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Lithospheric evolution of the Andean fold–thrust belt, Bolivia, and the origin of the central Andean plateau

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Abstract

We combine geological and geophysical data to develop a generalized model for the lithospheric evolution of the central Andean plateau between 18° and 20° S from Late Cretaceous to present. By integrating geophysical results of upper mantle structure, crustal thickness, and composition with recently published structural, stratigraphic, and thermochronologic data, we emphasize the importance of both the crust and upper mantle in the evolution of the central Andean plateau. Four key steps in the evolution of the Andean plateau are as follows. 1) Initiation of mountain building by ~70 Ma suggested by the associated foreland basin depositional history. 2) Eastward jump of a narrow, early fold–thrust belt at 40 Ma through the eastward propagation of a 200–400-km-long basement thrust sheet. 3) Continued shortening within the Eastern Cordillera from 40 to 15 Ma, which thickened the crust and mantle and established the eastern boundary of the modern central Andean plateau. Removal of excess mantle through lithospheric delamination at the Eastern Cordillera–Altiplano boundary during the early Miocene appears necessary to accommodate underthrusting of the Brazilian shield. Replacement of mantle lithosphere by hot asthenosphere may have provided the heat source for a pulse of mafic volcanism in the Eastern Cordillera and Altiplano at 24–23 Ma, and further volcanism recorded by 12–7 Ma crustal ignimbrites. 4) After ~20 Ma, deformation waned in the Eastern Cordillera and Interandean zone and began to be transferred into the Subandean zone. Long-term rates of shortening in the fold–thrust belt indicate that the average shortening rate has remained fairly constant (~8–10 mm/year) through time with possible slowing (~5–7 mm/year) in the last 15–20 myr. We suggest that Cenozoic deformation within the mantle lithosphere has been focused at the Eastern Cordillera–Altiplano boundary where the mantle most likely continues to be removed through piecemeal delamination.

Keywords: Central Andes; Plateaus; Fold–thrust belts; Crustal structure; Foreland basins

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1. Introduction

The Central Andes have been generally regarded as a young orogenic belt because of their high elevation and topography, rugged peaks little affected by erosion, and broad internally drained areas of low relief. The initial pulse of tectonism that constructed this topographic edifice is generally considered to be late Oligocene–early Miocene (25–30 Ma) in age (Isacks, 1988; Sempere et al., 1990; Gubbels et al., 1993, Allmendinger et al., 1997; Jordan et al., 1997). In fact, the 25–30 Ma range of ages for initiation of shortening within the Central Andes is so deeply entrenched in the literature that most thermo-mechanical and kinematic models require the entire construction of the central Andean plateau within this time period (e.g., Isacks, 1988; Gubbels et al., 1993; Wdowinski and Bock, 1994; Pope and Willett, 1998), or use this timespan to determine long-term rates of deformation of the orogen (Norabuena et al., 1998; 1999, Gregory-Wodzicki, 2000; Liu et al., 2000; Hindle et al., 2002). On the other hand, several authors have recognized an older history of deformation in the Eastern Cordillera based on thermochronologic data (Benjamin et al., 1987; McBride et al., 1987; Farrar et al., 1988; Sempere et al., 1990; Masek et al., 1994; Lamb et al., 1997), angular unconformities (Kennan et al., 1995; Lamb and Hoke, 1997; Sempere et al., 1997; Allmendinger et al., 1997), and ages of foreland basin deposits (Lamb and Hoke, 1997; Sempere et al., 1997; Horton, 1998; Horton et al., 2001; DeCelles and Horton, 2003). Using probable foreland basin deposits as a signal of the initiation of mountain building, several authors have proposed that a topographically high fold–thrust belt existed in the western portion of the Central Andes as early as Cretaceous to early Paleocene time (Coney and Evenchick, 1994; Sempere et al., 1997; Horton et al., 2001). The consolidation and eastward propagation of the Andean orogenic belt during Late Cretaceous to early Paleocene time are supported by rapid (30–150 mm/year) convergence between the Nazca and South American Plates (Pardo-Casas and Molnar, 1987; Coney and Evenchick, 1994; Somoza, 1998; Norabuena et al., 1998; 1999).

Numerous studies have asserted that the thick crust (60–70 km) and high elevation of the central Andean plateau are linked to the large amount of tectonic shortening in the Central Andes (Lyon-Caen et al., 1985; Isacks, 1988; Roeder, 1988; Sheffels, 1990; Sempere et al., 1990; Gubbels et al., 1993; Schmitz, 1994; Allmendinger et al., 1997; Kley et al., 1996; Baby et al., 1997; Schmitz and Kley, 1997). This causative relationship between shortening and thickening directly ties the development of the central Andean plateau to the evolution and eastward propagation of the fold–thrust belt (McQuarrie, 2002). Nevertheless, the kinematic evolution of lithospheric-scale shortening remains elusive. Recent seismic studies in the Central Andes of Bolivia provide a detailed image of the upper mantle (Dorbath et al., 1993, Dorbath and Granet, 1996; Myers et al., 1998; Yuan et al., 2000), allowing mantle structure and topography to be correlated with surface topography and structural trends. Using recently published stratigraphic, thermochronologic, and geophysical data, we propose a lithospheric model describing how the Andean fold–thrust belt progressed with time to form the central Andean plateau and how the plateau itself reflects deformational processes that have been in action since the Late Cretaceous. To this end, we first combine structural, stratigraphic, and thermochronologic data to describe how upper crustal deformation progressed with time. This provides estimates of average long-term rates of geologic shortening and flexural wave migration that can be compared with modern GPS convergence rates. A review of the present lithospheric structure of the Central Andes, combined with the structural shortening history, allows for insight into mantle lithospheric deformation that must have accommodated crustal deformation.

2. Tectonic framework

The Central Andes have the highest average elevations, greatest width, thickest crust, and maximum shortening of the entire Andean orogenic belt (Isacks, 1988; Roeder, 1988; Sheffels, 1990; Sempere et al., 1990; Schmitz, 1994; Zandt et al., 1994; Beck et al., 1996; Kley et al., 1996; Allmendinger et al., 1997; Baby et al., 1997; Schmitz and Kley, 1997; Kley and Monaldi, 1998; McQuarrie, 2002). In the Central Andes of Peru, Bolivia, and parts of Argentina and Chile, an area of high elevation (greater that 3 km), generally internally drained basins, and low to moderate relief has been defined as the central Andean plateau (Isacks, 1988; Gubbels et al., 1993; Masek et al., 1994; Lamb and Hoke, 1997) (Fig. 1). The geology of the Central Andes between 18° and 20° S can be divided into nine longitudinal tectonomorphic zones (Fig. 1). From west to east, these include the Peru–Chile trench, the Coastal Cordillera (remnant of a Mesozoic arc), the Longitudinal Valley (a modern forearc basin), the Chilean Precordillera (remnant of a Paleogene arc), the Western Cordillera (the modern volcanic arc), the Altiplano (a high-elevation, internally drained, low-relief basin), the Eastern Cordillera (a bivergent portion of the Andean fold–thrust belt), the Subandean zone (the frontal, most active portion of the Andean fold–thrust belt), and the Chaco Plain (a low-elevation foreland basin underlain by the Brazilian shield) (Allmendinger et al., 1997; Horton et al., 2001). A simplified geologic map of the fold–thrust belt at 19–20° S from the Altiplano to the Subandean zone is shown in Fig. 2a.

3. Rationale

3.1. Two end member models

3.1.1. Short duration, low-magnitude shortening

A prevalent model of mountain building in the Central Andes argues that although there may have been earlier periods of deformation in both the Precordillera and the Eastern Cordillera, the deformation that culminated in the high-elevation plateau is predominantly Neogene (Isacks, 1988; Sempere et al., 1990; Gubbels et al., 1993, Allmendinger et al., 1997;

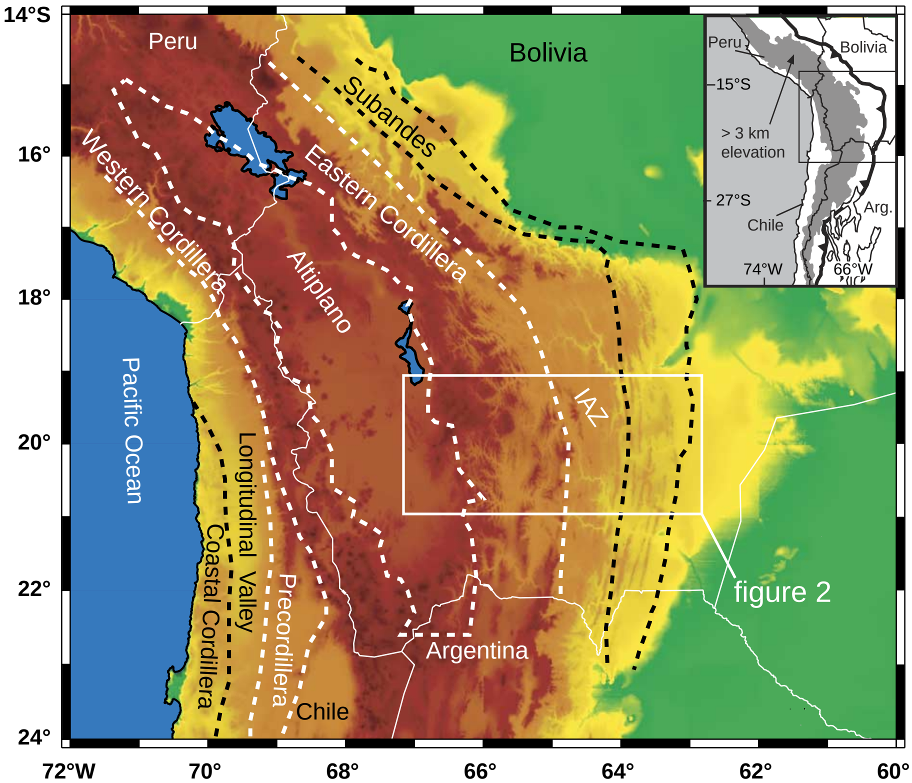

Fig. 1. Topography of the central Andes with major physiographic divisions separated by dashed lines. IAZ is the Interandean zone defined by Kley (1996). Inset (modified from Isacks, 1988) shows the geographical extent of the Andean plateau and the location of the central Andean portion of the plateau.

