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Comparison of crustal thickening budget and shortening estimates in southern Peru (12–14°S): Implications for mass balance and rotations in the "Bolivian orocline"

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ABSTRACT

Two fundamental questions of interest with regard to the Andean Plateau are the mass balance of material needed to create and sustain a 3–4-km-high plateau and the relationship between the plateau and the bending of the Bolivian orocline. The link between these two questions is the distribution of shortening through the Central Andes. Is crustal shortening sufficient to support an isostatically compensated 60–70-km-thick crust? Is differential shortening between the hinge of the orocline and its limbs sufficient to account for both physiographic curvature and measured rotations? Three-dimensional (3-D) models of deformation within curved orogens require, in addition to paleomagnetic data, viable two-dimensional estimates of displacements. To this end, we present new map data and use those data to derive estimates of shortening across the northern margin of the Andean Plateau. The cross-section extent, from the eastern edge of the volcanic arc to the foreland basin, is approximately one half of the physiographic width of the Andean Plateau in Peru. Cross-sectional shortening estimates in southern Peru (12–14°S) provide a preferred estimate of 123 km or 40% shortening, with a minimum estimate of 58 km or 25% shortening and maximum estimate of 333 km or 65% shortening. The largest controlling factor on the magnitude of shortening is where basement becomes involved in thrust belt deformation. Using our preferred shortening estimate, the differential shortening between our cross section on the northern margin of the Bolivian orocline (12–14°S) and cross sections across the axis of the Bolivian orocline (17–18°S) is 177 km, corresponding to 23° of rotation. The differential shortening for the young (younger than 15 Ma) deformation of the Altiplano and Subandean region between 12–14°S and 17–18°S is 88 km, correlating to 11.5° of rotation. While differential shortening can account for young (younger than 15 Ma) vertical-axis rotations, it cannot completely account for early 45–35 Ma rotations (37° ± 15°). This suggests that forearc rotations reflect both bending due to differential shortening, ~12°, as well as block rotations (~25°–30°), which facilitated deformation in a tightening core and transported material toward the center of the orocline. The preferred estimate of shortening is well short of the required 240–300 km of shortening needed to account for a 60–70-km-thick crust under the entire plateau. This suggests that for an isostatically equilibrated crust, either (1) there is a significant amount of shortening (~100–150 km) in the western half of the plateau that is hidden by the volcanic arc; or (2) crustal material is being added to the Peruvian section of the Andean Plateau, most likely through lower-crustal flow.

INTRODUCTION

The Central Andes are defined by two of South America's most distinguishing morphologic features, the ~4-km-high, 400-km-wide Andean Plateau, and a pronounced seaward concave bend of the western coast and cordillera between ~13°S and 29°S latitude, termed the Bolivian orocline (Fig. 1). Plate motion, changes in subduction angle, shortening, stratigraphy, delamination, and erosion have all been suggested as possible controls on the large-scale morphology of the Central Andes. Due to the coexistence of the plateau and orocline in space, many authors have proposed that both features codeveloped in time (Isacks, 1988; Kley, 1999; Rousse et al., 2005; Roperch et al., 2006; Arriagada et al., 2008), implying connections between along-strike shortening magnitudes, shape of the orogen, and measured paleomagnetic rotations (Isacks, 1988; Kley and Monaldi, 1998; Kley, 1999). Although the magnitude of shortening has often been thought of as the primary control of crustal thickening and thus surface uplift of the Andean Plateau through isostasy (Sheffels, 1990; Schmitz, 1994; Allmendinger et al., 1997; Baby et al., 1997; Kley and Monaldi, 1998; Kley et al., 1999; McQuarrie and DeCelles, 2001; McQuarrie, 2002a; McQuarrie et al., 2005), a long-lived debate centers on whether or not there is enough shortening to account for the 60–70-km-thick crust of the Andean Plateau (e.g., Allmendinger et al., 1997; Kley and Monaldi, 1998; McQuarrie, 2002a; Hindle et al., 2005; Roperch et al., 2006). Many initial shortening estimates were too low to account for the entire crustal area and at most only accounted for 70%–80% of the area (Kley, 1996; Allmendinger et al., 1997; Baby et al., 1997; Lamb and Hoke, 1997; Kley and Monaldi, 1998; Sheffels, 1990). However, recent shortening estimates from two-dimensional (2-D) cross sections through the Bolivian Andes indicate that there is enough shortening to account for a 2-D crustal area from the volcanic arc centered in the Western Cordillera to the undeformed foreland in the east (McQuarrie and DeCelles, 2001; McQuarrie, 2002a; Müller et al., 2002; Elger et al., 2005; McQuarrie et al., 2005). A more three-dimensional (3-D) approach to the problem (e.g., Kley, 1999; Hindle et al., 2005) requires accounting for the curvature of the orogen, which predicts motion of material toward the axis of the bend. This out-of-plane addition of material from north and south can increase the plateau volume by ~9% (Hindle et al., 2005), highlighting the 3-D nature of mass balance through the Central Andes.

