan 1989), therefore the drainage in the basin is not parallel to the foredeep and the range, but almost perpendicular (Flemings & Jordan 1989). This behaviour occurs when the rate of relief creation in the forebulge is surpassed by the sedimentation rates.

We hypothesize that during an underfilled stage, characterized by rivers dominantly flowing parallel to the Andes, the stream power of the rivers coming from the Andes should be capable of incising the Vaupés Swell. This is supported by the fact that rapidly uplifting ranges in the sub-Andes are always incised by rivers of different sizes, and by the fact that the Iquitos, Carauari and Purus arches (see Fig. 4.2 and Roddaz *et al.* 2005), which are basement arches similar to the Vaupés Swell, do not act as barriers for the eastward flow of the Amazon tributaries and Amazon River itself. In an overfilled stage of the Amazon basin, the influence of trasverse arches like the Vaupés Swell might be highlighted, as we will explain below. The transition from filled conditions in the Amazon foreland to overfilled conditions is therefore critical for the configuration of the Amazon flowing toward the Atlantic. Roddaz *et al.* (see Chapter 5) proposed that such a transition occurred in the Late Miocene all along the Amazonian and Llanos forelands. In relation to this the role of the Andes favouring or delaying such a transition should be assessed. Therefore the geological developments in the Andean hinterland that conditioned the transition from filled to overfilled stages in the Amazon foreland should be taken into account. In Chapter 5, Roddaz *et al.* explain in more detail the sedimentary signals of the transition from filled to overfilled.

A synchronism exists between many processes along the northern Central and Northern Andes. First, an Early Oligocene onset of deformation in the northern Central Andes compared with the Eastern Cordillera in the Northern Andes. Second, generalized Late Oligocene, Early Miocene deformation in the Andean Eastern Cordilleras of Bolivia and Colombia at least, and possibly in Peru. Maximum surface elevations in the Andean Eastern Cordilleras were also progressively reached during the post-Late Miocene. There is also a pattern of accelerated Late Miocene to Pliocene/Quaternary denudation in the eastern face of the Andean Eastern Cordilleras. Finally, after the Late Miocene, deformation migrated from the Eastern Cordilleras towards the sub-Andean zones. This partially synchronous chronology of geological processes and how these may affect the origin of the Amazon River is part of our hypothesis.

Flemings & Jordan (1989) show that the amount of sediment transported from an uplifting range toward the foreland is proportional to the slope, and to a transport coefficient that varies according to precipitation and climate. In turn, a critical palaeoelevation in the hinterland should be reached for the moisture-bearing winds to be trapped and generate focused precipitation (e.g. Bookhagen & Burbank 2006), as occurs today along the entire margin of the Andean Eastern Cordilleras (see precipitation maps in Chapter 14). Following Bookhagen's map, most of the rainfall in Amazonia should be focused along the Andean eastern front. The critical elevation for topography to focus precipitation has been deduced in the Central Andes and other mountain belts (Sobel & Strecker 2003; Blisniuk *et al.* 2005; Bookhagen *et al.* 2005; Strecker *et al.* 2007; Bookhagen & Strecker 2008). In such cases a value of ~2 km has been proposed. Deformation was present since the Oligocene in the Andean Eastern Cordilleras but palaeoelevation data show that such critical elevations were not reached until the Late Miocene. Thus, the data summarized here show that it is likely that rainfall was focused along the Andean Eastern Cordilleras since the Late Miocene in response to the Late Miocene major surface uplift of the Central and Northern Andes Eastern Cordilleras. An increased slope may also have been associated with this configuration. In turn, the amount of rainfall in the hinterland should also be proportional to the denudation rates in the same areas (Willet 1999). Therefore, the late Cenozoic uplift of the Andean Eastern Cordilleras may also be responsible for increased denudation rates by increasing rainfall, as documented in detail by Mora *et al.* (2008) for the Colombian Eastern Cordillera.

Noticeably increased granulometry in the Amazonian foredeep deposits (Christophoul *et al.* 2002; Roddaz *et al.* 2005; Parra *et al.* 2009b) is mostly synchronous with faster denudation rates (Mora *et al.* 2008; see Fig. 4.9). In addition, for the Ecuadorian foredeep Christophoul *et al.* (2002) document a late Cenozoic change in palaeocurrents from axial (parallel to the orogen) to transverse (perpendicular). Finally, in Chapter 5 Roddaz *et al.* show a shift in the Late Miocene record from tidal environments to fluvial all along the Central sub-Andean ranges. All these processes in the Amazon foreland reflect changing conditions in the Andean hinterland. In fact, based on a compilation of all the available sedimentological data in the sub-Andean basins, only partially mentioned here, Roddaz *et al.* (Chapter 5) propose a Late Miocene transition from filled Amazonian foreland to overfilled. This is in line with the processes happening in the Andean Eastern Cordilleras with increased surface uplift, inferred higher focused precipitation rates along the eastern side and faster denudation rates during the Late Miocene. It appears therefore that increased denudation rates in the Andean Eastern Cordillera have a causal relationship with the transition from filled to overfilled conditions in the adjacent foredeep, where higher sedimentation rates would have induced overfilled conditions. Such Late Miocene onset of overfilled conditions should have prompted the presence of an eastward-directed fluvial drainage pattern mostly typical of overfilled basins (Flemings & Jordan 1989). In this case, most of the sediment is transported outside the foredeep adjacent to the mountain front, as the foredeep is not capable of storing the sediment input.