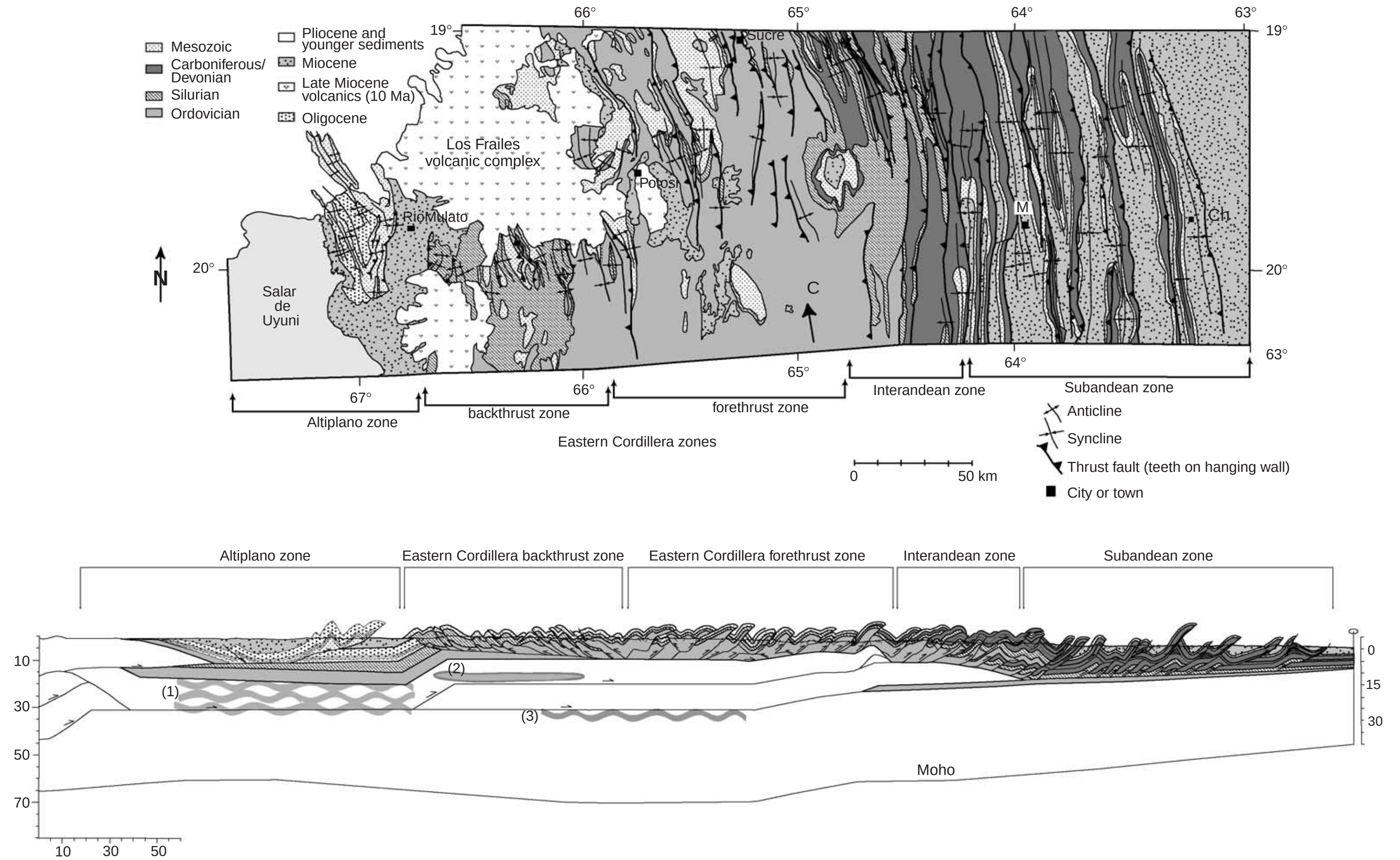

Fig. 2. (a) Simplified geologic map of the Andean fold–thrust belt between 19° and 20° S. The map is modified from Pareja et al. (1978), GEOBOL (1962a,b,c,d; 1996), and Suarez (2001). M=Monteagudo; Ch=Charagua; arrow and C=along-strike projection of Camargo. Structural zones discussed in paper identified by brackets. (b) Structural cross section from McQuarrie (2002). Low-velocity zones (gray waves and oval) from Beck and Zandt (2002) and Yuan et al. (2000). (1) Altiplano low-velocity zone (Yuan et al., 2000); (2) Los Frailes; and (3) Eastern Cordillera (Beck and Zandt, 2002).

Jordan et al., 1997). Timing estimates for the focused Neogene deformation are based on rapid (and possible low-angle) subduction at 26–27 Ma (Isacks, 1988; Allmendinger et al., 1997), 25–20 Ma synorogenic sediments deposited in basins preserved both east and west of the Eastern Cordillera in Bolivia (Sempere et al., 1990; Jordan et al., 1997), and a ~10 Ma low-angle erosion surface that post-dates deformation in the Eastern Cordillera (Gubbels et al., 1993). Shortening estimates for this period of mountain building rely heavily on shortening within the Subandes, where the geometry of deformation is well documented and decidedly Neogene (e.g., Roeder and Chamberlain, 1995; Dunn et al., 1995) (Table 1). The contribution of Eastern Cordilleran shortening has been hard to ascertain due to an incomplete understanding of the stratigraphy, pre-Andean deformation south of 20° S, uplift and erosion of strata prior to the Cretaceous, and low preservation of Tertiary strata (see discussions in Allmendinger et al., 1997; Müller et al., 2002). Low magnitudes of shortening (100–150 km) in the Eastern Cordillera suggest that much of the crustal thickness and subsequent elevation of the Andean plateau was the result of late (10 Ma–present) shortening of the Subandes. However, this scenario does not explicitly account for (a) the 70-km-thick crust of the Eastern Cordillera and Altiplano; (b) an Eocene cooling history in the Eastern Cordillera; and (c) the mid-Tertiary sedimentation history of the Eastern Cordillera and Altiplano.

3.1.2. Long-duration, large-magnitude shortening

The purpose of this paper is to present the other end member scenario for the evolution of the central Andean fold–thrust belt through the integration of several data sets (sedimentology/stratigraphy, structural geology, and geophysics) that are commonly considered separately. Our reconstructions are based on combining sequentially restored structural cross sections through the Central Andes with a simple four-part model for foreland basin systems that has been developed for modern and ancient fold–thrust belts (DeCelles and Giles, 1996).

Recent field work and accompanying maps and cross sections through the Eastern Cordillera have documented tight folds, rotated faults, and structures that propagate to the west and to the east (McQuarrie and DeCelles, 2001; McQuarrie, 2002; Müller et al., 2002). Dated sedimentary basins and mineral cooling ages suggest that this deformation started at least as early as 32 Ma, and possibly as early as 40 Ma (Horton, 1998; McQuarrie and DeCelles, 2001; Ege et al., 2001; Müller et al., 2002). Although there are discrepancies among authors as to how the brittle upper basement accommodates the shortening documented in the tightly folded and faulted Paleozoic and younger sedimentary rocks (compare McQuarrie, 2002; Müller et al., 2002), the style of basement shortening does not affect shortening estimates, which are as high as 285–330 km for the exposed portions of the central Andean fold–thrust belt (Table 1).

Table 1
Shortening estimates for the Andean fold–thrust belt and approximate rates (mm/yr)

Location	Altiplano	Eastern Cordillera Interandean zone	Subandean zone	Total	Eastern Cordillera			Subandes	
					30–20[a]	25–10[b]	40–15[c]	10–0[d]	15–0[e] Ma
Northern Bolivia									
15–17° S	–	157 km (1)	123 km (1)	279 km (1)	10	16	6	12	8
15–18° S	14 km (2)	103 km (2)	74 km (2)	191 km (2)	7	10	4	7	5
17–19° S	47 km (3)	181 km (3)	72 km (3)	300 km (3)	12	18	7	7	5
17–19° S	30 km (4)	–	–	240 km (4,5)					
Southern Bolivia									
20–21° S	20 km (2)	125 km (2)	86 km (2)	231 km (2)	8	13	5	9	6
21° S	50 km (7)	135 km (7,8)	100 km (9)	285 km (8)	8	14	5	10	7
20° S	41 km (3)	218 km (3)	67 km (3)	326 km (3)	15	22	9	7	5

References (in parentheses) are as follows: (1) Roeder and Chamberlain (1995); (2) Baby et al. (1997); (3) McQuarrie (2002); (4) Lamb and Hoke (1997); (5) Sheffels (1990); (6) Kley et al. (1997); (7) Kley (1996); (8) Müller et al. (2002); (9) Dunn et al. (1995).
Superscript a–e are time periods (myr) of Andean deformation: (a) Sempere et al. (1990); (b) Hindle et al. (2002), Allmendinger et al. (1997), and Isacks (1988); (c) this paper; (d) Gubbels et al. (1993); (e) this paper.