A 3-D kinematic analysis of any region first requires a series of 2-D sections to define lateral variations in across-strike motion (Kley, 1999; Hindle et al., 2005; McQuarrie and Wernicke, 2005; Hatcher et al., 2007; Arriagada et al., 2008). Thus, a critical piece of information to any mass balance arguments for the magnitude

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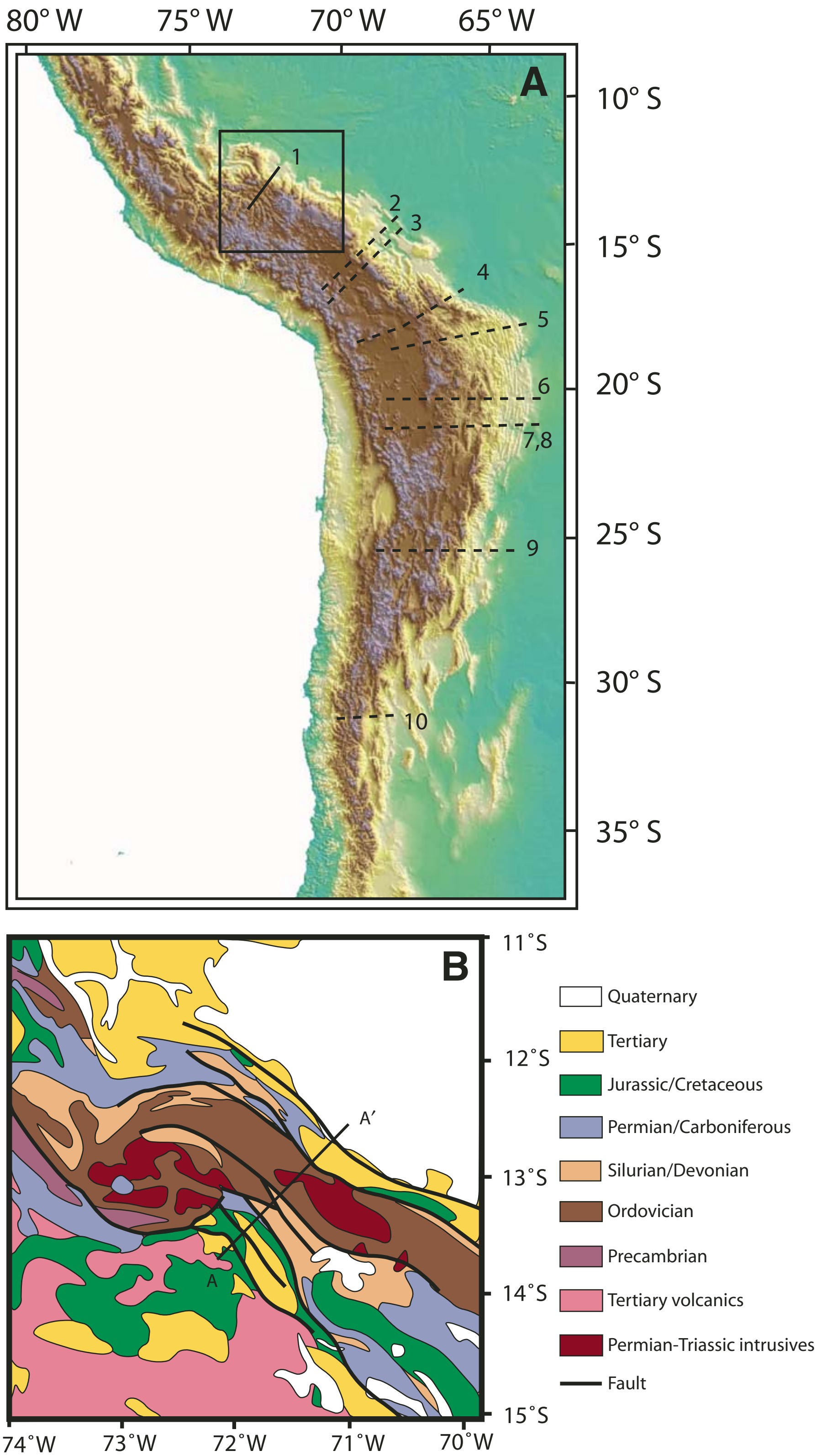


Figure 1. (A) Color topography of the Andean Plateau and the Bolivian orocline. Box indicates location of B and the study area. Black line represents the location of the cross-section line. Numbers and dashed lines refer to cross sections discussed in text and in Figure 8. The northern boundary of the Altiplano is located within the black box. (B) Generalized geologic map of the northern margin of the Andean Plateau.

of thickened felsic crust underneath the Andean Plateau (Zandt et al., 1994; Beck et al. 1996; Dorbath and Granet, 1996; Beck and Zandt, 2002) or research into the causes of vertical-axis rotations (Isacks, 1988; Kley, 1999; Lamb, 2001; Rousse et al., 2005; Roperch et al., 2006) is documented shortening variations around the Bolivian orocline. Although data on vertical-axis rotation (Arriagada et al., 2008) and internal deformation and strain (Kwon and Mitra, 2004) are required to complete a 3-D kinematic model, 2-D shortening estimates that span the region of interest are fundamental. While shortening estimates through Bolivia are abundant and fairly complete, shortening estimates through Peru are few and limited to the frontal Subandean zone (Sébrier et al., 1988; Kley and Monaldi, 1998; Gil et al., 2001; Gil, 2002) (Fig. 1). Shortening estimates from the northern margin of the Andean Plateau are particularly critical because this shortening influences the amount of material available for implied southward motion of material toward the axis of the bend in plane-view kinematic models (Kley, 1999; Hindle et al., 2005). Additionally, an understanding of how rotation of the northern limb is accomplished depends strongly on the details of how shortening estimates vary along strike.

