Revised deformation history of the central Andes: Inferences from Cenozoic foredeep and intermontane basins of the Eastern Cordillera, Bolivia

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1. Introduction

[1] Investigation of Cenozoic deformation, sediment accumulation, and provenance in the Eastern Cordillera of Bolivia at 17-21°S indicates major shortening (60-140 km) and foredeep development followed by limited internal shortening and intermontane basin development. Contrasting histories of shortening, deposition, sediment dispersal, and detrital composition distinguish a formerly extensive Paleogene foredeep exposed along an eastern belt of synclines from a zone of principally Neogene intermontane basins in the central Eastern Cordillera. New ⁴⁰Ar/³⁹Ar ages of 25–17 Ma for interbedded tuffs reveal latest Oligocene-early Miocene accumulation in intermontane basins at rates <80 m/Myr, several times lower than Andean foredeeps. Although poorly dated, foredeep evolution probably coincided with middle Eocene-Oligocene deformation and denudation in adjoining regions to the west. A mid-Cenozoic transition from foredeep to intermontane conditions may be attributable to emplacement of a basementinvolved tectonic wedge beneath the Eastern Cordillera. In this interpretation, wedge emplacement drove flexural foredeep subsidence from roughly 40 to 25 Ma, whereas subsequent accumulation occurred in localized internally drained basins in elevated intermontane areas. Regardless of the regional subsurface structural geometry, ⁴⁰Ar/³⁹Ar ages define a 25-21 Ma onset of intermontane sedimentation that signifies the termination of major upper crustal shortening over large parts of the Eastern Cordillera and possibly the age of initial shortening in the Interandean and Subandean zones to the east. In contrast to several popular models, the revised tectonic history presented here suggests significant pre-Neogene shortening in the Eastern Cordillera and underscores the large uncertainties in estimates of longterm shortening rates in the Andes. Citation: Horton, B.K. (2005), Revised deformation history of the central Andes: Inferences from Cenozoic foredeep and intermontane basins of the Eastern Cordillera, Bolivia, Tectonics, 24, TC3011, doi:10.1029/2003TC001619.

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[2] The central Andes exemplify the processes and rates of mountain building along ocean-continent convergent plate boundaries. Many investigations of retroarc deformation, crustal thickening, plateau uplift, and foreland basin generation have emphasized the importance of horizontal shortening in driving Andean orogenesis [Isacks, 1988; Roeder, 1988; Sheffels, 1990; Schmitz, 1994; Wigger et al., 1994; Beck et al., 1996; Allmendinger et al., 1997; Lamb and Hoke, 1997]. Recently, GPS studies have provided an opportunity to compare modern displacements to geologic rates of shortening [Norabuena et al., 1998; Horton, 1999; Liu et al., 2000; Lamb, 2000; Bevis et al., 2001; Hindle et al., 2002; Khazaradze and Klotz, 2003]. These comparisons, however, arrive at conflicting conclusions about whether shortening rates have increased, decreased, or remained steady during Cenozoic deformation. Meaningful assessments are precluded by disagreement over the magnitude and duration of long-term shortening, with estimates ranging from 50 to 500 km over the past 10 to 70 Myr [Isacks, 1988; Sempere et al., 1990; Gubbels et al., 1993; Jordan et al., 1997; Kley and Monaldi, 1998; McQuarrie, 2002a, 2002b; DeCelles and Horton, 2003; McQuarrie et al., 2005].

[3] No consensus exists for the deformation history of Bolivia, the region of maximum Andean shortening. Although shortening in the thin-skinned Subandean Zone at the front of the fold-thrust belt appears well understood [Dunn et al., 1995; Roeder and Chamberlain, 1995; Baby et al., 1997], debate continues over the magnitude, geometry, and timing of deformation in the Eastern Cordillera, a hinterland region containing complex basement-involved structures [Kley et al., 1997; Allmendinger and Zapata, 2000: McOuarrie and DeCelles, 2001]. Inconsistencies in the regional kinematic history can be overcome by structural and chronostratigraphic investigations of Cenozoic outcrop belts distributed throughout the Eastern Cordillera (Figure 1). These outcrop belts, traditionally attributed to accumulation in isolated intermontane basins [Ahlfeld and Branisa, 1960; Martinez, 1980; Marshall and Sempere, 1991], may in part represent relict foredeeps analogous to the modern Andean foreland basin system [e.g., Jordan, 1995; Horton and DeCelles, 1997].

[4] The objective of this paper is to synthesize the evolution of foredeep and intermontane basins affiliated with deformation in the Eastern Cordillera of the central Andean fold-thrust belt at $17-21^{\circ}$ S. Analyses of the structure, stratigraphy, sedimentology, provenance, and



Figure 1

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 40 Ar/ 39 Ar geochronology of five Cenozoic outcrop belts support a history of roughly 40–25 Ma shortening and foredeep development followed thereafter by intermontane sedimentation and severely reduced shortening. This tectonic record in which major upper crustal shortening of Neogene age is notably absent in the Eastern Cordillera requires revision of existing models for the kinematic history of shortening and surface uplift in the central Andes.

2. Geologic Framework

[5] Crustal shortening affected distinct morphotectonic zones in Bolivia and northern Chile: from west to east, the modern forearc region (including the Precordillera), the Western Cordillera magmatic arc, the Altiplano hinterland plateau, the retroarc fold-thrust belt consisting of the Eastern Cordillera, Interandean Zone, and Subandean Zone, and the margins of the active foreland basin of the Chaco and Beni plains (Figure 1). Cretaceous to Paleogene shortening and strike-slip deformation has been reported for the forearc region and Western Cordillera along a belt of poorly understood structures [Chong, 1977; Scheuber and Reutter, 1992; Kuhn and Reuther, 1999; Horton et al., 2001, 2002a; McQuarrie et al., 2005]. In the Eastern Cordillera, shortening was accommodated by complex east and west directed thrusts that probably involve crystalline basement, although no Precambrian rocks are exposed [Roeder, 1988; Sempere et al., 1990; Baby et al., 1992; Hérail et al., 1996; Tawackoli et al., 1996; Kley et al., 1997; Lamb and Hoke, 1997; Horton, 1998; McQuarrie and DeCelles, 2001; Müller et al., 2002]. In contrast, shortening in the Interandean and Subandean zones was concentrated along a series of relatively simple, east directed structures above a primary décollement in the lower Paleozoic section [Dunn et al., 1995; Roeder and Chamberlain, 1995; Moretti et al., 1996; Kley, 1996; Baby et al., 1997; Leturmy et al., 2000]. Pre-Andean deformation certainly affected parts of the Eastern Cordillera [Martinez, 1980; Baby et al., 1992; Allmendinger et al., 1997], but an extensive nonconformity to low-angle unconformity defining the Paleozoic-Cretaceous contact [Kley et al., 1997; McQuarrie and DeCelles, 2001] indicates that most shortening in Bolivia occurred during Andean deformation of Cretaceous and Cenozoic age.

[6] At $17-21^{\circ}$ S, nonmarine Cenozoic strata in the Eastern Cordillera are distributed among elongate outcrop belts parallel to the regional tectonic strike of the fold-thrust belt. These north to northwest trending outcrop belts fall into two categories. First, in the ~4-km-high central Eastern Cordillera, basin fill of principally Neogene age is exposed in isolated accumulations of small areal extent (<300 km²). These deposits of the Parotani, Bolívar, and Mondragón regions are unconformably overlapped in places by upper

Miocene ignimbrites of the Los Frailes and Morococala volcanic fields (Figure 1). Second, in the \sim 3-km-high eastern part of the Eastern Cordillera, successions of probable middle Eocene-Oligocene age are exposed along an arcuate, 600-km-long, discontinuous belt of largemagnitude synclines: from north to south, the Morochata, Torotoro, Incapampa, Otavi, and Camargo synclines (Figure 1). Previous investigations have addressed the stratigraphy of both the central zone and eastern syncline belt [Ahlfeld and Branisa, 1960; Lohmann and Branisa, 1962; Ahlfeld, 1965; Ponce de León, 1966; Evernden et al., 1977; Martinez, 1980; Sempere et al., 1989, 1997; Martinez et al., 1990; Marshall and Sempere, 1991; Hérail et al., 1996; Blanco, 1994; Pacheco and Fernandez, 1994; Kennan et al., 1995; Tawackoli et al., 1996; Horton, 1998; Horton and DeCelles, 2001; Müller et al., 2002; DeCelles and Horton, 2003]. However, it is unclear whether these outcrop belts developed as disconnected intermontane basins, integrated foredeep depocenters, or some other type of sedimentary basin. Furthermore, the precise ages of sediment accumulation and linkages with Cenozoic deformation remain uncertain.