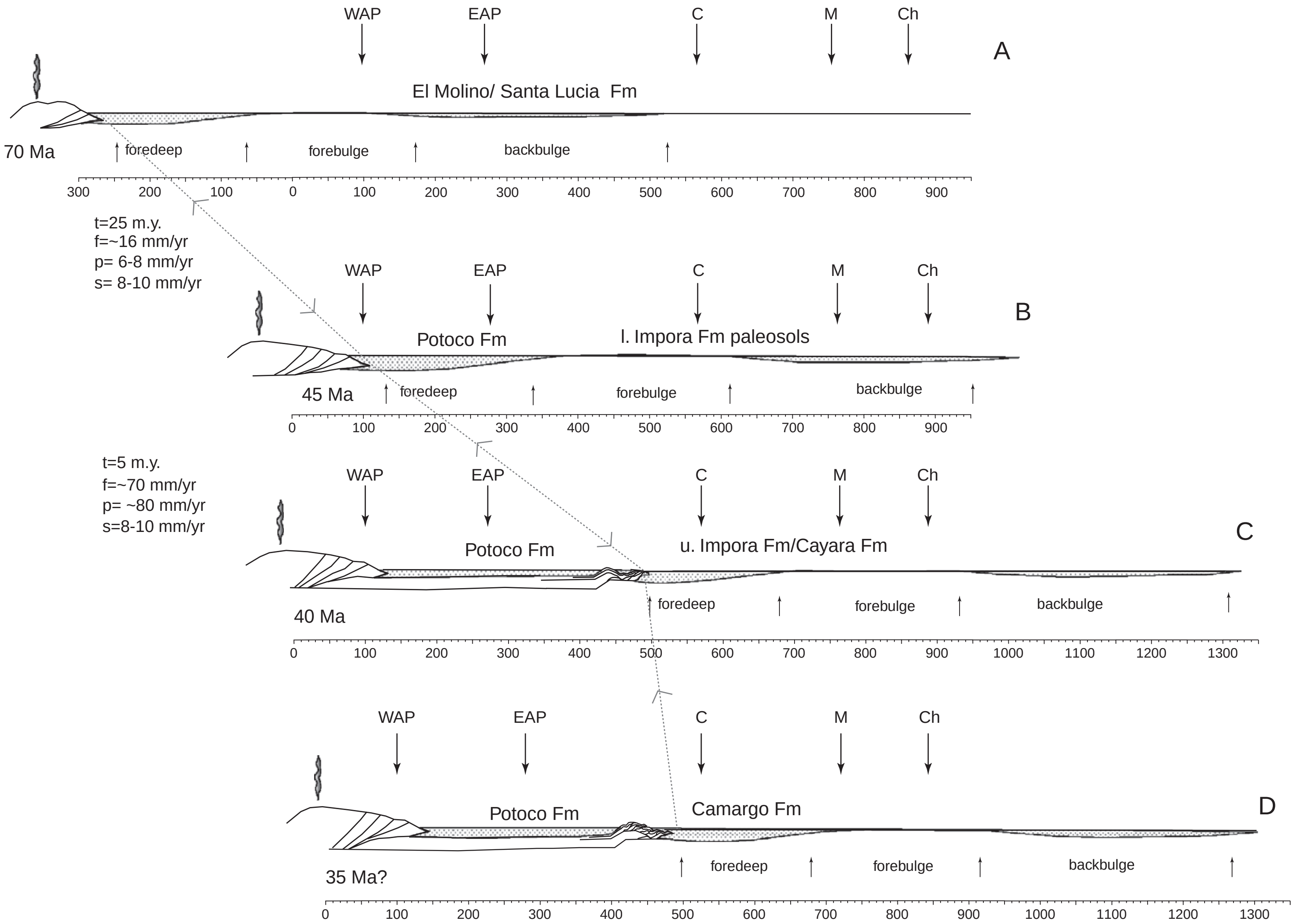

Fig. 3. Basin migration diagram integrating the foreland basin history with the evolution of the fold–thrust belt. WAP, Western edge of Altiplano; EAP eastern edge of Altiplano; C, Camargo; M, Monteagudo; Ch, Charagua; f, foreland basin migration rate; p, propagation rate of fold–thrust belt; s, shortening rate of fold–thrust belt. Locations of foreland basin boundaries determined from preserved basin deposits (where possible) and a lithospheric flexural rigidity of 3×10^{23} to 5×10^{23} Nm (Lyon-Caen et al., 1985; Goetze and Schmidt, 1999; DeCelles and Horton, 2003). Jagged line represents the modern western edge of the Brazilian shield. Structures portrayed west of the Altiplano are schematic; however, structures shown in the Altiplano through Subandean zone are from a sequentially restored balanced cross section simulated using 2-D MOVE (Midland Valley) (modified from McQuarrie, 2002).

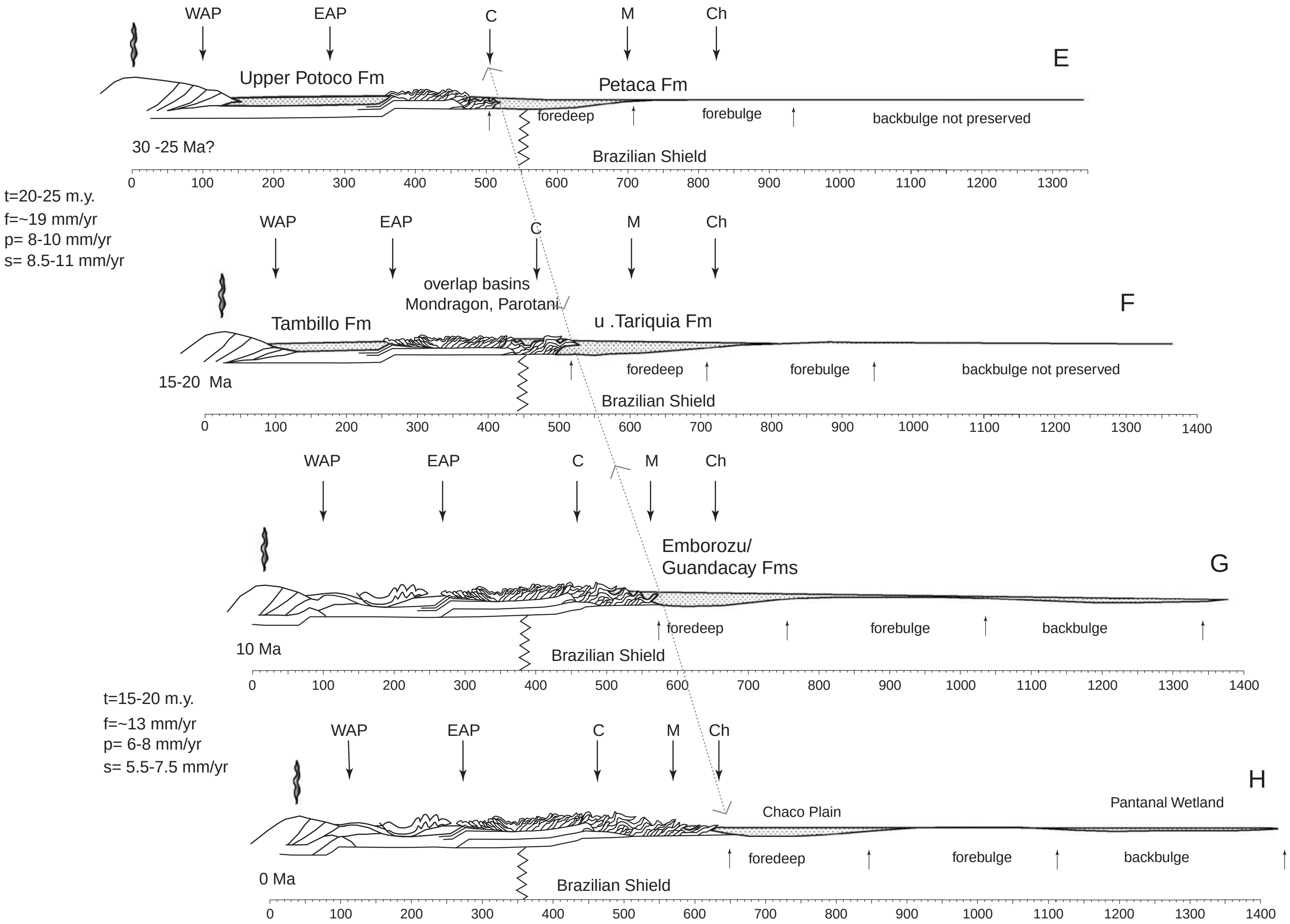

Fig. 3 (*continued*).

Historically, the influx of coarse sediment into a basin has been used to constrain the initiation of deformation within the adjacent fold–thrust belt. However coarse-grained proximal foreland basin deposits are also the first to be deformed and erosionally removed by the propagating fold–thrust belt (e.g., Burbank and Raynolds, 1988). Thus the earliest record of orogenesis may only be preserved in distal, fine-grained foreland basin deposits. In this paper, we use the sedimentological signature of distal and proximal foreland basin deposits to track the eastward evolution of the Andean fold–thrust belt with time (Fig. 3). The basin migration history is a robust record of long-term fold–thrust belt migration, and can be compared to total shortening estimates, growth of the fold–thrust belt, and kinematic behavior of the orogenic wedge (DeCelles and DeCelles, 2001).