If isostatic balance of a thickened crust is the primary control of high elevations through the Central Andes, then changes in shortening along strike would result in changes in magnitude and width of surface uplift. This suggests that regions with more shortening should have a wider extent of high-elevation topography than areas with less shortening (Kley and Monaldi, 1998), regardless of the exact time at which high elevations were achieved (Gregory-Wodzicki, 2000; Lamb and Davis, 2003; Garzione et al., 2006; Molnar and Garzione, 2007). This paper presents new shortening estimates from southern Peru and evaluates the magnitude of shortening needed to account for the crustal thickness of the plateau and rotation of the northern limb of the Bolivian orocline. In addition, it addresses the robustness of shortening estimates across orogens and evaluates the factors (dip of décollement, displacement, how space is filled) that have the largest influence on shortening estimates. Although we give maximum and minimum estimates in addition to our preferred estimate, our preferred estimate is considered most viable in that it bests honors available geologic data and is kinematically permissible.

GEOLOGIC BACKGROUND

The Central Andes are commonly divided into four geomorphic zones: the Western Cordillera, which is dominated by a Tertiary volcanic arc related to subduction of the Nazca plate; the Altiplano, which is an internally drained basin containing several kilometers of Tertiary deposits and has its northern limit in southern Peru; the Eastern Cordillera, which contains a fold-and-thrust belt with both east- and west-verging thrusts; and the Subandean zone, which is currently the most tectonically active part of the mountain belt. The Andean Plateau includes the high peaks of the Eastern Cordillera, the Altiplano, and the Western Cordillera. It is commonly defined as a broad area of internally drained basins and moderate relief with average elevations of ≥3 km (Isacks, 1988). The northern boundary of the Andean Plateau is located in southern Peru and is associated with an abrupt transition in the width of the high-elevation (>3 km) portion of the Andes, from ~350 km across the orogen at 14°S to ~150 km across at 11°S (Fig. 1). The map pattern of Paleozoic through Tertiary stratigraphy curves in this transition region, and the strike of folds and faults bends from northwest-southeast to more east-west. This bend defining the northern edge of the Andean Plateau is also known as the Abancay deflection, or Vilcabamba hinge (Sempere et al., 2002; Roperch et al., 2006). A transition in the thickness of units also occurs across the northern boundary of the Andean Plateau, where there is a much thicker early Paleozoic section, and thinner Mesozoic section in the plateau region (~14°S) compared to north of the plateau (~11°S) (Carlotto et al., 1996; Megard et al., 1996). The sedimentary distribution is attributed to Permian-Triassic rifting (Megard, 1978; Sempere, 1995; Sempere et al., 2002; Perello et al., 2003).

In southern Peru, near the northern boundary of the Andean Plateau, the volcanic arc of the Western Cordillera is ~270 km wide. The width of the volcanic arc is much narrower to the north and south of this region, ~60 km wide in central Peru and ~140 km wide in northern Bolivia. The Subandean zone is also fairly narrow in southern Peru, ~35 km, compared to ~50 km in northern Bolivia. In southern Peru, there is no Interandean zone, a region of transitional topographic and structural elevations that is sandwiched between the Subandes and Eastern Cordillera (Kley, 1996), as there is further to the south in Bolivia.

Most of the shortening in the Andean fold-and-thrust belt is thought to have occurred from ca. 40 Ma to present, although age constraints are from more data-rich regions in Bolivia (Elger et al. 2005; Horton, 2005; McQuarrie et al., 2005, 2008; Barnes et al., 2006, 2008; Gillis et al., 2006; Ege et al., 2007). Limited ^{40}Ar/^{39}Ar and apatite fission-track (AFT) thermochronology in southern Peru suggests cooling from ca. 37 to 18 Ma (Kontak et al., 1990), very similar to that recorded farther south (Barnes et al., 2006, 2008; Gillis et al., 2006; Ege et al., 2007). Early work in the Andes (e.g., Megard, 1978) proposed significant pre-Andean (late Paleozoic) deformation of Peru and Bolivia based on angular unconformities between Devonian and Carboniferous rocks located ~12°S in the Eastern Cordillera. Although late Paleozoic pre-Andean deformation has been documented in southernmost Bolivia (21°S) (e.g., Müller et al., 2002), the presence of conformable Jurassic and younger strata on Paleozoic strata exposed in numerous synclines throughout the Eastern Cordillera from 20°S to 17°S indicates that ~20.5°S is the northern limit of pre-Andean deformation in Bolivia (Sempere, 1995; McQuarrie and Davis, 2002). Documenting the absence of pre-Andean deformation from 14°S to 12°S is difficult due to large regions of the Eastern Cordillera (~70 km) that have only Devonian and older rocks exposed. Where present, Permian and younger strata parallel the older Paleozoic units, suggesting Andean-age deformation. However, at the outcrop scale, the Permian-Triassic igneous intrusive rocks do not appear to be folded along with the highly folded older Paleozoic strata, suggesting that (1) some or all of the outcrop-scale deformation observed in the older Paleozoic strata may be pre-Andean in age, or (2) deformation within the intrusive rocks is accommodated differently (Gillis et al., 2006; McQuarrie et al., 2008).