Therefore, transverse drainage – similar to present – dominates. In such a scenario, features like the Fitzcarrald Arch (Pliocene) and Vaupés Swell (Late Miocene-Pliocene) are less likely to be incised by rivers flowing parallel to the range, as in the case of the Iquitos Arch, and their role as basement arches confining the river flow patterns is highlighted. In turn, more rivers flowing east, in an overfilled basin, should find an easier way to flow towards the Atlantic. With this evidence, a two-phase evolution of the Amazon River can be outlined.

First, during the Late Miocene the transcontinental drainage started under overfilled conditions and tributaries flowing perpendicular to the Andean deformation front. Second, unambiguously fluvial and coarser-grained sedimentary environments preserved in the record (see Chapter 5) document the presence of the modern Amazonian network since the Pliocene. We hypothesize that this final stage was prompted by Late Miocene-Pliocene uplift of the Vaupés Swell and Pliocene uplift of the Fitzcarrald basement arches, the fast deformation along the sub-Andean ranges, and an even more widespread drainage network in the Andean headwaters.

Migration of the sub-Andean ranges

The previous discussion raises the question why deformation only migrated towards the sub-Andean ranges after more than 20 Ma of deformation events in the Andes and why deformation migration towards the Subandes is virtually coeval with the birth of the Amazon river. Many studies propose that there is a mechanical interaction between deformation migration towards the foreland and elevation in the hinterland, explained in terms of the critical taper wedge (Davis *et al.* 1983; Dahlen *et al.* 1984; Dahlen & Suppe 1988). Roeder & Chamberlain (1995) proposed that the Late Miocene and later migration of deformation from the Eastern Cordillera and Altiplano towards the sub-Andean ranges could be due to the Eastern Cordillera having reached a critical palaeoelevation along extensive areas in the Central Andes. This is in agreement with available palaeoelevation data in the Andes and has been documented in detail in the Colombian Eastern Cordillera by comparing the timing of the structures in the sub-Andes and the palaeoelevation data inferred from pollen analysis (e.g. Mora 2007). However, Roeder & Chamberlain (1995) say that simple wedge mechanics involving a simple shear mechanism of deformation for the entire Altiplano, Eastern Cordillera and sub-Andean ranges requires a complex succession of events, including a plateau collapse after a main phase of crustal thickening. Alternatively, Oncken *et al.* (2006) and Babeyko *et al.* (2006) suggest, based on modelling, that a change in the mechanical properties (cohesion and friction) of the sub-Andean sedimentary wedge is needed for deformation to migrate from the Eastern Cordillera to the sub-Andean foredeep.

As already suggested by Babeyko *et al.* (2006), two possible scenarios can favour the change in mechanical conditions: increased pore pressure along the basal detachment due to hydrocarbon generation, and increased pore pressure in the wedge due to higher precipitation rates caused by the surface uplift of the Eastern Cordillera. The latter scenario has the problem that higher precipitation would make the entire sedimentary wedge weaker, not just the detachment. Wedge mechanics suggest that a less resistant wedge increases basal friction and decreases the effectiveness of the detachment horizon (Davis *et al.* 1983; Dahlen *et al.* 1984; Dahlen & Suppe 1988). In contrast, a documented late Cenozoic event of hydrocarbon generation from Colombia to Bolivia (Baby *et al.* 1995; Cazier *et al.* 1995) is more likely to facilitate the presence of a weaker detachment, given the fact that the hydrocarbon source rocks are also the basal detachments for the sub-Andean thrustbelt. Late Miocene hydrocarbon generation in the sub-Andean ranges from Colombia to Bolivia occurred at that time because only then was burial sufficient to heat the source rocks enough for them to enter the oil generation window. An increase in denudation rates in the Eastern Cordilleras producing a considerable amount of molasse deposits favours this scenario of rapid source rock burial and heating. Importantly, these molasse deposits are also coeval with the proposed onset of overfilled conditions.