3.2. Model rationale

Probable variations in deformation rates and lithospheric strength properties prevent a simple, smoothly migrating wave of deformation across the Andes. Nevertheless, we apply a first-order analysis of the structure, sedimentology, and geophysics of the Central Andes, which provides a reasonable, albeit imprecise, evolutionary history. Existing timing constraints are strengthened through new age data emerging for early Tertiary rocks (e.g., Horton et al., 2001) and by integrating thermochronologic cooling ages with sequentially balanced cross sections across the fold–thrust belt. By combining the basin migration history with the structural evolution of the fold–thrust belt, age constraints from both parts of the system suggest an internally consistent model for the growth of the Central Andes with time. This analysis is not intended to convey the precise kinematics of deformation on a million year timescale, but rather a broad understanding of the location, timing, and mechanisms of deformation across the belt on the order of tens of millions of years. Our purpose is to develop an integrated picture of the Cenozoic history of the region that satisfies the existing constraints offered by each sub-discipline. In this endeavor, we hope to stimulate discussion and identify particular areas in need of further analysis. We view the following synthesis as a work in progress that is suitable to modification as further age control, better constrained cross sections, and a clearer crustal/upper mantle picture become available.

4. Crustal evolution

4.1. Cretaceous (?)–Paleocene

4.1.1. Sedimentology

The Upper Cretaceous–mid-Paleocene sedimentary succession in Bolivia consists of ~1–2 km of mostly nonmarine mudrock, sandstone, and carbonate deposited in distal fluvio-lacustrine systems and marginal marine settings over much of the eastern Altiplano and Eastern Cordillera. This succession includes the upper Puca Group, which is composed of the Aroifilla, Chaunaca, El Molino, and Santa Lucia Formations (Sempere et al., 1997). The regional distribution, limited thickness, notable decrease in strata thickness to both the east and west, and preponderance of low-gradient depositional systems indicate an extensive zone of minor to moderate subsidence during Late Cretaceous–mid-Paleocene time. Some studies have attributed this deposition to thermal subsidence following a short pulse of rifting around ~120–100 Ma, similar to strata in northwestern Argentina that have been linked to Late Cretaceous normal faulting in the Salta rift system (Salfity and Marquillas, 1994; Welsink et al., 1995; Comínguez and Ramos, 1995; Viramontes et al., 1999). However, the sediment accumulation rates calculated for the upper Puca Group (Sempere et al., 1997) exceed typical values of post-rift subsidence, suggesting a different or additional process (DeCelles and Horton, 2003).

Cretaceous sedimentation may represent, in part, distal deposition in an early foreland basin system related to initial shortening in the westernmost parts of the Central Andes (Sempere et al., 1997; Horton and DeCelles, 1997). By analogy with the modern Andean foreland, this fill can be attributed to deposition in the backbulge depozone of the foreland basin system (DeCelles and Giles, 1996)—a region that would have been situated between an elevated forebulge to the west and the craton to the east (Horton and DeCelles, 1997; Lundberg et al., 1998). Although the age of initial compression is uncertain, the Maastrichtian to mid-Paleocene El Molino–Santa Lucia interval could represent part of the backbulge depozone, thus providing a minimum age estimate for the onset of regional shortening and crustal loading (see also Sempere et al., 1997).

Mid-Paleocene to middle Eocene rocks represent a period of minimal subsidence and limited sediment accumulation in the eastern Altiplano and Eastern Cordillera. Development of 20–100-m-thick zones of pervasive paleosol deposits in the uppermost Santa Lucia Formation and lowermost Potoco and Impora Formations of the eastern Altiplano (Horton et al., 2001) and Eastern Cordillera (DeCelles and Horton, 2003) attests to long intervals (10–20 myr) of limited sediment accumulation. Such conditions may be representative of negligible long-term accumulation over the structurally elevated forebulge depozone (e.g., DeCelles and Currie, 1996), as observed over a potential modern forebulge in the eastern Chaco Plain (Litherland and Pitfield, 1983; Litherland et al., 1986; Horton and DeCelles, 1997). Although the Upper Cretaceous–mid-Paleocene strata could be interpreted as rift-related, foreland basin-related, or a combination of both systems, the abrupt cessation of subsidence (represented by the pervasive paleosol deposits) documented in the Altiplano (Horton et al., 2001) and inferred in the Eastern Cordillera (DeCelles and Horton, 2003) is best explained as a result of forebulge migration, suggesting that at least part of the Upper Cretaceous–mid-Paleocene strata may be related to backbulge sedimentation.

Based on the information discussed above, initial Cretaceous–mid-Paleocene backbulge deposition followed by mid-Paleocene–middle Eocene forebulge deposition suggests eastward migration of part of the foreland basin system through the Altiplano and Eastern Cordillera (Horton and DeCelles, 1997). By inference, the adjacent fold–thrust belt to the west propagated eastward during Cretaceous–Paleocene time (Fig. 3A and B).

Correlative proximal foredeep and wedge-top deposits may be represented by Upper Cretaceous to Paleocene rocks of the Purilactis Group in eastern Chile (Mpodozis et al., 2000; Arriagada et al., 2002; Horton et al., 2001). These deposits are up to 5 km thick and contain growth structures in the basal Tonel Formation (Upper Cretaceous), and exhibit proximal alluvial facies in conjunction with progressive unconformities in the Eocene Loma Amarilla strata (Arriagada et al., 2000, 2002).

4.1.2. Structure

Eastward migration of the forebulge and backbulge depozones of a foreland basin system from the Altiplano region into the Eastern Cordillera implies that the front of the fold–thrust belt migrated 400 km. Possible structures from this early fold–thrust belt may be mostly concealed beneath the extensive Tertiary to Quaternary volcanic cover of the Western Cordillera. Support for shortening, as early as the Cretaceous to Paleocene, in the modern arc to forearc region (Fig. 1) is found in a growing body of structural, stratigraphic, and paleomagnetic evidence in Chile (Chong and Ruetter, 1985; Bogdanic, 1990; Hammerschmidt et al., 1992; Harley et al., 1992; Scheuber and Reutter, 1992; Mpodozis et al., 2000; Somoza et al., 1999; Arriagada et al., 2000, 2002, 2003; Roperch et al., 2000a, 2000b). Evidence of shortening and associated strike–slip deformation includes clockwise rotations of up to 65° in Mesozoic to Paleocene volcanic rocks (Arriagada et al., 2003), rapid cooling of the Cordillera Domeyko (in the Chilean Precordillera) between 50 and 30 Ma in conjunction with significant E–W shortening (Maksaev and Zentilli, 1999), and shortening-related growth structures in Cretaceous and Paleogene age sedimentary rocks (Arriagada et al., 2000, 2002).

Structures indicating Cretaceous through Eocene compression within the Precordillera and Cordillera Domeyko are tight to overturned folds, imbricated reverse faults, and thrust faults with >5 km of displacement (Chong, 1977; Scheuber and Reutter, 1992). Preliminary shortening estimates for the region are >25% (Scheuber and Reutter, 1992). The relationships between foreland basin migration and shortening and propagation of a fold–thrust belt (DeCelles and DeCelles, 2001) permit tentative estimates to be made of both shortening and propagation for this time period even though the structures to document that shortening may not be exposed or preserved. The distance of forebulge migration is equal to the sum of the total propagation of the fold–thrust belt (relative to a static marker on the Brazilian craton) plus the total shortening (DeCelles and DeCelles, 2001). If taken at face value, the stratigraphic record of backbulge and forebulge deposition suggests that from 70 Ma to 45

Ma, the foreland basin migrated approximately 400 km to the east (Fig. 3A and B). The propagation distance of this early Andean fold–thrust belt is partially constrained by the width of potential deformation in the Cordillera Domeyko and Western Cordillera. Taking the 400 km migration estimate, the current width (150–200 km) of these cordilleras suggests that shortening from 45 to 70 Ma was 200–250 km (50–60%) at 8–10 mm/year. Eastward advance of the fold–thrust belt 150–200 km over a 25 myr time span yields a propagation rate of 6–8 mm/year, a value similar to the long-term average propagation rate of 7.8 mm/year for the central Andean fold–thrust belt (its present width (550 km) divided by ~70 myr).

4.2. Late Eocene

4.2.1. Sedimentology

By late Eocene time, forebulge depositional conditions in the eastern Altiplano and Eastern Cordillera had been replaced by major sediment accumulation in large-scale fluvial systems. Several-km-thick successions of westerly derived fluvial sandstone and mudstone occupy both the Altiplano (main body of the Potoco Formation) and the Camargo syncline in the Eastern Cordillera (Camargo Formation). However, the broadly age-equivalent deposits in these two regions cannot represent a single depositional system because the Eastern Cordillera strata (Camargo Formation) are significantly coarser than the Altiplano (Potoco Formation) strata (Horton and DeCelles, 2001; Horton et al., 2001). Whereas the Altiplano deposits were derived from a source area in the present-day Western Cordillera (Horton et al., 2001), the Eastern Cordillera strata were derived from a separate source area in the central to western Eastern Cordillera (Horton and DeCelles, 2001). For the Altiplano, age control (discussed below) indicates rapid rates of sediment accumulation, and, by inference, rapid subsidence. The two separate zones of rapid accumulation probably represent foredeep depozones associated with two separate fold–thrust systems: one in the Western Cordillera and one in the central to western Eastern Cordillera.