Kley et al. (1999) argued that changes in the amount of along-strike shortening in the Andes are related to changes in structural style and preexisting stratigraphy; specifically, high amounts of shortening in thin-skinned fold-and-thrust belts correlate to thick preexisting stratigraphic sections, and low amounts of shortening in thick-skinned fold-and-thrust belts correlate to thin preexisting stratigraphic sections. In Peru, there is a dramatic change in the stratigraphic architecture and structural style at ~12°S. Seismic sections and associated cross sections across this transition illustrate a progressive change from a thick-skinned fold-and-thrust belt in the narrow northern section of the Andes to a thin-skinned fold-and-thrust belt to the south in the wide Andean Plateau region (Gil et al., 2001; Gil, 2002).

BALANCED CROSS SECTION

Methods

Geologic Mapping

Geologic mapping was conducted at 1:100,000 scale in southeastern Peru between 12°S and 14°S and 71°W and 73°W (Fig. 2), with the goal of cross-section construction. The mapping provided information on lithology, the distribution of rock units, and the orientation of bedding, foliation, and faults. In locations where faults were not exposed, stratigraphic separations were assumed to be the result of Tertiary thrust faulting. Data collected during field work were combined with preexisting geologic maps to create a regional map at 1:250,000 scale, which was then used as the basis for the construction of a balanced cross section (Fig. 3).

Cross-Section Construction

The cross-section line extends from the eastern edge of the volcanic arc to the undeformed foreland (12.5°S–14°S). The roads on which data were collected do not follow a straight path perpendicular to strike, so the cross section was broken up into three segments to allow the cross section to follow the mapped geology and still be perpendicular to the structural trend (Fig. 2). The breaks in the section were chosen in locations with continuous along-strike structures, allowing the section to be projected north or south along strike.

The depth to basement and the slope of the décollement toward the orogen at the undeformed eastern edge of the cross section were projected from a series of seismic-reflection profiles located 50–200 km to the south (Gil, 2002). The continuity of basement dip along strike over 200 km and similar strata (and strata thickness) involved in the deformation suggest a similar basement dip along our line of section. Based on the seismic-reflection profiles, we assumed a 10 km depth to basement with a 2° dip. Since the basement-cover interface is one of the major décollements in many fold-and-thrust belts, including this one, the décollement dip in the hinterland was also assumed to be 2°–4°. An increase in dip toward the hinterland has been inferred in other cross sections through the Andes (Müller et al., 2002; McQuarrie, 2002a) and may be due to the weight of the basement fault on the décollement.

Cross-Section Balance

The balanced cross section was constructed using the sinuous bed method, where bed line length and area are conserved (Dahlstrom, 1969). All motion was assumed to be within the plane of the cross section with no strike-slip or oblique motion causing material to move through the plane of the section. A pin line was set in the undeformed foreland basin on the eastern edge of the cross section as the reference point for the restoration.

As described in the following section on stratigraphy, mechanically weak units through this portion of the Andes are commonly highly folded on an outcrop scale, which is impossible to accurately represent on the scale of the cross section. The cross section as drawn is a