3. Structural Geology

[7] Geologic mapping of Cenozoic basin fill, interpretation of satellite images, and construction of simplified local cross sections reveal structural disparities between the central Eastern Cordillera and eastern syncline belt. In the central zone, strongly deformed Ordovician to lower Paleogene rocks of the Parotani, Bolívar, and Mondragón regions are unconformably overlain by subhorizontal to gently dipping successions of principally Neogene age that are capped by undeformed upper Miocene ignimbrites (Figures 2 and 3). No basin-bounding structures or growth strata are observed. In the eastern syncline belt, the Incapampa and Torotoro synclines contain continuous Cenozoic successions deformed within large-wavelength, largeamplitude folds related to flanking thrust structures. Across a large expanse of the Eastern Cordillera, one of the youngest markers in the deformation history is the undeformed, subhorizontal San Juan del Oro erosion surface, which bevels Ordovician through Cenozoic rocks and is covered by thin upper Miocene-Pliocene deposits (Figure 1). In terms of older markers, a nonconformity to low-angle unconformity defines the basal Cretaceous contact over most of the Eastern Cordillera. The Cretaceous section, however, overlies a range of Ordovician through Devonian rocks, suggesting significant pre-Andean erosion of the upper Paleozoic-lower Mesozoic section that is preserved elsewhere in the central Andes.

[8] No new estimates of slip magnitude are made for the numerous thrust faults in the region. However, previous

Figure 1. Regional tectonic map of the Eastern Cordillera, Bolivia [after *Pareja et al.*, 1978]. Boxes denote study areas (Figure 2). Cenozoic basin fill includes principally Neogene strata of the Parotani, Bolívar, and Mondragón regions and Paleogene strata of the Morochata, Torotoro, Incapampa, Otavi, and Camargo synclines. For clarity, numerous fold-thrust structures are not shown. Inset map (top right) shows major tectonic provinces of the central Andes and late Miocene volcanic fields of the Eastern Cordillera (LF, Los Frailes; MR, Moroccala).



Figure 2. Geologic maps (at same scale) and simplified cross sections (at magnified scale): (a) Parotani region, (b) Bolívar region, (c) Mondragón region, (d) Incapampa syncline, and (e) Torotoro syncline. Locations of satellite images (Figure 3), measured sections (Figure 4) and ⁴⁰Ar³⁹Ar samples (Figure 6) are shown. From field mapping and *Geobol* [1962, 1966a, 1966b, 1966c, 1993a, 1993b, 1994].



Figure 3. Landsat images: (a) Mondragón region, (b) Incapampa syncline, and (c) Torotoro syncline. Dashed lines define formation boundaries (O, Ordovician; O-D, Ordovician–Devonian; D, Devonian; K, Cretaceous; C, Cenozoic sedimentary rocks; Cv, Cenozoic volcanic rocks). See Figure 2 for locations (map area of Figure 2e approximately corresponds to Figure 3c). See color version of this figure at back of this issue.

structural studies provide estimates on the basis of observed stratigraphic separation, assumed stratal thicknesses, and construction of admissible, viable cross sections utilizing mapped cutoff relationships [Servicio Geológico de Bolivia (Geobol), 1962, 1966a, 1966b, 1966c, 1993a, 1993b, 1994, 1996; Pareja et al., 1978; Martinez, 1980; Sheffels, 1990; McQuarrie and DeCelles, 2001; McQuarrie, 2002a]. These estimates, however, are subject to inherent uncertainties related to assumptions of plane strain, fold-growth mechanisms, stratigraphic thickness, and original stratal dip prior to Andean deformation [e.g., Kley et al., 1997; Horton, 2000].

3.1. Parotani Region

[9] At 17.5°S, the northwest trending outcrop belt of the Parotani Formation overlies folded and faulted Ordovician, Devonian, and Cretaceous strata in angular unconformity, exhibits $<30^{\circ}$ dips, and is exposed over ~ 110 km² (Figure 2a). Underlying rocks are involved in southwest directed fold-thrust structures. Fault cutoff relationships indicate hanging wall flats preferentially located in the Ordovician section. Faults exhibit up to ~ 8 km of stratigraphic separation, assuming reported stratal thicknesses [Martinez, 1980; Geobol, 1994; McQuarrie and DeCelles, 2001]. These hinterland-directed structures compose part of the central Andean back thrust belt, which forms the western half of the Eastern Cordillera in Bolivia. Individual thrusts within the back thrust belt accommodated 5-40 km of reverse slip, generated many large folds, and tilted pre-Parotani bedding, faults, and fabrics to steep dips [Martinez, 1980; McQuarrie and DeCelles, 2001; McQuarrie, 2002a]. In comparison, Parotani strata are involved only in localized small folds and are not cut by major faults. A simplified cross section (Figure 2) suggests that Parotani basin fill sustained <0.5 km of shortening.

[10] The nature of the eastern margin of Parotani strata is a matter of debate, mainly due to partial Quaternary cover along a modern river. Whereas Kennan et al. [1995] report a minor west dipping normal fault, mapping by Geobol [1994] shows an unconformity with localized zones of <1.5 km thrust offset. Mapping in this study suggests an unconformity, with gently dipping Cenozoic strata overlapping deformed Devonian and Ordovician rocks in both the hanging walls and footwalls of older thrust faults, notably in the southeastern corner of the map area (Figure 2a). Regardless of these discrepancies, the original dimensions of the Parotani basin may have exceeded the present outcrop distribution. Collectively, the Parotani map relationships indicate a deformation history in which substantial shortening was followed by basin development and a subsequent period of limited shortening and possible minor normal faulting.

3.2. Bolívar Region

[11] At 18°S, the Bolívar Formation composes a northwest trending outcrop belt overlying deformed Ordovician-Devonian and Cretaceous rocks in angular unconformity, exhibits generally $<30^{\circ}$ dips, and is preserved over $\sim 210 \text{ km}^2$ (Figure 2b). The Cretaceous section unconformably overlies Orodovician, Silurian, and Devonian strata, suggesting differential erosion possibly linked to pre-Andean deformation. Pre-Bolívar strata are involved in southwest directed fold-thrust structures. Cutoff relationships indicate up to ~ 8 km of stratigraphic separation along individual faults and suggest tens of kilometers of reverse slip above a major décollement in the Ordovician section, similar to other parts of the central Andean back thrust belt [Sheffels, 1990; Geobol, 1993a, 1993b; McQuarrie and DeCelles, 2001; McQuarrie, 2002a]. No basin margin faults have been identified. Bolívar strata define an open syncline with limbs dipping 20-30°. In places, small-wavelength

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Figure 4

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folds affect the succession, but overall map relationships indicate <0.5 km of postdepositional shortening (Figure 2). Bolívar strata are overlain in angular unconformity by sub-horizontal, undeformed upper Miocene ignimbrites of the \sim 1400 km² Morococala volcanic field. In summary, the Bolívar deformation history involved large-scale predepositional shortening of Paleozoic–Mesozoic rocks, limited postdepositional folding, and late Miocene ignimbrite emplacement.