The vertical stacking of foredeep deposits upon the underlying backbulge and forebulge deposits indicates continued eastward migration of the central Andean fold–thrust belt. The isolation and evolution of two separate yet coeval foredeeps suggest that a major forward advance of the frontal tip of the fold–thrust belt occurred during late Eocene time. This eastward “jump” in the thrust front effectively isolated the Altiplano basin so that it became a crustal-scale piggyback basin (Fig. 3C and D).

4.2.2. Structure

The relationship between the level of structural exposure within the Eastern Cordillera and the level of structural exposure within the internally drained Altiplano basin requires a 10–12 km structural step at this boundary (McQuarrie and DeCelles, 2001). Although previous interpretations of this step have included mid-Miocene normal faulting (Rochat et al., 1996; Baby et al., 1997; Rochat et al., 1999), McQuarrie and DeCelles (2001) and McQuarrie (2002) argue that this step is the result of a 10–12-km-thick basement megathrust propagating up and over a crustal-scale ramp (Fig. 2). This assertion is supported by the along-strike continuity of the offset in Paleozoic rocks, Paleozoic hanging wall and Tertiary footwall cutoff relationships between west-vergent structures in the Eastern Cordillera, and large amplitude synclines in Tertiary rocks, as well as seismic interpretations in the Poopo basin (McQuarrie and DeCelles, 2001).

The interpretation of this step as the result of a thrust ramp implies that the overlying Paleozoic through Tertiary rocks were passively folded into a monocline as the basement thrust cut up-section eastward beneath what is now the Eastern Cordillera. The eastward emplacement of the basement thrust sheet was accommodated by both east- and west-vergent, thin-skinned deformation in the Paleozoic cover (Fig. 2). Insertion of the basement thrust sheet beneath the Eastern Cordillera was the means by which deformation was allowed to propagate or “jump” eastward into the Eastern Cordillera, dividing the foreland basin into two distinct basins (Fig. 3C and D). Timing for this eastward propagation of the fold–thrust belt is provided by apatite and zircon fission track and $^{40}Ar/^{39}Ar$ thermochronologic data that record cooling ages and suggest initial shortening-related exhumation in the Eastern Cordillera at ~40±5 Ma (Benjamin et al., 1987; McBride et al., 1987; Farrar et al., 1988; Sempere et al., 1990; Masek et al., 1994; Kennan et al., 1995, Lamb et al., 1997; Ege et al., 2001). We propose that the shortening-related exhumation tracks the progression of Eastern Cordillera rocks up and over the basement ramp.

4.3. Oligocene

4.3.1. Sedimentology

Oligocene rocks represent the continued development of separate depocenters within the Altiplano and eastern Eastern Cordillera. Deposition is recorded by the 2–7-km-thick Potoco Formation in the Altiplano, and the 2-km-thick Camargo Formation in the Camargo syncline (Horton et al., 2001; Horton and DeCelles, 2001, DeCelles and Horton, 2003). For the Altiplano, long-term average sediment accumulation rates for the late Eocene–Oligocene Potoco Formation are as high as 500 m/myr, a value similar to the highest accumulation rates anywhere in the Subandean foreland (Jordan, 1995). We attribute this rapid accumulation to flexural subsidence in a basin influenced by dual topographic loading in both the Western Cordillera and Eastern Cordillera (Horton et al., 2002). Although age control is very limited for the Camargo syncline, similar rapid subsidence is inferred for the eastern Eastern Cordillera. A back-bulge depozone in this Oligocene foreland basin system is predicted to have been located in the present-day Subandean zone (Fig. 3D and E). In the Subandean zone, a several-km-thick, upward-coarsening fluvial succession rests disconformably on Cretaceous strata (Dunn et al., 1995; Jordan et al., 1997) without any evidence for backbulge deposits. The lowest portion of this section is probably late Oligocene in age (Jordan et al., 1997; Marshall et al., 1993; Marshall and Sempere, 1991). Possible explanations for missing backbulge deposits include damping out of the flexural wave due to increasing rigidity of the Brazilian shield to the east or erosion of backbulge strata during forebulge migration (DeCelles and Horton, 2003). Reconstruction of the fold–thrust belt indicates that the western limit of the modern Brazilian shield (as defined through geophysical studies) was near the eastern limit of thrust front by this time (Fig. 3D and E).

Within deformed regions of the central Eastern Cordillera, several zones of sediment accumulation became active during the Oligocene. Narrow sedimentary basins of the Tupiza region are associated with fold–thrust deformation as early as Oligocene time (Horton, 1998). These basins are considered the most proximal zone of sediment accumulation of the Oligocene foreland basin system. Growth strata are present in the Tupiza area basins (Horton, 1998) but lacking in the Camargo syncline (Horton and DeCelles, 2001; DeCelles and Horton, 2003), suggesting that the eastern front of the Andean fold–thrust belt was situated within the central Eastern Cordillera by this time (Fig. 3E).

4.3.2. Structure

The accruing slip on the proposed basement megathrust was initially fed eastward into the east-verging structures of the Eastern Cordillera, but a significant portion of that slip was fed westward along the westward verging backthrust belt (Fig. 3D and E) (McQuarrie and DeCelles, 2001; McQuarrie, 2002). Cross-section balancing and the sequential restoration of the fold–thrust belt suggest that both east- and west-verging structures in the southern portion of the central Andean fold–thrust belt were active simultaneously (McQuarrie, 2002; Müller et al., 2002). The eastward propagation of the proposed basement thrust sheet created an east- and westward widening zone of shortened and thickened (up to 60 km) crust (McQuarrie, 2002). Timing constraints for this period of deformation include: the formation of basins associated with west-verging faults (Horton, 1998; Horton et al., 2002), overlap sediments that provide a minimum age bracket on deformation (Kennan et al., 1995; Hérail et al., 1996; Tawackoli et al., 1996; Horton, 1998, 2005), and cooling ages of minerals within the Eastern Cordillera (Ege et al., 2001). The wedgetop Tupiza basins described above suggest that activity in the backthrust belt initiated at ~34 Ma. These west-verging faults are capped by 20–29 Ma volcanic and sedimentary rocks (Hérail et al., 1996; Tawackoli et al., 1996; Horton, 1998). Preliminary apatite fission track ages from the Eastern Cordillera at approximately 21° S suggest cooling ages of 38±7 Ma to 30±5 Ma near the transition from east-verging to west-verging thrusts, with ages decreasing to 17±5 Ma at the western edge of the backthrust belt and the eastern edge of the Eastern Cordillera zone (Ege et al., 2001). Thus, the structural, stratigraphic, and thermochronologic data all indicate that late Eocene through

Oligocene time was a period of major structural growth and crustal thickening within the Eastern Cordillera.

4.4. Early Miocene

4.4.1. Sedimentology

Coarse-grained strata in the eastern Altiplano and western Subandean zone record the forward propagation of two fold–thrust systems, the west-vergent backthrust belt of the western Eastern Cordillera, and the east-vergent Interandean zone. In the eastern Altiplano, the latest Oligocene–earliest Miocene tuffs are overlain by several-km-thick conglomeratic sections; e.g., the Peñas, Coniri, Tambo Tambillo, and possibly San Vicente Formations (Baby et al., 1990; Sempere et al., 1990; Kennan et al., 1995). Available paleocurrent data indicate transport from east to west (Sempere et al., 1990; Horton et al., 2002). These successions tend to coarsen upward and in one case exhibit growth strata in their upper levels (LaReau and Horton, 2000). We attribute these strata to deposition in a crustal-scale piggyback basin in the Altiplano that was subjected to progressively greater deformation as it was incorporated into the westward-propagating, west-vergent backthrust belt.

In the western Subandean zone, the thick-bedded sandstone of the 1–2-km-thick, lower Miocene (24.4 Ma) Tariquia Formation (Erikson and Kelley, 1995; Jordan et al., 1997) suggests erosion of the westward adjacent Interandean zone. By early Miocene time, the leading edge of the east-vergent orogenic wedge had propagated approximately to the boundary between the Interandean and Subandean zones, creating a foredeep depozone in the modern Subandean zone. In the Tupiza region of the central Eastern Cordillera, continued wedge-top deposition occurred synchronous with fold–thrust deformation (Hérail et al., 1996; Horton, 1998) (Fig. 3F).

4.4.2. Structure

The structures within the backthrust zone on the eastern side of the Eastern Cordillera display older units and deeper exposures in the western part of the zone. The folds and faults tend to verge westward and regional structures cut up-section to the west incorporating progressively younger rocks (Fig. 2). The style of deformation within the backthrust zone is similar for hundreds of kilometers along strike. Duplexes in the northern portion (17–18° S) of the backthrust zone require that at least in those sections, the thrusts broke in sequence from east to west (McQuarrie and DeCelles, 2001). The combination of these features provides strong support for a westward-verging, westward-younging backthrust system that occupies a significant part of the central Andean fold–thrust belt (McQuarrie and DeCelles, 2001). Because the proposed basement megathrust carried cover rocks up and over the basement ramp and into the westward-propagating backthrust belt, the cessation of deformation within the backthrust zone marks the cessation of slip on the basement megathrust. Fold–thrust structures in the backthrust belt are overlapped by the gently folded sedimentary rocks of the 24–21 Ma Salla beds (Sempere et al., 1990), indicating that most of the structural activities within the backthrust belt ceased at ~20 Ma (McFadden et al., 1985; Sempere et al., 1990; Marshall and Sempere, 1991, McQuarrie and DeCelles, 2001). Intermontane basins in the Tupiza area contain growth structures associated with east-vergent, out-of-sequence thrust faults that were active between ~18 and ~13 Ma (Horton, 1998), suggesting that minor amounts of motion on the basement megathrust and deformation within the backthrust zone persisted until mid-Miocene time (~13 Ma) (Fig. 3F). The 20–13 Ma minimum age for deformation within the backthrust belt has significant implications for the bracketing of deformation within the Interandean zone. Because both zones are associated with slip on the basement megathrust sheet (McQuarrie, 2002), the ~20 Ma age represents the end of major deformation in the Interandean zone with minor amounts of deformation ending at ~13 Ma. Preliminary apatite fission track dates within the Interandean zone range in age from 19±3 Ma to 13±3 Ma (Ege et al., 2001). These ages may reflect uplift of the Interandean zone along a lower basement thrust that fed slip into the Subandean zone.