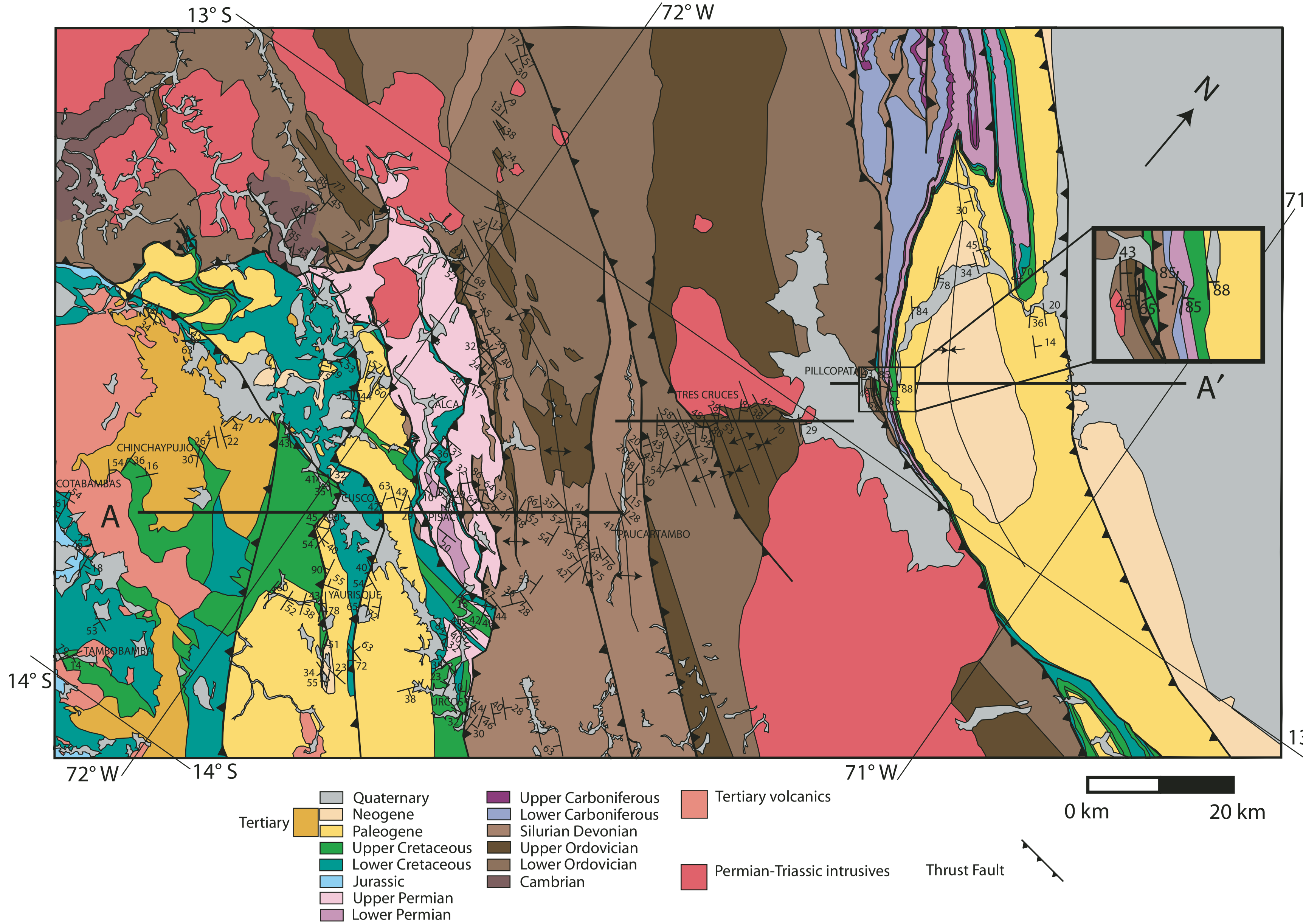
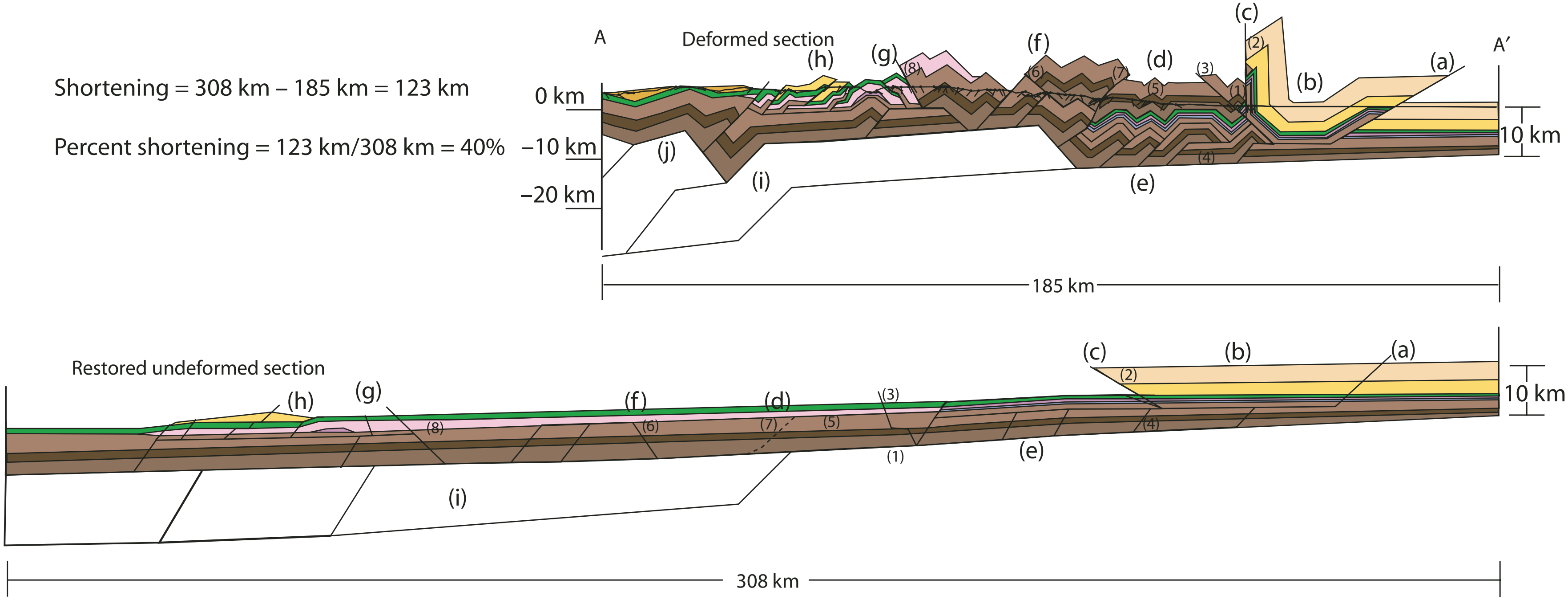


Figure 2. Geologic map of the northern margin of the Andean Plateau. The map data are from Marocco (1975); Sanchez and León (1994); Carlotto et al. (1996); Vargas and Hipólito (1998); León et al. (1999); and our own mapping.



Annotations

1. Large regional east-vergent thrust fault observed in the field placing Lower Ordovician rocks on top of Cretaceous rocks. The fault is regionally continuous in map view and we hypothesize that the older Permian-Triassic intrusive rocks allowed the thrust sheet to behave as a mechanically strong, single thrust. The fault is older than the Tertiary basin deposits (b) and west-vergent fault (2). The fault terminates in a hanging-wall cutoff against Cretaceous units (1).

2. West-vergent thrust fault displacing the large older regional east-vergent thrust fault. The fault contact was observed in the field. The cutoff angle is not preserved but this fault terminates into the regional east-vergent fault (1) to the north and south of the cross section line suggesting displacement is minimal. The displacement shown on this fault was drawn to maintain an equal amount of shortening in the units both above and below the detachment in the Devonian.