Conclusions

From the data presented above we suggest that the evolution of the Andes went through a stage of generalized rock uplift and – for the first time – deformation spread over similar geographical areas as today. This stage could be as old as 30 Ma, an age that is in line with the observations of Oncken *et al.* (2005) in the Central Andes. We propose that the origin of the Amazon River and its flow towards the Atlantic is not precisely related to this stage. Rather, we suggest that it is related to Late Miocene-Pliocene faster denudation rates in the Andes. Faster denudation rates probably started when the eastern margin of the Andes reached a critical elevation during the Late Miocene, in an already growing mountain range. We hypothesize that faster denudation rates coincide with threshold elevations in the Andean Eastern Cordilleras, because it was only at that time when orographic precipitation was fully developed as a consequence of the newly created topographic barrier. Probably then, drainage areas of high topography were also much more extended than in the previous times.

In this scenario, it is likely that the Late Miocene transition from filled to overfilled conditions (see Chapter 5) in the Amazon drainage basin is the result of a sequence of processes in the Andean Orogen. Thus, a greater amount of sediments coming from the Andes may have caused the basin to evolve from filled to overfilled conditions during the Late Miocene. Our main hypothesis suggests that those conditions favoured the initial development of the transcontinental Amazonian drainage. An instrumental piece of evidence supporting this hypothesis, which links the development of the transcontinental network and the orogenic processes, is the fact that all the exposed Late Miocene timing of surface uplift and accelerated subsidence in the Andes and change in sedimentation patterns in the Amazon foredeep coincide with the initiation of the Amazon transcontinental network by 11 Ma, suggested by the sedimentary record from offshore Brazil (Dobson *et al.* 2001; Figueiredo *et al.* 2009).

However, the modern Amazon network, namely the large drainage area, with similar distribution of depositional systems and flow patterns, sourced in the Andean Eastern Cordilleras and eastward drainage into the Atlantic Ocean, was only fully established after the Pliocene (see Chapter 5 and Figueiredo *et al.* 2009). We consider that the Late Miocene to Pliocene uplift of the Vaupés Swell, the Pliocene and younger uplift of the Fitzcarrald Arch, the fast deformation along the sub-Andean ranges, and an even more widespread drainage network in the Andean headwaters may have prompted the final Pliocene development of the modern Amazon network. In addition we also propose that the following were the main factors that caused the palaeo-Amazon to stop flowing towards the Caribbean and be redirected to the Atlantic. We suggest that first the transition from filled to overfilled Amazonian foreland during the Late Miocene and second the Late Miocene-Pliocene uplift of the Vaupés Swell permanently separated the Orinoco from the Amazon drainage basins. Thus, in our view, surface uplift and deformation processes in the Andes not only controlled denudation rates, depositional systems, basin processes and thrust belt migration, but also the origin of the largest fluvial system in the world.

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References

Albert, J., Lovejoy, N., Crampton, W. (2006) Miocene tectonism and the separation of cis and trans–Andean river basins: evidence from Neotropical fishes. *J S Am Earth Sci* 21, 14–27.

Allmendinger, R.W., Jordan, T.E., Kay, S.M., Isacks, B.L. (1997) The evolution of the Altiplano-Puna Plateau of the Central Andes. *Annu Rev Earth Pl Sci* 25, 139–174.

Anders, M.H., Gregory-Wodzicki, K.M., Spiegelman, M. (2002) A critical evaluation of late Tertiary accelerated uplift rates for the Eastern Cordillera, central Andes of Bolivia. *J Geol* 110, 89–100.

Babeyko, A., Sobolev, S., Vietor, T., Oncken, O., Trumbull, R. (2006) Numerical study of weakening processes in the central Andean backarc. In: Oncken, O. *et al.* (eds) *The Andes – Active Subduction Orogeny*. Berlin: Springer-Verlag. pp. 495–510.

Baby, P., Moretti, I., Guillier,B., Limachi, R., Mendez, E., Oller, J., Specht, M. (1995) Petroleum system of the northern and central Bolivian sub-Andean zone. In: Tankard, A.J., Suarez R., Welsink, H.J. (eds) *Petroleum Basins of South America.* Tulsa: AAPG, pp. 445–458.

Baby, P., Rochat, P., Mascle, G., Hérail, G. (1997) Neogene shortening contribution to crustal thickening in the back arc of the Central Andes. *Geology* 25, 883–886.

Baby, P., Rivadeneira, M., Christophoul, F., Barragan, R. (1999) Style and timing of deformation in the Oriente Basin of Ecuador. In: *4th International Symposium on Andean Geodynamics (ISAG ´99), Göttingen, Germany*, pp. 68–72.