We suggest that basin fill preserved in the Altiplano (Salla beds, Luribay, Tambo Tambillo, Coniri conglomerates) and western Subandean zone (Tariquia Formation) that have been used as evidence marking the initiation of mountain building within the Central Andes (~25 Ma) (Sempere et al., 1990) may instead be associated with the end of a period of significant shortening and thickening related to motion on an extensive basement megathrust. To conserve mass, the magnitude of shortening in the lower crust must match that of the upper crust. If lower crustal deformation had the same lateral boundaries as the brittle upper crust then by mid-Miocene time, the Eastern Cordillera was characterized by thick crust (~60 km) and possibly high elevation (3–4 km) (McQuarrie, 2002). At the southern end of the Altiplano, the chemistry of late Oligocene to early Miocene lavas suggests a thick crust (>50 km) beneath what is now the western edge of the plateau (Kay et al., 1994; Allmendinger et al., 1997).

4.5. Late Miocene

4.5.1. Sedimentology

By late Miocene time, the foreland basin system had propagated far to the east, close to its present-day configuration (Fig. 3G). In the Subandean zone, thick intervals of coarse-grained strata overlie the dated Oligocene–lower Miocene deposits (Gubbels et al., 1993; Moretti et al., 1996; Jordan et al., 1997). Although these deposits are mostly representative of foredeep depositional conditions, growth strata imaged in seismic profiles (Roeder and Chamberlain, 1995; Baby et al., 1995; Horton and DeCelles, 1997) suggest wedge-top conditions for at least part of the Subandean zone by late Miocene time (Fig. 3G).

In the Eastern Cordillera, the development of a regional low-relief geomorphic surface, the San Juan del Oro surface, is dated as ~10 Ma based on tuffs that overlie the surface (Servant et al., 1989; Gubbels et al., 1993). This surface is defined as an erosional surface that cuts Paleozoic bedrock in some areas, but a constructional surface capping Miocene basins in other areas (Gubbels et al., 1993; Kennan et al., 1997). Because this surface is not significantly tilted, most deformation in the Eastern Cordillera must have been complete by late Miocene time (Gubbels et al., 1993). In the Altiplano, sedimentary filling of the closed Altiplano basin continued (Horton et al., 2002). Late Miocene sedimentation predominantly involved deposition of volcanogenic units and low-gradient fluvial systems carrying volcaniclastic debris (Evernden et al., 1977; Lavenu et al., 1989; Marshall et al., 1993).

4.5.2. Structure

The Subandean zone between 18° S and 20° S is interpreted to be an in-sequence, thin-skinned thrust system detached along lower Silurian shale (Fig. 2) (Baby et al., 1992; Dunn et al., 1995; McQuarrie, 2002). The detachment for the Subandean zone dips 2° toward the west (Baby et al., 1992, Dunn et al., 1995), well below the décollement of the tightly deformed structures in the Interandean zone (Kley, 1996; Kley et al., 1996; McQuarrie, 2002). To resolve the discrepancy between the two detachment horizons, Kley (1996) proposed that a single basement thrust sheet, which fed slip into the Subandean zone, could fill this space. Evidence for basement involvement in this portion of the fold–thrust belt includes abrupt changes in structural elevation (Kley, 1996), gravimetric and magnetotelluric data (Kley et al., 1996; Schmitz and Kley, 1997), and a large change in gravity correlated with a seismic refraction discontinuity (Wigger et al., 1994; Dunn et al., 1995).

Timing of motion on the lower basement thrust sheet, and thus initiation of deformation within the Subandean zone, has generally been thought to be late Miocene (10 Ma and younger) based on several features correlated to Subandean shortening. These features include the age of the San Juan del Oro surface in the Eastern Cordillera (Gubbels et al., 1993; Kennan et al., 1997), the ~10 Ma age of conformable sedimentary rocks in the eastern Subandean zone (Gubbels et al., 1993), and apatite fission track ages suggesting increased exhumation in the Eastern Cordillera from 15 to 7 Ma (Benjamin et al., 1987; Masek et al., 1994). The overlap of structures within the backthrust belt by ~20 Ma synorogenic sediments limits significant motion on the upper basement thrust sheet after this time and suggests that deformation within the Subandean zone may have begun as early as 20 Ma. Pre-10 Ma deformation is supported by preliminary apatite fission track ages (19±3 Ma to 13±3 Ma) within the Interandean zone (Ege et al., 2001). These ages most likely reflect a combination of deformation within the Interandean zone and passive uplift of the zone along a lower basement thrust that fed slip into the Subandean zone. Although some overlap in deformation along the upper basement thrust and lower basement thrust may be expected, we propose that the most of the slip was transferred from the upper basement thrust to the lower basement thrust and into the Subandean zone by ~15 Ma.

Shortening within the Subandean zone occurred coevally with deformation in the hinterland of the fold–thrust belt from ~15 Ma to the present (Kennan et al., 1995; Lamb and Hoke, 1997). Examples of hinterland deformation include: (1) growth structures in synorogenic sediments imaged on seismic lines west of the Corque syncline that show west-directed thrusting between 25 and 5 Ma (Lamb and Hoke, 1997; McQuarrie and DeCelles, 2001), and (2) a pronounced angular unconformity on the east limb of the Corque syncline (Lamb and Hoke, 1997), which suggests major folding of basin sediments post-9 Ma.

4.6. Rates of deformation

The evolution presented above suggests that there was 200–250 km of shortening in a narrow Late Cretaceous to Paleocene fold–thrust belt and an additional 300–330 km of shortening accommodated by basement megathrusts and cover rocks as the fold–thrust belt propagated eastward into the Eastern Cordillera and Subandean zone. By combining structural geology with thermochronology and the basin migration history, it is possible to determine average, long-term rates of shortening through the fold–thrust belt (Fig. 3). Accurate timing constraints are essential to any proposed rates of deformation across the Andes. The critical ages that bracket deformation of the Andean fold–thrust belt are as follows: (1) coexisting wedge-top deformation in eastern Chile with backbulge sedimentation in the eastern Altiplano at 65–70 Ma (Horton et al., 2001; Arriagada et al., 2002, 2003); (2) 40±5 Ma cooling ages in the western part of the Eastern Cordillera (Benjamin et al., 1987; McBride et al., 1987; Farrar et al., 1988; Sempere et al., 1990; Masek et al., 1994; Kennan et al., 1995, Ege et al., 2001), interpreted to represent the cooling age of rocks moving over a basement ramp (McQuarrie and DeCelles, 2001); and (3) capping of deformation within the backthrust belt (and by correlation Eastern Cordillera and Interandean zone) by synorogenic sedimentary rocks between 20 and 15 Ma (McFadden et al., 1985; Sempere et al., 1990; Marshall and Sempere, 1991; Horton, 1998; McQuarrie, 2002). Although imprecise, these timing constraints can be combined with sequentially restored cross sections describing the kinematic evolution of the fold–thrust belt (McQuarrie, 2002) to provide broad rates of deformation through the Cenozoic. Although the ages bracketing each of the panels on Fig. 3 are inexact, they represent general limits on the kinematic history within 5–10 myr. Comparing rates of deformation across 20–30 myr time intervals reveals that shortening rates in the Central Andes have remained fairly constant (~8–10 mm/year) through time with possible slowing (~5–7 mm/year) in the last 15–20 myr.

5. Modern lithospheric structure

The combined shortening of a narrow Late Cretaceous to Paleocene fold–thrust belt, in addition to documented shortening within the Eastern Cordillera and Subandean zone, is more than sufficient to build the 70-km-thick crust of the central Andean plateau and suggests a similar magnitude of shortening within, or removal of, the mantle lithosphere. Seismic images of the mantle lithosphere under the Central Andes indicate that the mantle is not anomalously thick, nor has it been completely removed, suggesting a mantle deformation process that has been continuous through time (Fig. 4). Because understanding the modern lithospheric structure is essential to understanding any lithospheric deformation processes in the Andes, we discuss five major regions from the Western Cordillera to the foreland basin in terms of their crustal thickness, composition, and mantle lithospheric characteristics.