3. Young west-vergent thrust fault displacing the older regional east-vergent thrust fault.

4. Duplexing of the underlying stratigraphic section provides a mechanism to explain the larger-scale folding observed at the surface, allow equal magnitudes of shortening in the units above and below the detachment in Devonian age rocks, and also fill the space between the Paleozoic rocks mapped at the surface and the basal detachment.

5. The rocks in this region are folded at a broad scale indicated by changes in stratigraphic levels, as well as tighter, isoclinal map-scale folding. The cross section honors strike-and-dip measurements that displayed more regional folding, which we associate with Andean age deformation. The wavelength of folds measured in the field are too small to be shown at the scale of the cross section, given the thickness of the Paleozoic section. The relationship between folding in the Paleozoic section and the Permian-Triassic intrusive rocks was not observed, in outcrop.

6. The cutoff for this west-vergent thrust fault is not preserved and additional displacement is possible. However, the magnitude of displacement on the original detachment is already large (60 km, 15 km is west-verging), and no other fault in this region has substantial displacement through the erosion surface.

7. Out-of-sequence east-vergent thrust fault. Faulting occurred after movement on west-vergent thrust fault (6) and before duplex faulting (4).

8. Steep fault ramps shown here and at other locations on the cross section were drawn in order to maintain back-limb dip measurements. Steep ramps are a common feature throughout the Central Andes (Sheffels, 1990; McQuarrie and DeCelles, 2001; McQuarrie and Davis, 2002; Muller et al., 2002)

Figure 3. Balanced cross-section A-A′, extending from the eastern edge of the Western Cordillera to the undeformed foreland. Deformed length is 185 km, and restored undeformed length is 308 km, resulting in 123 km of shortening, or 40%. Letters refer to items discussed in the text; numbers refer to annotations. Stratigraphic color key is given in Figure 2.

compromise between stratigraphic contacts and dip data, and hence the dip data that represent outcrop-scale folding are not always observed in the cross section. The cross section is drawn to depict the larger-scale regional folding, as supported by changes in stratigraphy, regional map patterns, and orientation of strata in mechanically stronger units.

The critical data for determining the magnitude of shortening are (1) the amount of shortening required by exposed surface structures, (2) the thickness of the strata prior to deformation, (3) the area between the surface geology and the depth to the décollement that needs to be filled, and (4) the way in which this area is filled (i.e., is it filled by doubling or repeated stratigraphy or doubling or repeating basement). Drill-hole data and reflection seismology provide constraints on both the strata involved in filling space above a basal décollement as well as the geometry of the structure involved. Since these data are not available along our line of section, we evaluate the robustness of our section by looking at two extremes, (1) filling the available space with as much basement as possible (minimizing the shortening estimate), and (2) filling the available space with repeated sedimentary rocks (maximizing the shortening estimate). Our preferred cross section and shortening estimate lies between these end members, is kinematically the most viable, and thus most accurately reflects the amount of shortening in this region.

Stratigraphy

The age of the units observed along the cross-section line ranges from early Ordovician shale and sandstone to late Tertiary synorogenic sediments. Total stratigraphic thickness of pre-Tertiary sediments ranges from ~5 km in the foreland to ~10 km in the hinterland. Stratigraphic thicknesses are determined by standard geometric techniques utilizing the location of stratigraphic contacts observed in the field during geologic mapping, the dip of these beds, and topography (Compton, 1985). The thickness of the stratigraphic section in the foreland is constrained to the north and south of the cross-section line by seismic profiles. Since observed thicknesses are similar both to the north and to the south, it is assumed that they maintain a similar thickness in the study area, consistent with our field estimates. East-to-west thickness variations are based on both changes in mapped strata thicknesses as well as variations inferred from the seismic profiles. Generally, we find that Paleozoic through Mesozoic units thicken gradually to the west, as documented farther south in Bolivia (McQuarrie, 2002a). Because the names of formations vary between map locations in the region, we describe the stratigraphy along the cross-section line based on mechanical stratigraphy and formation age.

Ordovician

The Lower Ordovician units are 700–2500 m thick and are predominantly composed of blue-gray shales and quartzites that are locally metamorphosed near shallow-level plutons. Overlying the Lower Ordovician shale, there is a 700–2000-m-thick Upper Ordovician quartzite (Fig. 4). The Upper and Lower Ordovician units both behave as mechanically strong formations with little outcrop-scale deformation.

Silurian-Devonian

The Silurian-Devonian stratigraphic section ranges from ~1000 m thick in the foreland to ~2500 m thick along the western margin of the cross section (Fig. 4). The Silurian-Devonian units are predominantly shale and siltstone and are compositionally similar to the Lower Ordovician shale. Due to the similarity in composition, the Silurian-Devonian units are sometimes depicted in unfaulted contact with the Lower Ordovician shales in many geologic maps of Peru. Along our mapped transect lines, these units and contacts have been corrected; however, for areas located off our mapped transects, several of these contacts remain (Fig. 2). The Silurian-Devonian units are mechanically weaker than the Lower Ordovician shale and often display intense folding at an outcrop scale and fold wavelengths of a few centimeters to several meters. Mesoscale folds and a weak pencil structure parallel to fold axes all indicate that transport direction is perpendicular to the strike of faults and fold axes. The weak pencil structure also indicates a minor amount of layer-parallel shortening (Fig. 5). A major detachment horizon observed at the front of the fold-and-thrust belt is in a shale layer located in the middle of the Silurian-Devonian units. A mid-Silurian–Devonian detachment is also present in the western half of the Eastern Cordillera, but at a higher stratigraphic level.