Baby, P. *et al.* (2005) Geodinámica mio-pliocénica de las cuencas subandinas peruanas: un mejor entendimiento de los sistemas petroleros. *V International Seminar INGEPET, Lima, Perú.*

Barke, R., Lamb, S. (2006) Late Cenozoic uplift of the Eastern Cordillera, Bolivian Andes. *Earth Planet Sci Lett* 249, 350–367.

Barnes, J.B., Ehlers, T.A., McQuarry, N., O Sullivan, P., Pelletier, J.D. (2006) Eocene to recent variations in erosion across the central Andean fold-thrust belt, northern Bolivia: Implications for Plateau evolution. *Earth Planet Sci Lett* 248, 118–133.

Barragan, R., Baby, P., Duncan, R. (2005) Cretaceous alkaline intraplate magmatism in the Ecuadorian Oriente Basin: Geochemical, geochronological and tectonic evidence. *Earth Planet Sci Lett* 236, 670–690.

Bayona, G., Cortes, M., Jaramillo, C., Ojeda, G., Aristizabal, J.J., Reyes-Harker, A. (2006b) Distinguishing tectonic versus eustatic surfaces in the Llanos Basin of Colombia, and implications for stratigraphic correlations. Extended abstracts, *IX Simposio Bolivariano de Exploración Petrolera en las Cuencas Subandinas, Cartagena de Indias, Colombia.*

Bayona, G., Cortés, M., Jaramillo., C., Ojeda, G., Aristizabal, J.J., Reyes-Harker, A. (2008) An integrated analyisis of an orogen-sedimentary basin pair: Latest Cretaceous-Cenozoic evolution of the linked Eastern Cordillera orogen and the Llanos foreland basin of Colombia. *Bull Geol Soc Am* 120. doi: 10.1130/B26187.1

Benjamin, M.T., Johnson, N.M., Naeser, C.W. (1987) Recent rapid uplift in the Bolivian Andes: Evidence from fission-track dating. *Geology* 15, 680–683.

Blisniuk, P.M., Stern, L.A., Chamberlain, C.P., Idleman, B., Zeitler, P.K. (2005) Climatic and ecologic changes during Miocene surface uplift in the southern Patagonian Andes. *Earth Planet Sci Lett* 230, 125–142.

Bookhagen, B., Burbank, D.W. (2006) Topography, relief, and TRMM-derived rainfall variations along the Himalaya. *Geophys Res Lett* 33, L08405. doi: 10:1029/2006GL026037

Bookhagen, B., Strecker, M.R. (2008) Orographic barriers, high-resolution TRMM rainfall, and relief variations along the eastern Andes. *Geophys Res Lett* 35, l06403. doi: 10.1029/2007gl032011

Bookhagen, B., Thiede, R.C., Strecker, M.R. (2005) Extreme monsoon events and their control on erosion and sediment flux in the high, arid NW Himalayas. *Earth Planet Sci Lett* 231, 131–146.

Burgos, J.D.Z. (2006) Genese et progradation d'un cone alluvial au front d'une chaine active: example des Andes Equatoriennes au Neogene. PhD thesis, Université Paul Sabatier, Toulouse, 373 pp.

Casero, P., Salel, J.F., Rossato, A. (1997) Multidisciplinary correlative evidences for polyphase geological evolution of the foot-hills of the Cordillera Oriental (Colombia). In: *VI Simposio Bolivariano de Exploración Petrolera en las Cuencas Subandinas, Bogotá, Colombia*, pp. 100–118.

Cazier, E.C., Hayward, A.B., Espinosa, G., Velandia, J., Mugniot, J.F., Leel, W.G. (1995) Petroleum geology of the Cusiana Field, Llanos Basin foothills, Colombia. *AAPG Bulletin* 79, 1444–1463.

Cediel, F., Shaw, R., Cáceres, C. (2003) Tectonic assembly of the northern Andean Block. In: Bartolini, C., Buffler, R.T., Blickwede, J. (eds) *The Circum-Gulf of Mexico and the Caribbean: Hydrocarbon Habitats, Basin Formation and Plate Tectonics. American Association of Petroleum Geologists Memoir* 79, 815–848.

Christophoul, F., Baby, P., Dávila, C. (2002) Stratigraphic responses to a major tectonic event in a foreland basin: the Ecuadorian Oriente basin from Eocene to Oligocene times. *Tectonophysics* 345, 281–298.

Cobbold, P.R., Rosello, E.A., Roperch, P., Arriagada, C., Gómez, L.A., Lima, C. (2007) Distribution, timing and causes of Andean deformation across South America. *Geol Soc Lond* 272, 321–343.