5.1. Western Cordillera

Crustal thickness variations for the Western Cordilleran arc have been investigated using a variety of geophysical analyses (Myers et al., 1998; Graeber and Asch, 1999; Baumont et al., 2001, 2002; Beck and Zandt, 2002). Receiver function analysis at the Western Cordillera/Altiplano boundary suggests a crustal thickness of 70 km. This thickness corresponds to a P_s conversion at 9.8 s (Beck and Zandt, 2002) and assumes a uniform felsic crust as indicated by the bulk crustal velocity of 6.0 km/s and V_p/V_s ratio of 1.73 constrained by Swenson et al. (2000). However, a

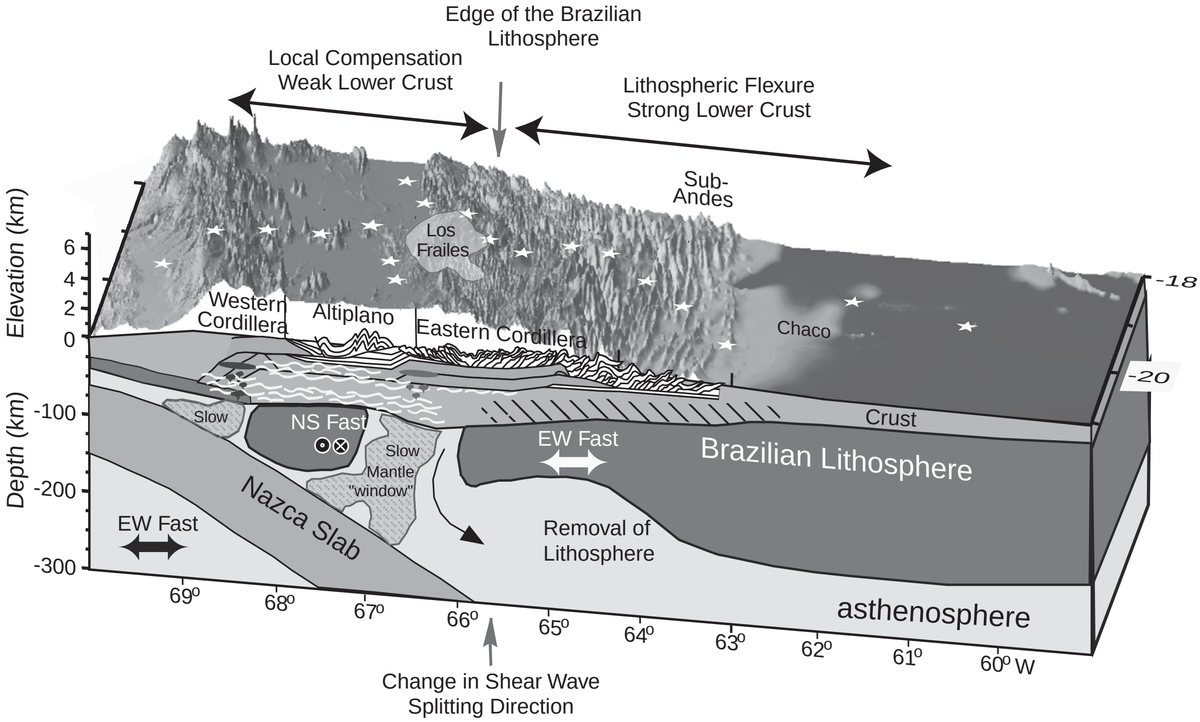

Fig. 4. Lithospheric-scale cross-section for 18–20° S showing our interpretation of the crustal and mantle structure for the Central Andes. White stars are BANJO/SEDA stations (Beck and Zandt, 2002). Dark gray lithosphere reflects fast upper mantle P wave velocities and white to gray shaded mantle represents slow P wave velocities. Crustal low-velocity zones are shown in white waves. Low-velocity zones possibly associated with crustal melt are shown in black for the Western Cordillera and dark gray for Los Frailes volcanic complex. Crustal structure from balanced cross section (2-D MOVE) (modified from McQuarrie, 2002): basement megathrusts in gray; sedimentary cover rocks in white. Note the spatial overlap of the mantle lithospheric window, the crustal-scale ramp, and the elevation steps associated with the basement thrust sheets.

more local tomography study of the coastal region to the Western Cordillera, farther south at 22–23° S, found a thick (70 km), two-layer crust with an upper crustal P wave velocity of 6.0 km/s and a lower crustal P wave velocity of 7.0 km/s (Graeber and Asch, 1999). Seismic refraction studies (Wigger et al., 1994) also show a higher bulk V_p through the Western Cordillera, perhaps reflecting processes related to arc magmatism.

Shallow mantle structure as determined by mantle tomography studies also highlights the importance of arc-magmatic processes under the Western Cordillera (Myers et al., 1998; Graeber and Haberland, 1996). Arc processes are indicated by localized zones of low V_p and V_s. Despite good correlation between low seismic velocities and arc volcanism, Myers et al. (1998) found that the single mantle parameter that corresponds best with arc volcanism is high V_p/V_s which they interpreted to result from the presence of subduction related fluids (Geise, 1996; Myers et al., 1998).

5.2. Altiplano

Crustal composition, crustal thickness, and mantle lithospheric characteristics are all well known for the Altiplano in the region of 18–20° S. Several geophysical studies have documented the presence of a thick (60–75 km) crust with much lower than average crustal P wave velocities (~6.0 km/s) beneath the Altiplano (Schuessler, 1994; Beck et al., 1996; Zandt et al., 1996; Myers et al., 1998; Swenson et al., 2000). Regional waveform modeling of intermediate and shallow earthquakes by Swenson et al. (2000), receiver function analysis (Beck and Zandt, 2002), and surface wave dispersion measurements (Baumont et al., 2002) indicate crustal thickness values of 59–64 km under the central Altiplano with average crustal V_p of 5.80–6.25 and a Poisson's ratio of 0.25. None of the different techniques provides any evidence for a higher velocity lower crust that is typical in many continental regions and might have been present before the Andean orogeny (Swenson et al., 2000; Baumont et al., 2002; Beck and Zandt, 2002).

Receiver function analysis shows a small negative-polarity arrival in many of the receiver functions at 2.5–3.0 s. This corresponds to a mid-crustal low-velocity zone that can be traced across the entire width of the Altiplano and corresponds to a depth of 14–30 km (Yuan et al., 2000). The basal décollement for the fold–thrust belt is too deep to be located at the top of this low-velocity zone. Geometrically, the upper limit of the basal décollement is ~30 km, which correlates with the base of the Altiplano low-velocity zone (Figs. 2 and 4). There is no lower limit on the depth to the basal décollement (e.g., Müller et al., 2002).

Myers et al. (1998) reported a high V_p as the dominant feature of the shallow mantle under the central Altiplano. High V_p and V_s and moderately high Q in the shallow Altiplano mantle are used to argue for the presence of mantle lithosphere that extends at least to 125–150 km depth.

5.3. Eastern Cordillera

Significant changes in crustal thickness and mantle lithosphere structure occur between the Altiplano and the Eastern Cordillera. Receiver function analysis shows generally decreasing crustal thickness with decreasing elevations across the Eastern Cordillera (Beck and Zandt, 2002). Relationships between crustal composition (as indicated by velocity) and crustal thickness determined from receiver function analysis, waveform modeling, and seismic refraction all indicate that the crust of the Eastern Cordillera is thick (60–74 km) and slow (5.75–6.0 km/s) (Wigger et al., 1994; Beck et al., 1996; Swenson et al., 2000; Beck and Zandt, 2002). Seismic stations used for waveform modeling and receiver function analysis for the western Eastern Cordillera are located above the massive Los Frailes ignimbrite field (Beck et al., 1996; Myers et al., 1998; Swenson et al., 2000, Beck and Zandt, 2002) (Fig. 4). Although the receiver functions are complicated, they can be modeled with a low-velocity layer at 15–20 km depth and a crustal thickness of 74 km. The areal extent of the low-velocity zone correlates strongly with the aerial extent of the volcanic field. Although there is strong evidence for thick crust in this area, receiver functions vary as a function of event backazimuths, indicating that they are sampling different Moho depths at different locations. This suggests local topography on the Moho at the Eastern Cordillera/Altiplano boundary (Beck et al., 1996; Swenson et al., 2000; Beck and Zandt, 2002).

East of Los Frailes volcanic field, receiver function analysis indicates gradually decreasing crustal thicknesses (60–50 km) from ~65° to 64° W. These stations show low-velocity zones at ~30 km, which is substantially deeper than the low-velocity zones associated with Los Frailes volcanic field (Beck and Zandt, 2002). However, the depth of these low-velocity zones is consistent with the detachment horizon for the lower (Subandean) basement mega-thrust, which separates upper crustal brittle shortening from lower crustal ductile shortening in this region (McQuarrie, 2002) (Fig. 2).

Mantle tomography images in the Eastern Cordillera indicate that there is a very low-velocity anomaly that extends through the lithosphere and is centered on the Los Frailes ignimbrite field. This suggests that the volcanism associated with this massive ignimbrite field is rooted in the mantle and that the lithosphere in this location has been strongly altered or removed (Myers et al., 1998). East of the Los Frailes anomaly, shallow mantle V_p and V_s values increase to slightly greater than normal values, with Q increasing dramatically (Myers et al., 1998). This transition to high velocities and high Q is in agreement with other geophysical features that also suggest the presence of strong, cold lithosphere indicative of the Brazilian shield. These include strong lateral changes in upper mantle velocities in the region of 16° S (Dorbath and Granet, 1996), a change in the fast direction of shear-wave anisotropies from N–S under the Altiplano to E–W at 65.5° W (Polet et al., 2000) and Bouguer gravity anomalies showing that lithospheric flexure is supporting the Subandean zone and part of the Eastern Cordillera (Lyon-Caen et al., 1985; Watts et al., 1995; Whitman et al., 1992).

5.4. Subandean zone

Less is known about the average crustal velocities and Poisson's ratio of the Subandean crust, making determination of crustal thickness more difficult. However by using the low average velocities (5.9 km/s) recorded in the seismic refraction study by Wigger et al. (1994), Beck and Zandt (2002) proposed values of 42 km and 40 km for crustal thicknesses in the eastern Subandean zone. The low average velocity is in part a result of the low-velocity Paleozoic, Mesozoic, and Tertiary sedimentary rocks involved in the fold–thrust belt (Wigger et al., 1994; Beck and Zandt, 2002).