Carboniferous to Tertiary

Immediately above the Silurian-Devonian units, there are shallow marine limestone and sandstone of Carboniferous age, ≤300 m thick

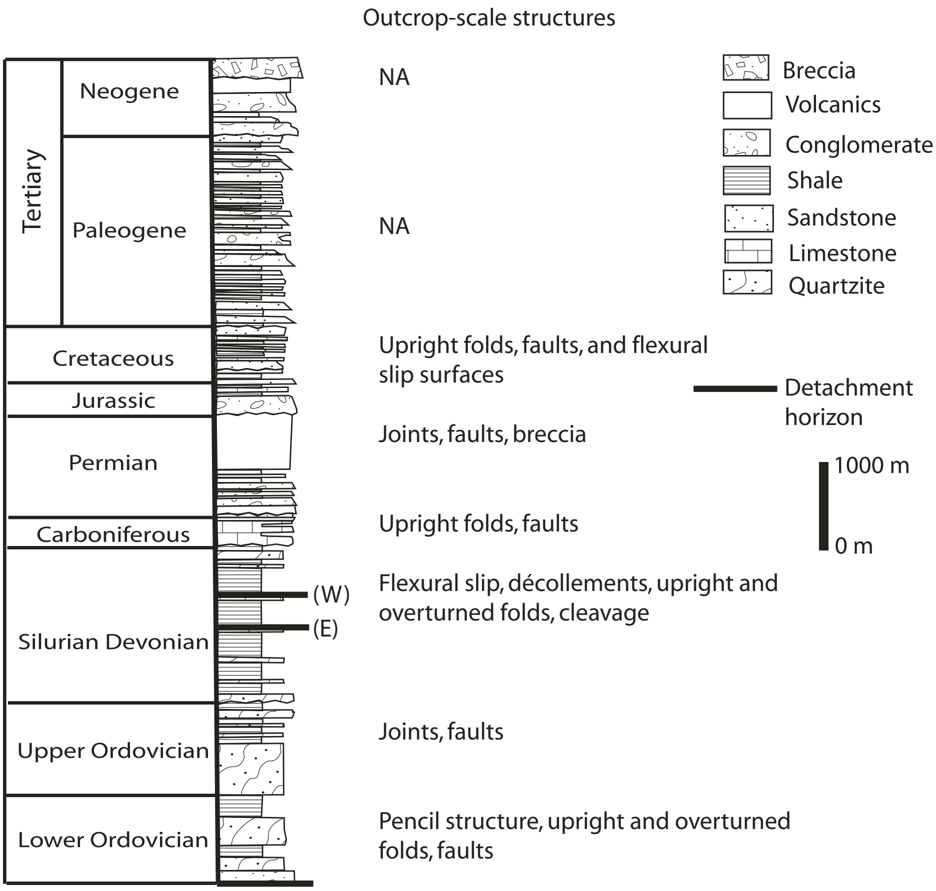


Figure 4. Stratigraphic column for cross-section line modified from Carlotto et al. (1996). W—Silurian-Devonian detachment in the west, E—Silurian-Devonian detachment in the east.

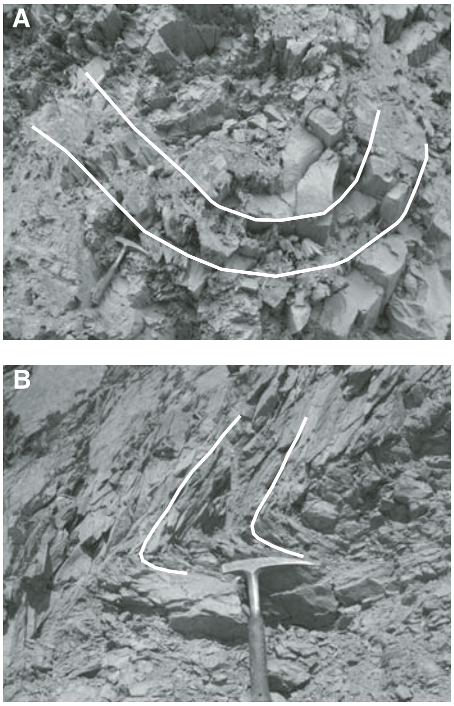


Figure 5. Field photos of mesoscale deformation and possible layer-parallel shortening (LPS) structures. (A) Weak layer-parallel shortening-induced cleavage in Lower Ordovician rocks; hammer for scale. Cleavage is into and out of the plane of the photo; bedding is highlighted in white. (B) Mesoscale folding in Silurian-Devonian siltstone; fine-grained sandstone bedding is highlighted in white. Note the brittle behavior in hinge.

(Fig. 4). The overlying Permian rocks consist of conglomerate, sandstone, mudstone, and volcaniclastic units that reach thicknesses of 2000 m and are thought to be syntectonic rift deposits (Sempere et al., 2002). Cretaceous limestone with interbedded shales and sandstone are 300–1200 m thick. The Carboniferous through Cretaceous units are exposed in the foreland and hinterland regions but are separated by an ~70-km-wide region where they are not exposed. Syntectonic Tertiary conglomerate and sandstone deposits are up to 6000 m thick in the foreland basin and up to 6000–8000 m thick in the Altiplano in southern Peru (Carlotto, 2002; Carlotto et al., 2005). The thickness of Tertiary strata in the intermontane basins along our cross section line is 2000–3000 m (Fig. 3).