Colletta, B., Roure, F., De Toni, B., Loureiro, D., Passalaqua, H., Gou, Z. (1997) Tectonic inheritance, crustal architecture, and contrasting structural styles in the Venezuelan Andes. *Tectonics* 16, 777–794.

Dahlen, F.A., Suppe, J. (1988) Mechanics growth and erosion of mountain belts. In: Clark, S.P.J. (ed.) *Processes in Continental Lithospheric Deformation.* Geological Society of America Special Paper #218, pp. 161–178.

Dahlen, F.A., Suppe, J., Davis, D. (1984) Mechanics of fold-and-thrust belts and accretionary wedges: Cohesive Coulomb theory. *J Geophys Res* 89(B12), 10087–10101.

Damuth, J.E., Kumar, N. (1975) Amazon Cone: Morphology, Sediments, Age and Growth Pattern. *Geol Soc Am Bull* 86, 863–878.

Davis, D., Suppe, J., Dahlen, F.A. (1983) Mechanics of fold-and-thrust belts and accretionary wedges. *J Geophys Res* 88(B2), 1153–1172.

Dobson, D.M., Dickens, G.R., Rea, D.K. (1997) Terrigenous sedimentation at the Ceara Rise. In: *Proceedings of ODP, Scientific Results 154.* College Station, TX: Ocean Drilling Program, pp. 465–473.

Dobson, D.M., Dickens, G.R., Rea, D.K. (2001) Terrigenous sediment on Ceara Rise: a Cenozoic record of South American orogeny and erosion. *Palaeogeogr Palaeocl* 165, 215–229.

Dunn, J., Hartshorn, K., Hartshorn, P. (1995) Structural styles and hydrocarbon potential of the Sub Andean thrust belt of southern Bolivia. In: Tankard, A.J., Suarez, S.R., Welsink, H.J. (eds) *Petroleum Basins of South America.* Tulsa: American Association of Petroleum Geologists, pp. 523–543.

Echavarria, L., Hernandez, R., Allmendinger, R., Reynolds, J. (2003) Subandean thrust and fold belt of northwestern Argentina: geometry and timing of the Andean evolution. *AAPG Bulletin* 87, 965–985.

England, P., Molnar, P. (1990) Surface uplift, uplift of rocks, and exhumation of rocks. *Geology* 18, 1173–1177.

Erslev, E.A. (1986) Basement balancing of Rocky Mountain foreland uplifts. *Geology* 14, 259–262.

Espurt, N., Baby, P., Brusset, S., Roddaz, M., Hermoza, W., Regard, V. *et al.* (2007) How does the Nazca Ridge subduction influence the modern Amazonian foreland basin? *Geology* 35, 515–518.

Espurt, N., Barbarand, J., Brusset, S., Baby, P., Roddaz, M., Hermoza, W. (2008) *Kinematic evolution of the Camisea Subandean thrust belt from apatite fission-track thermochronology, Peru.* Extended abstract in 7th International Symposium of Andean Geodynamics, pp. 199–202.

Fields, R.W. (1959) Geology of the La Venta badlands, Colombia, South America. *Univ Calif Publ Geol Sci* 32, 405–444.

Figueiredo, J., Hoorn, C., van der Ven, P., Soares, E. (2009) Late Miocene onset of the Amazon River and the Amazon deep-sea fan: evidence from the Foz do Amazonas Basin. *Geology* 37, 619–622.

Flemings, P.B., Jordan, T. (1989) A synthetic stratigraphic model of foreland basin development. *J Geophys Res* 94, 3851–3866.

Garver, J.I., Reiners, P.R., Walker, L.J., Ramage, J.M., Perry, S.E. (2005) Implications for timing of Andean uplift from thermal resetting of radiation-damaged zircon in the Cordillera Huayhuash, Northern Peru. *Geology* 113, 117–138.

Garzione, C.N., Molnar, P., Libarkin, J.C., MacFadden, B.J. (2006) Rapid late Miocene rise of the Bolivian Altiplano: Evidence for removal of mantle lithosphere. *Earth Planet Sci Lett* 241, 543–556.

Garzione, C.N., Hoke, G.D., Libarkin, J.C., Withers, S., MacFadden, B., Eiler, J. *et al.* (2008) Rise of the Andes. *Science* 320, 1304–1307.

Gil, W. (2001) Evolution latérale de la déformation d'un front orogénique: Exemple des bassins subandins entre 0° et 16°S. PhD thesis, l'Université Paul Sabatier, Toulouse.

Gil, W., Baby, P., Ballard, J.F. (2001) Structure and palaeogeographic control of the Peruvian Subandean zone. *Cr Acad Sci II A* 333, 741–748.