3.3. Mondragón Region

[12] The north to northwest trending outcrop belt of the Mondragón Formation at 19.5°S overlies folded and faulted Ordovician and Cretaceous rocks in angular unconformity, exhibits $<20^{\circ}$ dips, and is exposed over ~ 280 km² (Figure 2c). About 20 km west of the map area (and 8-20 km south of Challa Mayu; Figure 1), Mondragón strata possibly overlie deformed Eocene-Oligocene(?) strata of the Suticollo Formation in angular unconformity. Pre-Mondragón strata are folded and cut by east to northeast directed thrust faults. Most faults carry Ordovician strata in their hanging walls and exhibit up to ~ 8 km of stratigraphic separation. In this part of the Eastern Cordillera, individual thrusts accommodated several to tens of kilometers of reverse slip [Geobol, 1996; McQuarrie, 2002a]. Closely spaced folds along the southern part of the map area (Figure 2c) may be attributed to mutual folding of the Ordovician and Cretaceous sections or to buckle folding above a local Cretaceous décollement [Geobol, 1966a, 1966b, 1966c]. Map relationships and satellite imagery indicate no basin margin faults or intrabasinal folds affecting the Mondragón succession, which forms a nearly flatlying overlap assemblage displaying <0.1 km of shortening (Figures 2 and 3a). Mondragón strata are overlain by subhorizontal, undeformed upper Miocene ignimbrites of the \sim 8500 km² Los Frailes volcanic field. These observations indicate significant predepositional shortening, and little or no deformation associated with basin development and subsequent late Miocene ignimbrite emplacement.

3.4. Incapampa Syncline

[13] At 19.5°S, the Incapampa Formation concordantly overlies lower Paleogene strata, is folded into a large-magnitude, north trending, doubly plunging syncline, and

is exposed over an area of $\sim 60 \text{ km}^2$ (Figures 2d and 3b). The basal contact shows no angularity and Incapampa and older strata are folded to the same extent in the roughly 20 \times 40 km syncline. The western and eastern limbs dip 25-40° and $20-30^{\circ}$, respectively. The syncline is situated between a large-wavelength (~50 km wide), north trending structural culmination (the northern continuation of the Sama-Yunchará anticlinorium [Kley, 1996; McQuarrie, 2002a]) and the Eastern Cordillera-Interandean Zone boundary (Figure 1). To the west, east directed thrusts near the crest of the culmination carry Ordovician rocks in their hanging walls, exhibit up to ~ 6 km of stratigraphic separation, and display cutoff relationships consistent with a lower Ordovician décollement. These thrust faults accommodated several to tens of kilometers of reverse slip [Geobol, 1996; McQuarrie, 2002a]. The thrust shown in the southwestern part of the map area (Figure 2d) consistently exhibits a hanging wall flat in the Ordovician section, but its footwall is characterized by two flats (in Ordovician and Devonian rocks) linked by a ramp through Silurian rocks. This relationship is interpreted as transport of Ordovician hanging wall rocks over an oblique footwall ramp. East of the Incapampa syncline, the Interandean Zone and westernmost Subandean Zone consist of thin-skinned fold-thrust structures involving mostly Silurian and younger rocks.

[14] Growth of the Incapampa syncline postdated deposition of the Incapampa Formation, preceded the ~ 10 Ma San Juan del Oro surface, and was apparently linked to significant shortening accommodated by the aforementioned east directed thrusts. Syncline development may have been related to tilting of both limbs during synchronous thrusting along the western and eastern margins. Alternatively, the western limb may have formed during initial culmination uplift and the eastern limb during subsequent translation over a west dipping ramp in the Interandean-Subandean décollement, as proposed for the Bolivia-Argentina border [*Kley*, 1996; *Allmendinger and Zapata*, 2000].

3.5. Torotoro Syncline

[15] Structural relationships at 18° S in the Torotoro syncline (Figures 2e and 3c), also referred to as the Caine syncline [*Geobol*, 1966a, 1966b] are analogous to the

Figure 4. Measured sections (Figure 4a) and facies photographs (Figures 4b–4g). (a) Sections depicting basic lithologies, paleocurrent data, and stratigraphic levels for conglomerate clast counts, sandstone point counts, and 40 Ar/ 39 Ar samples. See Figure 2 for locations. (b, c) Incapampa Formation, steep canyon walls (1000 and 150 m vertical relief, respectively) composed of gently to moderately dipping, thick-bedded fluvial channel sandstone and subordinate conglomerate (facies S1 and S2). Arrow denotes approximate base of formation. (d) Incapampa Formation, moderately dipping, interbedded sandstone and mudstone of overbank fluvial origin (facies S3 and S4); 1.6 m jacob staff for scale. (e) Mondragón Formation, ~100 m thick succession of subhorizontal basin fill (C) displaying basal angular unconformity (lower dashed line) with steeply dipping Ordovician strata (O, foreground) and onlap unconformity (upper dashed line) with light colored Cretaceous strata (K, background). (f) Mondragón Formation, subhorizontal, interbedded mudstone (facies C1), minor sandstone (facies C2), and volcanic tuffs (arrows, facies C5); person for scale. (g) Bolívar Formation, gently dipping, interbedded mudstone (facies C1), minor gypsum (facies C4), and volcanic tuffs (arrows, facies C5) overlain by subhorizontal Morococala ignimbrite (Cv, background); person for scale. See enlarged version of this figure in the HTML. See color version of this figure at back of this issue.

syncline.

Incapampa syncline. The stratigraphic succession is nearly identical to the Incapampa section, with an $\sim 80 \text{ km}^2$ axial exposure of basin fill (Soyco or Soykho Formation) concordantly overlying a Cretaceous-lower Paleogene succession. The structure consists of a large-magnitude, northwest trending, roughly 20×40 km syncline with Ordovician-Devonian, Cretaceous, and Cenozoic strata folded uniformly on both limbs. Although Cenozoic strata are folded symmetrically and display moderate dips on both limbs, the Cretaceous section defines a steeper northeastern limb and shallower southwestern limb. Few recent maps are available [Geobol, 1966a, 1966b], but northwest trending folds and northeast directed thrust faults appear to be associated with syncline development. The evolution of the fold remains uncertain, but geometric similarities may imply a kinematic history comparable to the Incapampa

4. Sedimentology and Stratigraphy

[16] Stratigraphic sections were measured along the most continuous exposures (Figure 2) and are presented at a simplified scale (Figure 4a). Deposits in the central Eastern Cordillera (Mondragón, Bolívar, and Parotani regions) are grouped into five lithofacies (facies C1–C5); deposits in the eastern syncline belt (Incapampa and Torotoro synclines) are grouped into six lithofacies (facies S1–S6). Complete facies descriptions and interpreted depositional processes are presented in Tables A1 and A2 (available as auxiliary material¹) and representative examples are shown in Figures 4b–4g.

4.1. Mondragón, Bolívar, and Parotani Regions

[17] Cenozoic outcrop belts in the central Eastern Cordillera contain a variety of fine- and coarse-grained strata with interbedded volcanic rocks. Strata unconformably overlie a wide range of deformed Ordovician to lower Paleogene rocks. Although preserved thicknesses vary within each belt, the maximum values are about 300, 400, and 600 m for the Parotani, Mondragón, and Bolívar regions, respectively (Figure 4). Each belt exhibits significant lateral variability in lithofacies over <1 km distances. No large-scale stratigraphic trends, such as systematic coarsening or fining upward packages several hundred meters thick, are observed.

[18] Five lithofacies (facies C1-C5; Table A1; Figures 4e-4g) are interpreted to represent a range of depositional processes. The volumetrically dominant facies is very thin-bedded to massive siltstone (facies C1); common are thin-bedded sandstone (facies C2) and conglomerate (facies C3); thin beds of laminated gypsum (facies C4) and tuff (facies C5) are rare. The dominant depositional process was fallout of silt to fine-grained sand from suspension in lacustrine, overbank fluvial, or distal alluvial fan environments. Additional processes included deposition of sand and gravel by hyperconcentrated flows on alluvial fans and turbidity flows in lakes. Deposition of ashfall tuffs and evaporative precipitation of gypsum were relatively minor processes.