5.5. Chaco Plain

The brief operating time of the seismic stations on the Chaco Plain provided much less data to determine crustal composition and thickness. The data are best fit with average crustal velocities of 6 km/s, giving crustal thicknesses of 30 km (Beck and Zandt, 2002). Snoke and James (1997) also found thin crust (32 km) in the Chaco farther to the east in Brazil based on surface wave group velocities. However, because both the Paleozoic basin and the foreland basin thin dramatically to the east where the stations were located (Dunn et al., 1995; Sempere, 1995), the 30 km crust most likely reflects the thickness of the pre-Paleozoic basement.

6. Lithospheric evolution

The modern lithospheric structure is shown in Fig. 4. The cross section of the fold–thrust belt indicates that the upper crust has shortened 330 km. Under the thickened crust of the fold–thrust belt is the strong, seismically fast mantle lithosphere of the Brazilian shield (eastern side) and an adjacent mantle "window" of thermally or chemically altered lithosphere (western side) (Lyon-Caen et al., 1985; Myers et al., 1998; Beck and Zandt, 2002) (Fig. 4). The strong/fast nature of the Brazilian lithosphere argues against the mantle lithosphere under the fold–thrust belt being orogenically thickened, while the mantle window suggests a possible location of mantle removal (Beck and Zandt, 2002). The total length of mantle that needs to be removed to produce the current lithospheric cross section of the Andes is ~400 km (330 km shortening plus a ~75-km-wide mantle lithosphere window). The cross-sectional area of removed lithosphere could range from 2×10^4 km^2 for an initially thin (50 km) mantle lithosphere to 4.8×10^4 km^2 for a thick (120 km) mantle lithosphere. Fig. 4 also illustrates the spatial co-existence of the crustal-scale ramp at the Eastern Cordillera/Altiplano boundary and the zone of mantle weakness. We suggest that the correlation between these crustal and lithospheric scale features gives insight into lithospheric-scale deformation processes within the Central Andes. We propose the following scenario as a means of linking zones of significant crustal and mantle deformation.

6.1. Focusing deformation 600 km from the trench

During the early 70–45 Ma deformation period, the fold–thrust belt and subsequent mantle deformation were most likely focused adjacent to the magmatic arc ~150–200 km from the subduction zone. The proximity of the fold–thrust belt to the trench and volcanic arc would have allowed for continual removal of mantle lithosphere through thermal or mechanical (subduction ablation) processes. However, at ~40 Ma deformation in the crust, and perhaps the mantle jumped ~400 km inboard. We propose that this jump may have exploited a pre-existing weakness in the crust and or mantle due to rifting in Ordovician, Triassic/Jurassic, and Cretaceous time (Viramontes et al., 1999; Sempere et al., 2002), and/or flat slab subduction hydrating and weakening of the lithosphere (Sandeman et al., 1995; James and Sacks, 1999).

Several periods of lithospheric rifting affected the Central Andes of Bolivia: during Late Cambrian/Early Ordovician (Bahlburg, 1990; Sempere, 1995), from Permian through mid-Jurassic (Sempere et al., 2002), and Late (120–100 Ma) Cretaceous (Viramontes et al., 1999). The Permo-Triassic rift axis in northern Bolivia and the Cretaceous rift axis in southern Bolivia, as determined by the localization of syn-rift sedimentary, intrusive, and volcanic rocks, are along the axis of the Eastern Cordillera nearly coincident with the marked change in vergence of east- and west-verging faults (Viramontes et al., 1999;

Sempere et al., 2002). The crustal-scale basement ramp that facilitated the "jump" of deformation from the Western Cordillera into the Eastern Cordillera restores below the location of the proposed rift axis in balanced cross sections of the fold–thrust belt (McQuarrie, 2002), suggesting that the location of the ramp could be controlled by pre-existing crustal structures.

The angle of subduction of the Nazca plate seems to be a transient feature through the Andes and has been proposed as a mechanism that controls deformation not only under the current flat slab segments of the Andean arc but also the style and location of deformation in the past (Isacks, 1988; Kay et al., 1995; Sandeman et al., 1995; Kay and Abbruzzi, 1996; Allmendinger et al., 1997; James and Sacks, 1999). Flat slab initiation during the Eocene may be supported by the cessation of Western Cordillera arc volcanism at ~50 Ma in southern Peru to ~38 Ma in northern Chile with accompanying mantle hydration along the Eastern Cordillera (Sandeman et al., 1995; James and Sacks, 1999). Flat slab subduction from 50 to 38 Ma would allow fluids escaping from the subducting slab to weaken the overriding plate, and create a zone where deformation is focused first in the mantle and then in the crust.

6.2. Concurrent mantle and crustal deformation

Early (40 Ma) deformation within the mantle lithosphere may have been critically important in determining both where and how deformation was accommodated in the overlying crust. Westward intracontinental "subduction" of the mantle lithosphere has sufficient negative buoyancy to create space for the basement megathrust to propagate eastward (McQuarrie and Lavier, 2003) (Fig. 5). The westward-subducting lithosphere could have been removed systematically through continual mantle ablation from 40 Ma to present (Pope and Willett, 1998), or significant portions of mantle lithosphere could have interfered with the subducting Nazca plate, causing pronounced delamination events (Kay and Abbruzzi, 1996). Perhaps a signal of significant mantle melting at depth or significant lithospheric removal is the pulse of basaltic to dacitic volcanism with island arc affinities that occurred throughout the central Altiplano at ~23–24 Ma (Davidson and de Silva, 1993; Hoke et al., 1993; 1994, Kennan et al., 1995; Lamb and Hoke, 1997). If this pulse of mafic magmatism and the crustal melt ignimbrites that followed (12–7 Ma) are a result of delamination of the lithosphere, this may suggest westward subduction of the mantle lithosphere under the fold–thrust belt from 40 to 25 Ma with delamination of the slab at ~25 Ma due to mechanical or thermal instabilities. The continued subduction of the Brazilian shield under the fold–thrust belt from ~30 Ma to present argues for continual removal of mantle lithosphere focused at the Eastern Cordillera/Altiplano boundary. Either ablative subduction, or piecemeal delamination of unstable lithospheric blocks initiating between ~40 and 25 Ma and continuing to the present, is in good agreement with both the timing of volcanism in the

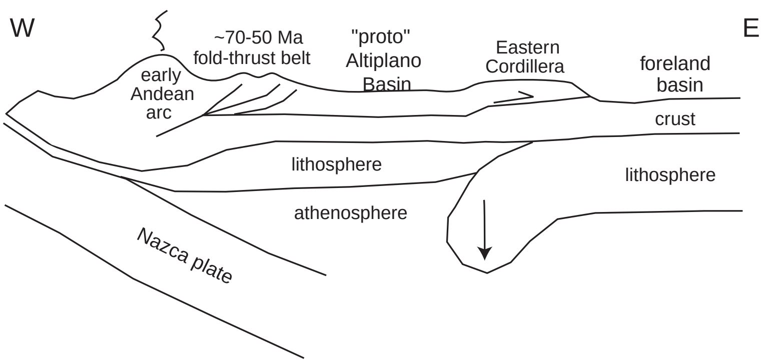

Fig. 5. Cartoon drawing of mantle and crustal deformation in the central Andes at ~35 Ma. A negatively buoyant mantle lithosphere provides the space needed for the eastward propagation of the basement megathrust (McQuarrie and Lavier, 2003).

Eastern Cordillera and the Altiplano and the current mantle structure at this boundary (Myers et al., 1998) (Fig. 4).

7. Summary

Combining the history of foreland basin migration with palinspastically restored regional cross sections across the Bolivian Andes between 18° and 20° S argues for an eastward-migrating fold–thrust belt/foreland basin system possibly as early as the Late Cretaceous. Addition of 40–45 myr to the long held 25–30 Ma age for initiation of "Andean deformation" has important implications for shortening amounts, rates of shortening, and the elevation history of the central Andean plateau. The 70–50 Ma history of the Andean fold–thrust belt, as constrained by the migration of the foreland basin, suggests 200–250 km of shortening in the Late Cretaceous to Paleocene fold–thrust belt. Balanced regional cross sections across the central Andean plateau from the eastern edge of the volcanic arc to the foreland suggest an additional 330 km of shortening of the brittle upper crust and equal amounts of shortening in the more ductile mid- and lower crust (McQuarrie, 2002). The eastward propagation of a 15-km-thick basement thrust sheet allowed for deformation to jump from the Western Cordillera into the Eastern Cordillera at ~40 Ma, linking these two zones of shortening into an eastward-evolving Andean system. With an initial crustal thickness of 35–40 km, the combination of these two estimates (530–580 km) can more than account for the cross-sectional area of the Central Andean plateau in Bolivia.

The combination of basin migration, balanced cross sections, and thermochronology provides reasonable estimates of the average long-term rates of shortening within the Central Andes. Long-term average rates of shortening of the fold–thrust belt indicate that shortening has remained fairly constant (8–10 mm/yr) through time with possibly slower rates (5–7 mm/yr) over the last 15–20 myr. The combination of basin migration, balanced cross sections, and thermochronology also suggests that the area defined as the central Andean plateau was shortened, thickened, and perhaps elevated as early as 20 Ma, with minor amounts of additional thickening (7–10 km) due to subduction of the Brazilian shield and shortening within the Subandean zone.

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