Gómez, E., Jordan, T.E., Allmendinger, R.W., Hegarty, K., Kelley, S., Heizler, M. (2001) Controls on architecture of the Late Cretaceous to Cenozoic southern Middle Magdalena Valley Basin, Colombia. *Geol Soc Am Bull* 115, 131–147.

Gómez, E., Jordan, T.E., Allmendinger, R.W., Cardozo, N. (2005) Development of the Colombian foreland-basin system as a consequence of the diachronous exhumation of the northern Andes. *Geol Soc Am Bull* 117, 1272–1292.

Graham, A., Gregory-Wodzicki, K.M., Wright, K.L. (2001) Studies in neotropical paleobotany. XV. A Mio-Pliocene palynoflora from the Eastern Cordillera, Bolivia: Implications for the uplift history of the Central Andes. *Am J Bot* 88, 1545–1557.

Gregory-Wodzicki, K.M. (2000) Uplift history of the Central and Northern Andes: A review. *Bull Geol Soc Am* 112, 1091–1105.

Gregory-Wodzicki, K.M., McIntosh, W.C., Velasquez, K. (1998) Climatic and tectonic implications of the late Miocene Jakokkota Flora, Bolivian Altiplano. *J S Am Earth Sci* 11, 533–560.

Gutscher, M., Olivet, J., Aslanian, D., Maury, R., Eissen, J. (1999) The lost Inca Plateau: cause of flat subduction beneath Peru? *Earth Planet Sci Lett* 171, 335–341.

Hampel, A. (2002) The migration history of Nazca Ridge along the Peruvian active margin: A re-evaluation. *Earth Planet Sci Lett* 203, 665–679.

Hermoza, W., Brusset, S., Baby, P., Gil, W., Roddaz, M., Guerrero, N., Bolaños, M. (2005) The Huallaga foreland evolution: thrust propagation in deltaic environment, in the northern Peruvian Andes. *J S Am Earth Sci* 19, 495–512.

Higgs, R. (1993) Sedimentological evidence for early uplift (Oligocene) of the Venezuelan Andes. *AAPG Bulletin* 77, 325.

Hooghiemstra, H., Van der Hammen, T. (1998) Neogene and Quaternary development of the neotropical rain forest: the forest refugia hypothesis, and a literature overview. *Earth Sci Rev* 44, 147–183.

Hooghiemstra, H., Wijninga, V.M., Cleef, A.M. (2006) The Paleobotanical record of Colombia: Implications for biogeography and biodiversity. *Ann MO Bot Gard* 93, 297–325.

Hoorn, C. (2006) The birth of the mighty Amazon. *Sci Am* 294, 52–59.

Hoorn, C., Guerrero, J., Sarmiento, G.A., Lorente, M.A. (1995) Andean tectonics as a cause for changing drainage patterns in Miocene northern South America. *Geology* 23, 237–240.

Horton, B. (1998) Sediment accumulation on top of the Andean orogenic wedge: Oligocene to late Miocene basins of the Eastern Cordillera, southern Bolivia. *Geol Soc Am Bull* 110, 1174–1192.

Horton, B.K. (2005) Revised deformation history of the central Andes: Inferences from Cenozoic foredeep and intramontane basins of the Eastern Cordillera, Bolivia. *Tectonics* 24; doi:10.1029/2003TC001619.

Jordan, T.E., Isacks, B.L., Allmendinger, R.W. *et al.* (1983) Andean tectonics related to geometry of subducted Nazca plate. *Geol Soc Am Bull* 94, 341–361.

Jordan, T., Reynolds, J., Erikson, J. (1997) Variability in age of initial shortening and uplift in the Central Andes. 16–33.30°S. In: Ruddiman, W.F. (ed.) *Uplift and Climate Change.* New York: Plenum Press, pp. 41–61.

Kaandorp, R.J.G. *et al.* (2005) Seasonal Amazonian rainfall variation in the Miocene climate optimum. *Palaeogeogr Palaeoclim* 221, 1–6.

Kennan, L., Lamb, S.H., Hoke, L. (1997) High altitude paleosurfaces in the Bolivian Andes: Evidence for late Cenozoic surface uplift. In: *Paleosurfaces: Recognition, Reconstruction, and Interpretation.* Geological Society of London, Special Publication, pp. 307–324.

Kleinert, K., Strecker, M. (2001) Climate change in response to orographic barrier uplift: Paleosol and stable isotope evidence from the Late Neogene Santa Maria basin, northwestern Argentina. *Geol Soc Am Bull* 113, 728–746.

Kley, J. (1996) Transition from basement-involved to thin-skinned thrusting in the Cordillera Oriental of southern Bolivia. *Tectonics* 15, 763–775.

Kley, J., Monaldi, C.R., Salfity, J.A. (1999) Along-strike segmentation of the Andean foreland: causes and consequences. *Tectonophysics* 301, 75–94.