[19] Comparable depositional processes have been documented in narrow intermontane basins within many foldthrust belts [e.g., Ori and Friend, 1984; Lawton and Trexler, 1991]. Similar elongate basins paralleling fold-thrust structures have been described in the Eastern Cordillera and Puna plateau of southern Bolivia and northern Argentina [e.g., Schwab, 1985; Vandervoort et al., 1995; Hérail et al., 1996; Horton, 1998; Kraemer et al., 1999; Coutand et al., 2001]. In this study, the rapid facies changes and absence of organized, laterally continuous stratigraphic packages favor a history of locally fluctuating depositional patterns in lacustrine, alluvial fan, and minor fluvial environments. By analogy with other intermontane settings, sediment was probably derived from a collection of small drainage networks. In such systems, temporal and spatial shifts in depositional conditions may be products of local tectonic and climatic controls rather than regional forcing mechanisms [e.g., Talling et al., 1995; Horton et al., 2002b; Sobel et al., 2003].

4.2. Incapampa and Torotoro Synclines

[20] Preserved Cenozoic successions in the Incapampa and Torotoro synclines contain 1200-1500 m of uniform, sandstone-dominated, upward coarsening basin fill. Absences of volcanic rocks and diagnostic fossils prevent accurate age assignments, but the successions must postdate the underlying ~2-km-thick Cretaceous-Paleogene section of nonmarine terrigenous clastic rocks and minor marine carbonates of the La Puerta, El Molino, Santa Lucía, and Cayara formations [Geobol, 1996; Sempere et al., 1997]. Maastrichtian to mid-Paleocene ages (\sim 73 to \sim 58 Ma) for the El Molino and Santa Lucía formations are defined on the basis of magnetostratigraphy and 40 Ar/ 39 Ar ages of two interbedded tuffs [Sempere et al., 1997]. As observed in other parts of the Eastern Cordillera, a 10- to 30-m-thick zone of well-developed paleosols marks the Santa Lucía-Cayara transition (Figure 5). Although poorly dated, a similar paleosol zone at an equivalent stratigraphic level in the Camargo syncline is considered to be roughly late Paleocene to middle Eocene in age [DeCelles and Horton, 2003]. By analogy, the overlying Incapampa and Torotoro successions are considered to be of principally middle Eocene-Oligocene age. Mappable stratigraphic intervals within both synclines exhibit uniform thicknesses over >5 km distances (Figures 3b and 3c). Large-scale, upward coarsening stratigraphic trends (Figure 4b) are well expressed along the ~ 40 km outcrop lengths of both synclines.

[21] Six lithofacies (facies S1–S6, Table A2 and Figures 4b–4d) are attributed to different depositional processes within the Incapampa and Torotoro synclines. Thick-bedded channel sandstone and conglomerate (facies S1 and S2) form the principal facies and are most

¹Auxiliary material is available at ftp://ftp.agu.org/apend/tc/ 2003TC001619.



Figure 5. Photograph of well-developed paleosol zone defining transition from white, coarse-grained sandstone of Paleocene Santa Lucía Formation (SL) to concordantly overlying siltstone and fine-grained sandstone of Cayara Formation (C), Incapampa syncline. Strata dip 30° to right (east). Pervasive subvertical rhizocretions (10-20 cm wide by 30-300 cm long) and calcareous nodules (1-10 cm diameter) define zone of intense pedogenic alteration. See color version of this figure at back of this issue.

pronounced in the upper half of both successions. Thinbedded sandstone (facies S3), laminated siltstone (facies S4), and pedogenic siltstone (facies S5) are common; laminated gypsum beds (facies S6) are rare. Deposition is interpreted to represent a range of processes operative in large fluvial systems. Within channels, transport and accumulation of sand and gravel occurred in fluvial bars and bed forms. In overbank areas, silt and sand was deposited by crevasse splays and suspension fallout during water flows. During periods of nondeposition (between flooding events), overbank areas underwent pedogenic alteration and, at times, evaporative precipitation of gypsum in restricted lakes.

[22] Incapampa and Torotoro sedimentation is interpreted to be analogous to deposition in fluvial megafans [e.g., *DeCelles and Cavazza*, 1999; *Raynolds*, 2002]. The lithofacies and inferred depositional processes are strikingly similar to possible age-equivalent megafan deposits in the Camargo syncline and to active megafans in the modern Chaco Plain foredeep, which receive sediment from large drainage networks (typically >10,000 km²) [*Horton and DeCelles*, 2001]. Well-developed upward coarsening successions imply a systematic stratigraphic evolution governed by regional factors, possibly linked to longterm eastward propagation of the fold-thrust belt and expansion of drainage networks [e.g., *Friend et al.*, 1999].

[23] Stratigraphic and facies relationships equivalent to the Incapampa and Torotoro synclines define the entire syncline belt of the Eastern Cordillera (Figure 1). Each syncline in the belt contains a <2500-m-thick, upward coarsening succession of Paleogene fluvial strata that concordantly overlies lower Paleogene strata and exhibits laterally continuous exposures that form steep cliffs in upper stratigraphic levels (Figures 4b and 4c).

5. Geochronology

5.1. Methods

[24] To resolve depositional ages, ⁴⁰Ar/³⁹Ar step heating and total fusion analyses of biotite and K-feldspar from andesitic to rhyodacitic volcanic rocks were performed at the University of California, Los Angeles. An overview of analytical procedures is given by *Quidelleur et al.* [1997] and McDougall and Harrison [1999]. High-purity mineral separates were obtained using conventional density and magnetic separation techniques followed by hand selection under a binocular microscope. Selected crystals showed minimal effects of weathering, alteration, or recrystallization. Samples were irradiated for 15 hours in the 5C position of the research nuclear reactor at McMaster University, Ontario, Canada using Fish Canyon sanidine (27.8 Ma [Renne et al., 1994]) as a flux monitor to calculate J factors and using K₂SO₄ and CaF₂ salts to calculate correction factors for interfering neutron reactions. Incremental step heating experiments, consisting of 10 release steps per sample, were conducted in a Ta crucible within a double-vacuum resistance furnace. Total fusion analyses were conducted using a continuous 5-W Ar ion laser; \sim 30 mg aliquots of \sim 0.5 mm crystals were fused per analysis. The 40Ar/39Ar isotopic measurements were made using a VG 1200S automated mass spectrometer. Ages were calculated using conventional decay constants and isotopic abundances.

[25] Most ⁴⁰Ar/³⁹Ar step heating analyses (Figure 6) exhibit straightforward degassing behavior in which age spectra display plateaus and the weighted mean, total gas, and isochron ages of individual samples are indistinguishable within error. However, several samples (e.g., BO138 and VBO240) show evidence of excess argon in the initial lowest-temperature steps, possibly the result of trace amounts of inherited older grains or nonatmospheric trapped Ar components [e.g., Heizler and Harrison, 1988; Kelley, 2002]. Because the initial steps show larger errors, the weighted mean ages are considered representative of depositional ages. Unless otherwise noted, reported ages are weighted mean ages calculated for the complete (10-step) release spectra for each sample. To facilitate comparison with previously published isotopic ages, reported errors represent total age uncertainties, which include analytical errors and uncertainties in J factors and irradiation parameters. However, for the purpose of calculating sediment accumulation rates, only the analytical errors are considered, as appropriate for internal comparison among samples from a single irradiation and ⁴⁰Ar/³⁹Ar analytical program. All errors are reported at the 1σ level. Detailed results are reported in Tables A3 and A4 (see auxiliary material).

5.2. Results

[26] Twelve ⁴⁰Ar/³⁹Ar ages were obtained on volcanic rocks from the central Eastern Cordillera (Figure 6). Eight



Figure 6. The ${}^{40}\text{Ar}{}^{39}\text{Ar}$ age spectra for biotite from volcanic rocks of the Mondragón, Bolívar, and Parotani regions. Reported ages are weighted mean ages; total age uncertainties are reported at the 1σ level.

analyses of interbedded tuffs produced latest Oligocene to early Miocene ages; four analyses of ignimbrites yielded late Miocene ages.

[27] In the Mondragón region, three volcanic units have been dated by five ${}^{40}\text{Ar}{}^{39}\text{Ar}$ analyses. Samples VMON3 and AD1 from a 1- to 5-m-thick tuff in the lower Mondragón succession (75 m level; Figure 4) produced biotite step heat ages of 21.45 \pm 0.36 Ma and 21.45 \pm 0.37 Ma, respectively. Sample VMON1 from a ~30-cm-thick tuff in the upper Mondragón succession (150 m level; Figure 4) yielded a biotite step heat age of 20.51 \pm 0.33 Ma and a Kfeldspar total fusion age of 20.69 \pm 0.63 Ma. Previous studies on adjacent tuffs and possible lateral equivalents such as the Agua Dulce and Canteria formations report K-Ar ages of 20.3-18.8 Ma for three volcanic units [*Evernden* et al., 1977], 21.9 ± 0.36 Ma to 20.9 ± 0.32 Ma for six tuffs [*Grant et al.*, 1979], and 19.7 ± 0.8 Ma and 19.0 ± 0.7 Ma for a single tuff [*Kennan et al.*, 1995].

[28] Unconformably overlying the Mondragón strata are subhorizontal ignimbrites erupted from the Condor Nasa eruptive center of the Los Frailes volcanic field. A biotite step heat age of 7.34 ± 0.15 Ma has been obtained on sample LF1 from the easternmost part of the volcanic field (Figure 2c). Similarly, a biotite step heat age of 7.58 ± 0.15 Ma was acquired on sample CV320 from the Ayo Ayo locality (Figure 1), where a Los Frailes ignimbrite unit unconformably overlies deformed strata of the Eocene–Oligocene(?) Suticollo Formation. These two ages coincide with the 8–6 Ma main phase of magmatic activity in the Los Frailes region [*Evenden et al.*, 1977; *Grant et al.*, 1975; *Schneider*, 1985; *Schneider and Halls*, 1985; *Kennan et al.*, 1995] and provide a younger age bound for the cessation of shortening in the central Eastern Cordillera.

[29] In the Bolívar region, three volcanic units have been dated by four ^{40}Ar , ^{39}Ar analyses. Samples BO138 and VBO240 from tuffs in the Bolívar succession (138 and 570 m levels; Figure 4) yielded biotite step heat ages of 24.84 ± 0.92 Ma and 17.75 ± 0.48 Ma, respectively (Figure 6). An extensive ash flow tuff erupted from the Tahnka Tahnka caldera of the Morococala volcanic field [Ahlfeld and Branisa, 1960; Soria, 1988; Luedke et al., 1990; Ericksen et al., 1990; Morgan et al., 1998] unconformably overlies the western margin of the Bolívar succession. Sample MRC1 from this subhorizontal ignimbrite produced a biotite step heat age of 6.64 ± 0.14 Ma and a Kfeldspar total fusion age of 6.46 ± 0.11 Ma. Previously reported K-Ar and ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages for the Morococala volcanic field range from 8.4 to 6.0 Ma, contemporaneous with Los Frailes volcanism [Grant et al., 1979; Lavenu et al., 1989; Koeppen et al., 1987; Luedke et al., 1990; Ericksen et al., 1990; Morgan et al., 1998].

[30] Two Parotani tuffs have been dated by ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ analyses. Samples PARO1 and PARO3 (35 and 85 m levels; Figure 4) yielded biotite step heat ages of 20.71 ± 1.40 Ma and 20.18 ± 0.78 Ma, respectively (Figure 6). Although separated by 50 m of stratigraphic section, the weighted mean ages are indistinguishable within error. Previously reported K-Ar biotite ages for a single tuff in the lower Parotani Formation range from 20.5 ± 0.7 Ma to 18.6 ± 1.3 Ma [Kennan et al., 1995].

5.3. Accumulation Rates

[31] The 40 Ar/ 39 Ar ages facilitate calculation of sediment accumulation rates, utilizing the measured stratigraphic thicknesses between successive tuffs (Figure 4a) and the analytical age uncertainties. Thickness values are not corrected for the effects of compaction due to limited postdepositional burial. The Mondragón section exhibits an accumulation rate of 80 m/Myr (65–103 m/Myr with analytical error) on the basis of 75 m deposited over 0.94 ± 0.21 Myr. The Bolívar section displays a rate of 61 m/Myr (52–74 m/Myr with analytical error) on the basis of 432 m



Figure 7. Summary of paleocurrent data. Rose diagrams for (a) Bolívar and Mondragón regions and (b) Incapampa and Torotoro synclines. Data are grouped into 10° bins. Large arrows and short arc segments depict vector means and standard deviations, respectively. (c) Map view summary (compare to Figure 2d) of paleocurrent data for Incapamapa syncline. Arrows and numerals represent vector means and total measurements for each station, respectively.

deposited over 7.09 ± 1.24 Myr. For the Parotani section, the weighted mean ages are indistinguishable within error, but the total gas ages suggest a rate of 46 m/Myr (30–93 m/Myr with analytical error) on the basis of 50 m deposited over 1.09 ± 0.55 Myr.

[32] These successions in the central Eastern Cordillera display average accumulation rates of 46–80 m/Myr over 1-7 Myr, significantly lower than reported rates for Andean foredeeps. Foredeep accumulation rates range from 100 to 1500 m/Myr over <1 Myr periods and average about 300–600 m/Myr over 1-12 Myr periods, on the basis of magnetostratigraphic studies of the Subandean and Santa Bárbara zones at $22-25^{\circ}$ S [*Reynolds et al.*, 2000, 2001; *Echavarria et al.*, 2003], the Bermejo basin at $29-31^{\circ}$ S [*Johnson et al.*, 1986; *Reynolds et al.*, 1990; *Beer*, 1990; *Jordan*, 1995; *Jordan et al.*, 2001], and the Manantiales basin at 32° S [*Jordan et al.*, 1996]. The sharp contrast in accumulation rates may be related to disparities in the rates of production, transport, and/or accommodation of sediment.

6. Sediment Provenance

[33] To assess patterns of exhumation and sediment dispersal, a study was carried out on paleocurrent histories and clastic sediment composition. Measured paleocurrent indicators include trough axes of cross-stratified sandstone and conglomerate, imbricated pebble-cobble clasts, and limited primary current lineations in sandstone. Conglomerate clast counts (identification of 100 clasts per station) were performed at selected outcrops along measured sections. Sandstone petrographic analyses involved point counting (450 framework grains per thin section) of samples collected from measured sections. Detailed methods and results are reported in Appendix A and Tables A5 and A6 (see auxiliary material).

6.1. Paleocurrents

[34] Paleocurrent data reveal conflicting sediment dispersal for the central zone and eastern syncline belt of the Eastern Cordillera. A total of 800 paleocurrent indicators were measured at 60 stations. Arrows plotted alongside measured sections (Figure 4a) represent mean paleocurrent vectors for individual stations. Grouping of station data and calculation of pooled vector means and standard deviations (reported at the 1σ level) show the distribution of paleocurrent directions (Figure 7). For the Bolívar and Mondragón sites, a vector mean of $131 \pm 9^{\circ}$ is calculated on the basis of 238 measurements at 18 stations. This southeastward transport approximately parallels predepositional structures within the fold-thrust belt, consistent with axial paleoflow. The Incapampa and Torotoro sites, however, yield a vector mean of $81 \pm 6^{\circ}$ on the basis of 562 measurements at 42 stations. This east-northeastward transport lies at a high angle to the overall structural grain of the fold-thrust belt, indicating transverse paleoflow. Data from stations spanning the Incapampa outcrop belt (Figure 7c) suggest a radiating dispersal pattern: northern stations show eastnortheastward vectors and southern stations show eastsoutheastward vectors. This configuration, expressed over \sim 15 km along strike, corresponds to the radial flow pattern and spatial scale of sediment dispersal in fluvial megafans [e.g., Sinha and Friend, 1994; Friend et al., 1999; DeCelles and Cavazza, 1999; Horton and DeCelles, 2001].