Kohn, B.P., Shagam, R., Banks, P.O., Burkley, L.A. (1984) Mesozoic-Pleistocene Fission Track ages on rocks of the Venezuelan Andes and their tectonic implications. *Geol Soc Am Bull* 162, 365–384.

Laubacher, G., Naeser, C.W. (1994) Fission-track dating of granitic rocks from the Eastern Cordillera of Peru: evidence for late Jurassic and Cenozoic cooling. *J Geol Soc Lond* 151, 473–483.

Macedo, J., Marshak, S. (1999) Controls on the geometry of fold-thrust belt salients. *Bull Geol Soc Am* 111, 1808–1822.

Marrett, R., Peacock, D.C.P. (1999) Strain and stress. *J Struct Geol* 21, 1057–1063.

Marshak, S., Wilkerson, M.S., Hsui, A. (1992) Generation of curved fold-thrust belts: insight from simple physical and analytical models. In: McClay, K.R. (ed.) *Thrust Tectonics and Petroleum Systems.* London: Chapman and Hall, pp. 83–92.

Martínez, J. (2006) Structural evolution of the Llanos foothills, Eastern Cordillera, Colombia: *J S Am Earth Sci* 21, 510–520.

Masek, J.G., Isacks, B.L., Gubbels, T.L., Fielding, E.J. (1994) Erosion and tectonics at the margins of continental plateaus. *J Geophys Res* 99(B7), 941–956.

Mégard, F. (1984) The Andean orogenic period and its major structures in central and northern Peru. *J Geol Soc Lond* 141, 893–900.

Mora, A. (2007) Inversion tectonics and exhumation processes in the eastern cordillera of Colombia. PhD thesis, Potsdam University, Potsdam.

Mora, A., Parra, M., Strecker, M.R. (2006) The role of penetrative deformation in orogenic processes. An example from the Eastern Cordillera of Colombia. 11. Abstracts of Symposium Tektonik, Struktur- und Kristallingeologie.

Mora, A., Parra, M., Strecker, M.R., Sobel, E.R., Hooghiemstra, H., Torres, V., Vallejo-Jaramillo, J. (2008) Climatic forcing of asymmetric orogenic evolution in the Eastern Cordillera of Colombia. *Bull Geol Soc Am* 120, 930–949. doi: 10.1130/B26186.1

Moretti, I., Baby, P., Mendez, E., Zubieta, D. (1996) Hydrocarbon generation in relation to thrusting in the Subandean zone from 18º to 22ºS, South Bolivia. *Petrol Geosci* 2, 17–28.

Navarro, L., Baby, P., Bolaños, R. (2005) Structural style and hydrocarbon potential of the Santiago basin. In: *V International Seminar INGEPET, Lima, Peru*, pp. 1–16.

Noble, D., McKee, E., Farrar, E., Petersen, U. (1974) Episodic Cenozoic volcanism and tectonism in the Andes of Peru. *Earth Planet Sci Lett* 21, 213–220.

Noble, D., Mégard, F., McKee, E. (1979) Early tertiary "incaic" tectonism, uplift and volcanic activity, Andes of Central Perú. *Geol Soc Am Bull* 90, 903–907.

Oncken, O., Hindle, D., Kley, J. Elger, K., Victor, P., Schemmann, K. (2006) Deformation of the Central Andean Upper Plate System – facts, fiction, and constraints for plateau models. In: Oncken, O. *et al.* (eds) *The Andes – Active Subduction Orogeny*. Berlin: Springer-Verlag, pp. 3–27.

Parra, M., Mora, A., Jaramillo, C., Veloza, G., Strecker, R. (2005) New stratigraphic data on the initiation of mountain building at the Eastern Front of the Colombian Eastern Cordillera. In: International Symposium on Andean Geodynamics, Barcelona.

Parra, M., Mora, A., Jaramillo, C., Strecker, M.R., Sobel, E.R., Quiroz, L.I. *et al.* (2009a) Orogenic wedge advance in the northern Andes: Evidence from the Oligocene-Miocene sedimentary record of the Medina Basin, Eastern Cordillera, Colombia: *Geol Soc Am Bull* 121, 780–800.

Parra, M., Mora, A., Sobel, E.R., Strecker, M.R., González, R. (2009b) Episodic orogenic-front migration in the northern Andes: constraints from low-temperature thermochronology in the Eastern Cordillera, Colombia: *Tectonics* 28, TC4004.