6.2. Conglomerate Composition

[35] Conglomerate clast counts from 34 stations within measured sections indicate a compositional distinction between the central zone and eastern syncline belt. Conglomerates of the Incapampa syncline contain dominantly Cretaceous clasts, a moderate amount of Silurian clasts, limited Ordovician clasts, and no Cenozoic volcanic clasts (Figure 8). A uniformly high proportion of Mesozoic clasts suggests erosion of shallow structural levels during an early stage of unroofing. A paucity of Devonian clasts is consistent with a sediment source >10 km west of the syncline, where Devonian strata are missing [Geobol, 1996]. Conversely, Bolívar and Mondragón conglomerates display abundant Ordovician clasts and generally limited proportions of Silurian, Devonian, Cretaceous, and Cenozoic clasts (Figure 8). The high proportion of Ordovician clasts suggests erosion of deeper levels during a later stage of unroofing.

[36] A discrepancy is also revealed by compositional fluctuations within each succession. Whereas Incapampa deposits exhibit limited stratigraphic variation in the ratio of Mesozoic–Cenozoic to total Mesozoic–Cenozoic and Ordovician clasts, the Bolívar and Mondragón sections display high variability (Figure 8b). Heterogeneous clast compositions in the Bolívar and Mondragón successions suggest



Figure 8. Summary of conglomerate clast compositional data for Bolívar, Mondragón, and Incapampa regions. (a) Pie charts of clast count data. (b) Plot of clast compositional ratio (defined as proportion of Cenozoic and Cretaceous clasts to total Cenozoic, Cretaceous, and Ordovician clasts) relative to normalized stratigraphic level, showing temporal (upsection) variations in clast composition.

that spatial or temporal changes in bulk rock exposure within the source areas were successfully transmitted to the depositional basins. Low clast variability in the Incapampa succession suggests that either bulk rock exposure in the source area did not change significantly through time or any changes were buffered within the drainage network by clast attrition or dilution by a dominant source [e.g., *Steidtmann and Schmitt*, 1988].

6.3. Sandstone Composition

[37] Sandstone petrographic analyses indicate contrasting compositions in the central Eastern Cordillera and eastern syncline belt. Using a modified Gazzi-Dickinson method of point counting [Gazzi, 1966; Dickinson, 1970; Ingersoll et al., 1984; Dickinson, 1985], 450 framework grains were identified per thin section for 48 samples. Petrographic counting parameters and recalculated detrital modes are shown in Tables A5 and A6 (see auxiliary material). Recalculated data are presented in quartz/feldspar/lithic fragment (Q/F/L) ternary diagrams (Figure 9). Principally quartzarenitic compositions characterize the Incapampa and Torotoro synclines (Q/F/L mean compositions of 95/2/3 and 93/3/4, respectively). Sublitharenitic compositions are recorded by the Mondragón and Bolívar localities (Q/F/L mean compositions of 90/1/9 and 82/3/15, respectively). Plots of mean standard deviations (reported at the 1σ level; Figure 9b) show the compositional difference between the central zone and syncline belt. Incapampa and Torotoro compositions are nearly equivalent to the mean sandstone compositions reported by DeCelles and Horton [2003] for the Camargo and Suticollo formations at the Cerro Tonka and Challa Mayu localities, respectively (Figure 1).

[38] One possibility for the higher proportion of detrital quartz within the syncline belt is breakdown of less resistant

feldspar and lithic fragments by physical and chemical weathering [e.g., *Johnsson*, 1990]. Alternatively, the increased quartz fraction may reflect quartz-rich compositions in the sediment source area. Both explanations have merit for the Incapampa region, where Cretaceous quartzarenites were a major source of gravel (Figure 8) and a larger drainage network (suggested by facies considerations) would have guaranteed greater transport and enhanced physical weathering.

7. Basin Development

[39] The preceding synthesis reveals inconsistencies in basin development between the central zone and eastern syncline belt of the Eastern Cordillera. (1) Large-scale upper crustal shortening preceded basin evolution in the central zone yet mostly postdated basin growth in the syncline belt. (2) 25-17 Ma volcanism coincided with deposition in the central zone, in contrast to the older nonvolcanogenic record of the syncline belt. (3) Whereas lithofacies assemblages in the syncline belt suggest extensive drainage networks capable of producing fluvial megafans, small drainage networks in the central zone fed localized alluvial fan, lacustrine, and minor fluvial systems. (4) Paleocurrent data demonstrate dominantly transverse flow within the syncline belt but axial flow paralleling older fold-thrust structures in the central zone. (5) Whereas sediment compositional data from the syncline belt recorded initial erosional unroofing of mainly Cretaceous and Silurian levels, data from the central zone indicate subsequent unroofing of lower, mostly Ordovician, levels. (6) Relatively lithic-rich sandstones and greater variability in gravel compositions in the central zone imply local sediment sources and shorter transport distan-



Figure 9. (a) Q/F/L (quartz/feldspar/lithic fragment) plots of sandstone petrographic data for Incapampa, Torotoro, Mondragón, and Bolívar regions. (b) Mean compositional values (open and solid symbols) and standard deviations (polygons) of data plotted in Figure 9a. Data from *DeCelles and Horton* [2003] are plotted for comparison.

ces, as opposed to the mature quartz-rich sandstones and more uniform clast compositions of the syncline belt.

[40] These disparities signify distinctive geomorphologic conditions during basin evolution. In the central Eastern Cordillera, a valley-and-ridge topography is suggested by axial sediment dispersal, small drainage networks, and predepositional fold-thrust deformation. Accordingly, the uppermost Oligocene-lower Miocene outcrop belts of the Parotani, Bolívar, and Mondragón regions are considered isolated intermontane basins. In contrast, successions in the Incapampa and Torotoro synclines concordantly overlie an extensive Cretaceous-Paleogene section of distal fluviolacustrine and marginal marine strata (Figure 1). Moreover, Incapampa and Torotoro lithofacies (Table A2) are indistinguishable from fluvial megafan deposits of the Camargo syncline [Horton and DeCelles, 2001]. On the basis of nearly identical stratigraphic and sedimentologic relationships, it is proposed here that Cenozoic outcrop belts in the eastern syncline belt (including the Incapampa, Torotoro, Camargo, Otavi, and Morochata synclines) collectively formed a single, continuous foredeep during middle Eocene-Oligocene time. Although foredeep growth could have initiated earlier [e.g., Sempere et al., 1997], an extensive paleosol zone capping the mid-Paleocene Santa Lucía Formation (Figure 5) suggests an abrupt hiatus in sedimentation prior to megafan accumulation. The duration of this hiatus is poorly known, but a similar paleosol zone in the Altiplano represents about 15-20 Myr [e.g., Horton et al., 2001]. Therefore, in the absence of precise age data, the onset of foredeep conditions is tentatively attributed to middle Eocene time.

[41] The division of foredeep and intermontane basins outlined here is based on contrasting structural, chronostratigraphic, lithostratigraphic, and sedimentological attributes (summarized in Figure 10). The studied intermontane basins are further distinguished from more commonly recognized piggyback and wedge top basins [e.g., *Ori and Friend*, 1984; *Horton and DeCelles*, 1997] by absences of growth strata, deformed basin fill, and underlying foredeep depositis. Whereas foredeep deposition is interpreted to be the result of flexural subsidence, the intermontane basins are associated with insignificant shortening, suggesting negligible flexural subsidence. Probable sediment accommodation mechanisms in the intermontane basins include sediment ponding and sediment loading in closed drainage basins.

[42] Interpretation of foredeep and intermontane basins is consistent with the sediment accumulation record. First, whereas intermontane basins contain 300–600 m of fill, foredeep successions in the syncline belt are generally 1000–2500 m thick. The thickest preserved foredeep succession (2300 m) occurs in the Camargo syncline at 20.8°S, where the original basin configuration is partially defined by apatite fission track data. The record of partial annealing of fission tracks along a sample transect at ~21°S is consistent with a foredeep attaining a maximum thickness of ~3.5 km and an east-west width exceeding 60–100 km [*Ege et al.*, 2003; *Ege*, 2004], values comparable to the modern foredeep [*Horton and DeCelles*, 1997]. Most of the



Figure 10. Summary of stratigraphic relationships for Cenozoic basin fill of the Eastern Cordillera, including basic lithologies, ages, and formation names. Note significant contrasts between foredeep and intermontane basins. See text for description of available age control.

original basin fill has been eroded away, leaving behind small preserved relicts of the foredeep in the eastern syncline belt. Second, the onset of conglomerate deposition, where recorded, occurred at different stages of basin evolution in the foredeep and intermontane basins. Gravel sedimentation commenced upon initial accumulation in the Bolívar and Mondragón intermontane basins, but was mostly limited to the upper halves of preserved foredeep sections in the Incapampa and Torotoro synclines (Figure 4). Third, although poor age control for studied foredeep fill prevents direct comparison, the intermontane accumulation rates are <80 m/Myr, several times lower than other Andean foredeeps. Conflicting accumulation rates may reflect disparate sediment fluxes or sediment accommodation mechanisms. For example, foredeeps receiving sediment from larger drainage networks would vield relatively higher sediment loads than intermontane basins fed by small networks [e.g., Damanti, 1993]. Alternatively, whereas thrust-induced flexure causes sustained foredeep subsidence, simple sediment ponding driven by closed drainage conditions [e.g., Sobel et al., 2003; Riller and Oncken, 2003] would not continually generate significant accommodation within intermontane basins.

[43] Interpretations of basin evolution and associated deformation can be summarized as follows. Middle Eocene–Oligocene growth of a flexural foredeep was driven by shortening in the western and central Eastern Cordillera, where balanced cross sections indicate up to 60–140 km of coeval shortening, although uncertainties in these estimates are poorly defined [*Sheffels*, 1990; *Baby et al.*, 1997; *Kley and Monaldi*, 1998; *McQuarrie and DeCelles*, 2001; *McQuarrie*, 2002a; *Müller et al.*, 2002]. A lack of volcanic material suggests that shortening and foredeep generation preceded the ~25 Ma onset of volcanism across the Western

Cordillera to Eastern Cordillera region [*Coira et al.*, 1982; *Isacks*, 1988; *Sempere et al.*, 1990; *Allmendinger et al.*, 1997; *Jordan et al.*, 1997]. Growth of intermontane basins at 25–17 Ma, as defined by 40 Ar/³⁹Ar analyses (Figure 6), occurred after the major episode of upper crustal shortening. A limited amount of postdepositional shortening recorded by the intermontane basins (<5 km) preceded development of the ~10 Ma San Juan del Oro erosion surface and emplacement of the 8–6 Ma Los Frailes and Morococala ignimbrites.

8. Discussion

[44] Resolving the age of the transition from foredeep to intermontane basin development provides a refined tectonic history of the central Andes, information crucial for determining the timing and rates of shortening. Several models propose initial distributed shortening across the Eastern Cordillera from 25 to 10 Ma, followed by thinskinned deformation in the Subandean Zone and associated underthrusting of the Brazilian Shield [Isacks, 1988; Gubbels et al., 1993; Allmendinger and Gubbels, 1996]. Contrary to these models, large-scale upper crustal shortening (60-140 km) over large parts of the Eastern Cordillera was accomplished prior to about 25-21 Ma, the age of initial intermontane basin development. Following middle Eocene-Oligocene foredeep development, uppermost Oligocene-lower Miocene successions in the intermontane basins were deposited in angular unconformity over highly deformed Ordovician to lower Paleogene rocks and have themselves undergone extremely limited postdepositional shortening (<5 km). These basins provide unique markers within the regional deformation history, offering crosscutting relationships comparable to the subhorizontal, undeformed ~10 Ma San Juan del Oro erosion surface (Figure 1) that bevels deformed rocks over large regions of the Eastern Cordillera [Servant et al., 1989; Gubbels et al., 1993; Kennan et al., 1997]. Moreover, localized thin successions (<250 m) of upper Miocene-Pliocene strata that cover this erosion surface exhibit stratigraphic and sedimentologic similarities to the intermontane basin fill. Therefore it is proposed here that the estimated termination age for major upper crustal shortening over most of the Eastern Cordillera at 17-21°S be revised from the generally accepted value of ~ 10 Ma to about 25-21 Ma. However, it must be emphasized that possible shortening after 25-21 Ma cannot be ruled out for the easternmost Eastern Cordillera, where syncline growth postdated Paleogene foredeep fill and predated the late Miocene erosion surface [Allmendinger and Zapata, 2000; Müller et al., 2002], or for the Altiplano-Eastern Cordillera boundary [Lamb and Hoke, 1997; Ege et al., 2003; Ege, 2004].

[45] The timing of initial shortening in the Eastern Cordillera remains poorly constrained. A probable middle Eocene–Oligocene age of foredeep sedimentation suggests that crustal thickening, surface uplift, and flexural loading were underway by roughly 40 Ma, considerably earlier than commonly envisioned [e.g., *Isacks*, 1998; *Sempere et al.*, 1990; Gubbels et al., 1993; Allmendinger and Gubbels, 1996; Allmendinger et al., 1997; Jordan et al., 1997]. Earlier shortening is consistent not only with the sedimentation record for the Eastern Cordillera [Kennan et al., 1995; Horton and DeCelles, 2001; DeCelles and Horton, 2003; McQuarrie et al., 2005], but also with paleomagnetic results requiring Paleogene vertical axis rotations in the Eastern Cordillera [Butler et al., 1995; Lamb, 2001] and thermochronological data from the western and central Eastern Cordillera suggesting increased denudation and short-lived (<15 Myr) heating episodes linked to thrusting [Benjanin et al., 1987; McBride et al., 1987; Farrar et al., 1988; Ege et al., 2003; Ege, 2004].

[46] Despite improved age control, fundamental questions persist over the geometry and magnitude of shortening. Previous studies depict varying degrees of east and west directed, basement-involved deformation and offer conflicting estimates of total upper crustal shortening in the Eastern Cordillera [Baby et al., 1997; Schmitz, 1994; Kley, 1996; Kley et al., 1997; McOuarrie, 2002a]. Although a range of shortening geometries have been proposed [Isacks, 1988; Sheffels, 1990; Gubbels et al., 1993; Schmitz, 1994; Baby et al., 1997], a basement-involved eastward tapering tectonic wedge may account for hinterland-directed (back thrust) structures long recognized in the western part of the Eastern Cordillera throughout Bolivia [Newell, 1949; Pareja et al., 1978; Martinez, 1980]. Such a tectonic wedge geometry, defined by a lower décollement (floor thrust) with eastward transport and a shallow décollement (roof thrust) with westward transport, characterizes not only regional profiles at 13-20°S [Roeder, 1988; Sempere et al., 1990; Roeder and Chamberlain, 1995; McQuarrie and DeCelles, 2001; McQuarrie, 2002a], but also cross sections for the 20-22°S region near the Bolivia-Argentina border [e.g., Kley et al., 1997; Allmendinger and Zapata, 2000; Müller et al., 2002]. Similar back thrust systems may accommodate large amounts of shortening in many fold-thrust orogens [e.g., Price, 1986; Erslev, 1993; Yin et al., 1994; Stockmal et al., 1998].