Pindell, J.L., Higgs, R., Dewey, J.F. (1998) Cenozoic palinspastic reconstruction, paleogeographic evolution, and hydrocarbon setting of the northern margin of South America. In: Pindell, J.L., Drake, C.L. (eds) *Paleogeographic Evolution and Non-glacial Eustasy, Northern South America.* Tulsa: Society for Sedimentary Geology, pp. 45–86.

Restrepo-Pace, P.A., Colmenares, F., Higuera, C., Mayorga, M. (2004) A fold-and-thrust belt along the western flank of the Eastern Cordillera of Colombia. Style, kinematics, and timing constraints derived from seismic data and detailed surface mapping. In: McClay, K.R. (ed.) *Thrust Tectonics and Hydrocarbon Systems.* American Association of Petroleum Geologists Memoir 82, pp. 598–613.

Rochat, P., Hérail, G., Baby, P., Mascle, G. (1999) Bilan crustale et contrôle de la dynamique sédimentaire sur les mécanismes de formation de l'Altiplano. *CR Acad Sci* 328, 189–195.

Roddaz, M., Baby, P., Brusset, S., Hermoza, W., Darrozes, J.M. (2005) Forebulge dynamics and environmental control in Western Amazonia: The case study of the Arch of Iquitos (Peru). *Tectonophysics* 399, 87–108.

Roeder, D., Chamberlain, C.P. (1995) Structural geology of sub-Andean fold and thrust belt in northwestern Bolivia. In: Tankard, A.J., Suarez, S.R., Welsink, H.J. (eds) *Petroleum Basins of South America.* Tulsa: American Association of Petroleum Geologists, pp. 459–479.

Sanderson, D., Marchini, W.R.D. (1984) Transpression. *J Struct Geol* 6, 449–458.

Sdrolias, M., Müller, R.D. (2006) Controls on back-arc basin formation. *Geochem Geophys Geosyst* 7, Q04016, doi: 10.1029/2005GC001090

Silver, P.G., Russo, R.M., Lithgow, B.C. (1998) Coupling of South American and African Plate motion and plate deformation. *Science* 279, 60–63.

Sobel, E.R., Strecker, M.R. (2003) Uplift, exhumation and precipitation: Tectonic and climatic control of Late Cenozoic landscape evolution in the northern Sierras Pampeanas, Argentina. *Basin Res* 15, 431–451.

Spikings, R.A., Seaward, D., Winkler, W., Ruiz, G. (2000) Low-temperature thermochronology of the northern Cordillera Real, Ecuador: Tectonic insights from zircon and apatite fission track analysis. *Tectonics* 19, 649–668.

Strecker, M.R., Alonso, R.N., Bookhagen, B., Carrapa, B., Hilley, G.E., Sobel, E., Trauth, M.H. (2007) Tectonics and climate of the southern Central Andes. *Ann Rev Earth Planet Sci* 35, 747–787.

Strub, M. (2006) Evolution géomorphologique et tectonique au front des Andes centrales (Nord de la Bolivie). PhD thesis, Université Paul Sabatier, Toulouse, 191 pp.

Toro, J., Roure, F., Bordas-Le Floch, N., Le Cornec-Lance, S., Sassi, W. (2004) Thermal and kinematic evolution of the Eastern Cordillera fold-and-thrust belt, Colombia. *AAPG Hedberg Series* 1, 79–115. http://www.aapg.org/pubs/aapgcopyright.cfm

Van der Hammen, T., Hooghiemstra, H. (2000) Neogene and Quaternary history of vegetation, climate, and plant diversity in Amazonia. *Quaternary Sci Rev* 19, 725–742.

Van der Hammen, T., Werner, J.H., Dommelen, V. (1973) Palynological record of the upheaval of the Northern Andes: a study of the Pliocene and Lower Quaternary of the Colombian Eastern Cordillera and the early evolution of its High-Andean biota. *Rev Paleobot Palynol* 16, 1–122.

Victor, P., Oncken, O., Glodny, J. (2004) Uplift of the western Altiplano Plateau: evidence from the Precordillera between 20 degrees S and 21 degrees S: northern Chile. *Tectonics* 23, 4, TC4004.

Wesselingh, F.P., Hoorn, M.C., Guerrero, J., Rasanen, M.E., Romero-Pittman, L., Salo, J. (2006) The stratigraphy and regional structure of Miocene deposits in western Amazonia (Perú, Colombia and Brazil), with implications for late Neogene landscape evolution. *Scripta Geologica* 133, 291–322.

Wijninga, V.M. (1996) Palynology and paleobotany of Neogene sediments from the high plain of Bogotá (Colombia): evolution of the Andean flora from an ecological perspective. PhD thesis, University of Amsterdam, 370 pp.

Willet, S.D. (1999) Orogeny and orography: The effects of erosion on the structure of mountain belts. *J Geophys Res* 104, 28957–28981.