[47] Proposed emplacement of a basement-involved tectonic wedge beneath the Eastern Cordillera may be reconciled with several observations, including the ages, locations, and stratigraphic records of foredeep to intermontane basin fill (Figure 10) and upper crustal structural geometries. In this model (Figure 11), initial shortening and eastward emplacement of the tectonic wedge coincides with foredeep development. As shortening continues, the frontal wedge tip and overlying monocline [e.g., Vann et al., 1986] advance systematically eastward, recycling the proximal foredeep [e.g., Covey, 1986]. At ~25 Ma, wedge emplacement terminates and slip initiates along a new décollement farther east in the Interandean Zone and possibly the Subandean Zone. In this model, intermontane basins develop in the elevated hinterland after ~ 25 Ma as the entire Eastern Cordillera translates eastward above the Interandean-Subandean décollement with little or no internal deformation. Translation over a west dipping ramp induces wholesale tilting of the eastern flank of the syncline belt [e.g., Kley, 1996; Allmendinger and Zapata,



Figure 11. Schematic model for evolution of foredeep and intermontane basins of the Eastern Cordillera, showing cross-sectional view of midcrustal décollements, upper crustal structures, and zones of basin development. Pin lines and open reference circles show relative fault offsets and fault tip propagation through time. (a) Middle to late Eocene, eastward translation of tectonic wedge (shaded) induces growth of Eastern Cordillera and flexural foredeep. (b) Oligocene, final emplacement of tectonic wedge causes tilting of western limb of syncline belt. (c) Early Miocene cessation of wedge emplacement and initiation of translation along structurally lower décollement feeding slip eastward into Interandean and Subandean zones. Transport over footwall ramp induces tilting of eastern limb of syncline belt.

2000]. Therefore the syncline belt is proposed to be the product of initial tilting of the western limb above the frontal tip of the tectonic wedge and subsequent tilting of the eastern limb above a ramp in the Interandean-Subandean décollement. An implication of this model is that Interandean and Subandean shortening may have started as early as 25 Ma.

[48] Potential drawbacks of the tectonic wedge model include the mechanical feasibility and expected flexural response to emplacement of a long (>100 km), thin (<15 km) basement sheet in the middle to upper crust. Under normal circumstances, such a basement sheet may not be expected to survive long-distance transport; if the sheet did remain intact, it would generate extreme flexural subsidence in the proximal foredeep. One possible explanation for both problems is exceptionally strong basement in the Eastern Cordillera during early to mid-Cenozoic time. Relatively rigid basement would not only retain its integrity over greater transport distances along a large-magnitude thrust but also, upon flexural loading, exhibit considerably reduced maximum foredeep subsidence. Further work is needed to address these issues.

[49] The synthesis of foredeep and intermontane basin evolution presented here suggests that the onset and cessation of major upper crustal shortening in the Eastern Cordillera (and by inference, the onset of Interandean and Subandean deformation) occurred roughly 15–20 Myr earlier than most studies suggest [Isacks, 1988; Sempere et al., 1990; Gubbels et al., 1993; Allmendinger and Gubbels, 1996; Moretti et al., 1996; Kley, 1996; Kley et al., 1997; Allmendinger et al., 1997; Jordan et al., 1997]. Because these age estimates play pivotal roles in the calculation of long-term rates of geologic shortening, all existing comparisons to short-term geodetic velocities in the central Andes [e.g., Norabuena et al., 1998; Horton, 1999; Liu et al., 2000; Lamb, 2000; Bevis et al., 2001; Hindle et al., 2002; Khazaradze and Klotz, 2003] must be regarded with extreme caution.

9. Conclusions

[50] 1. Two classes of Cenozoic basins are preserved in the Eastern Cordillera at 17–21°S. In the east, a formerly extensive, middle Eocene-Oligocene foredeep associated with major upper crustal shortening (60-140 km) is exposed at \sim 3 km altitude in regionally continuous synclines flanked by fold-thrust structures. Foredeep strata concordantly overlie lower Paleogene paleosols, contain at least 1200-1500 m of upward coarsening, nonvolcanogenic, fluvial megafan facies, yield quartzarenitic sandstone compositions, have mostly Cretaceous and Silurian conglomerate clasts, and show transverse sediment dispersal. In contrast, localized intermontane basins of latest Oligocene-early Miocene age in the ~4-km-high central Eastern Cordillera are characterized by limited upper crustal shortening (<5 km) and lack basin-bounding structures. Intermontane deposits unconformably overlie deformed Ordovician to lower Paleogene rocks, contain <300-600 m of variably fine- and coarse-grained alluvial fan, lacustrine, and minor fluvial facies with interbedded volcanic rocks, exhibit relatively lithic-rich sandstones, have predominantly Ordovician clasts, and display axial sediment dispersal.

[51] 2. New ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages of uppermost Oligocene– lower Miocene tuffs (25–17 Ma) indicate intermontane sediment accumulation at average rates <80 m/Myr, considerably lower than quantified rates in Andean foredeeps. The ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages of unconformably overlying, unde-

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formed upper Miocene ignimbrites require final intermontane accumulation and minor postdepositional shortening prior to 8–6 Ma. Although not directly dated, foredeep development is considered contemporaneous with middle Eocene–Oligocene deformation and denudation in the western and central Eastern Cordillera.

[52] 3. Foredeep and intermontane basin development may have been linked to emplacement of a basementinvolved tectonic wedge and subsequent translation along the Interandean-Subandean décollement farther east. Tectonic wedge emplacement along an east directed floor thrust and west directed roof thrust may have induced flexural subsidence and foredeep development in the Eastern Cordillera from roughly 40 to 25 Ma. Subsequent evolution of elevated intermontane basins from 25–21 Ma to possibly 10 Ma occurred during diminished shortening and negligible flexural subsidence, suggesting accumulation driven by closed drainage conditions that induced infilling of local topography.

[53] 4. A mid-Cenozoic transition from foredeep to intermontane sedimentation suggests the termination of

major upper crustal shortening in the Eastern Cordillera, and potential onset of shortening in the Interandean and Subandean zones, by about 25-21 Ma. This age estimate is 15-20 Myr earlier than assumed by previous conceptual models, highlighting the poor resolution of long-term shortening rates in the Andes.

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Figure 3. Landsat images: (a) Mondragón region, (b) Incapampa syncline, and (c) Torotoro syncline. Dashed lines define formation boundaries (O, Ordovician; O-D, Ordovician–Devonian; D, Devonian; K, Cretaceous; C, Cenozoic sedimentary rocks; Cv, Cenozoic volcanic rocks). See Figure 2 for locations (map area of Figure 2e approximately corresponds to Figure 3c).

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Figure 4

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Figure 5. Photograph of well-developed paleosol zone defining transition from white, coarse-grained sandstone of Paleocene Santa Lucía Formation (SL) to concordantly overlying siltstone and fine-grained sandstone of Cayara Formation (C), Incapampa syncline. Strata dip 30° to right (east). Pervasive subvertical rhizocretions (10–20 cm wide by 30-300 cm long) and calcareous nodules (1–10 cm diameter) define zone of intense pedogenic alteration.

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Figure 4. Measured sections (Figure 4a) and facies photographs (Figures 4b–4g). (a) Sections depicting basic lithologies, paleocurrent data, and stratigraphic levels for conglomerate clast counts, sandstone point counts, and $^{40}Ar/^{39}Ar$ samples. See Figure 2 for locations. (b, c) Incapampa Formation, steep canyon walls (1000 and 150 m vertical relief, respectively) composed of gently to moderately dipping, thick-bedded fluvial channel sandstone and subordinate conglomerate (facies S1 and S2). Arrow denotes approximate base of formation. (d) Incapampa Formation, moderately dipping, interbedded sandstone and mudstone of overbank fluvial origin (facies S3 and S4); 1.6 m jacob staff for scale. (e) Mondragón Formation, ~100 m thick succession of subhorizontal basin fill (C) displaying basal angular unconformity (lower dashed line) with steeply dipping Ordovician strata (O, foreground) and onlap unconformity (upper dashed line) with light colored Cretaceous strata (K, background). (f) Mondragón Formation, subhorizontal, interbedded mudstone (facies C1), minor sandstone (facies C2), and volcanic tuffs (arrows, facies C5); person for scale. (g) Bolívar Formation, gently dipping, interbedded mudstone (facies C1), minor gypsum (facies C4), and volcanic tuffs (arrows, facies C5) overlain by subhorizontal Moroccala ignimbrite (Cv, background); person for scale. See enlarged version of this figure in the HTML.