Structure

The following sections discuss the map patterns characteristic of different sections of the fold-and-thrust belt and outline the constraints imparted by each area to an orogen-scale cross section through the Central Andean fold-and-thrust belt in this region (Figs. 2 and 3). The size of structures, their detachment horizons, and vergence directions vary from east to west. Letters on the cross section refer to specific structures discussed in this section (Fig. 3).

Subandean Zone

The Subandean zone in northern Peru is a 40-km-wide synclinal basin that has been segmented by narrow, thrust-faulted anticlines (Figs. 1B and 2). In the region of this study, the Subandean zone is a simple syncline (Fig. 3b) bound to the east and west by thrust faults (Fig. 3a and c). The western limb of the syncline (3c) is marked by a steeply eastward-dipping to vertical section of Upper Devonian through Tertiary rocks bound on the western side by a steep fault that places Upper Devonian rocks adjacent to Cretaceous rocks. We propose that the base of the Upper Devonian strata is the detachment for the thrusts on both sides of the syncline.

Eastern Cordillera (East)

The rocks exposed in the Eastern Cordillera are folded Ordovician through Devonian strata. The folds have wavelengths of 3–5 km and faults are rare and widely spaced (~40 km) (3d and f). Directly to the northwest and southeast of our transect, Lower Paleozoic strata are intruded by Permian-Triassic age granitic rocks. Although the map pattern indicates that these plutons are extensive, much of the region is heavily vegetated with few trails or roads, making the outcrop extent of the granites highly speculative. Our own observations suggest that the Paleozoic rocks are metamorphosed to slates and quartzites locally near the granites, increasing their mechanical strength.

In order to explain the presence of Lower Paleozoic rocks at the surface and account for the thickness of material required to fill the space between the basal detachment and mapped geology at the surface, we propose that the Ordovician through Cretaceous section is repeated by a large east-vergent thrust sheet, doubling the stratigraphic section (Fig. 3d). A duplex fault system in the Ordovician to Lower Silurian–Devonian units located below the large thrust sheet can explain the folding observed at the surface in the Upper Silurian–Devonian to Cretaceous units, as well as balance the amount shortening of strata below the Silurian-Devonian detachment (Fig. 3e) with that documented above (Figs. 3a and c). The detachment level in the Silurian-Devonian strata may behave mechanically different than the strata in the duplex due to the presence of Permian-Triassic plutons. The relationships between Triassic-age shallow plutons and the deformed Ordovician through Devonian rocks that host them are a challenge to document, even in the well-exposed high Eastern Cordillera of Bolivia (Gillis et al., 2006). Faults from Paleozoic strata though the granites have been traced but a lack of planar fabric, inhibited tracing folds (Gillis et al., 2006). Similar to the geometry we show here, the Bolivian Triassic intrusions are carried along a large displacement (~100 km) thrust fault, the base of which is ~6 km below the modern-day exposure of the plutons. The thrust sheet is broadly folded and interpreted to be the roof thrust over a duplex system in Ordovician strata (McQuarrie et al., 2008). In Bolivia or Peru, it is difficult to distinguish whether the plutonic bodies are thin (~6 km) and sill-like or whether they have been decapitated by Andean deformation. Although it is clear the Triassic plutons have been deformed by Andean deformation in Bolivia (Gillis et al., 2006), it is not clear whether the plutons predate or postdate folding in Peru.

Eastern Cordillera (West)

From the Eastern Cordillera to the west, the rocks exposed at the surface get progressively younger, from Silurian-Devonian to Permian, and then from Cretaceous through Tertiary. The western edge of the Eastern Cordillera contains a locally thick deposit of Permian strata (Fig. 3g) exposed in an anticlinal structure. It is unclear whether the Permian strata located on the western edge of our section were ever laterally continuous with Permian strata preserved in the Subandean zone. The local, thick nature of the Permian strata suggests it was deposited in an extensional basin (e.g., Sempere et al., 2002). The geometries of those bounding structures are now obscured, and the relationship between the early extensional structures and the later thrust faults is unclear. To the east of the anticline, there are two west-verging faults, one placing Permian strata on Cretaceous and the other placing Silurian-Devonian strata on Permian strata. On the west side of the anticline, the vergence direction returns to an easterly direction (Fig. 3h). Faults displacing Upper Devonian through Tertiary strata are closely spaced (5–10 km), and folds have a narrower wavelength than structures to the east. We propose that the surface faults sole into a décollement in the Upper Devonian strata to explain the more closely spaced structures and strata exposed at the surface. To explain the difference in elevation of exposed strata in this region (Fig. 3g) with respect to the foreland (Fig. 3b), we propose a basement thrust that underlies this part of the Eastern Cordillera and an upper detachment for this basement thrust that is the major décollement surface for the fold-and-thrust belt, at the base of the Ordovician.

Altiplano

The western boundary of our cross section extends across the northern edge of the internally drained Altiplano basin (Fig. 2). West of exposed Permian strata (Fig. 3g), the Altiplano basin contains an ~2000-m-thick section of Tertiary sedimentary basin deposits (Carlotto, 2002). Tertiary volcanic rocks intrude and cover Cretaceous and older rocks in the Western Cordillera and western portion of the Altiplano, obscuring much of the evidence for prevolcanic deformation. For this reason, we have chosen to terminate our balanced cross section where it reaches a large Tertiary plutonic complex, ~35 km west of the Tertiary basin deposits of the Altiplano (Fig. 2). In our cross section, the plateau portion of the fold-and-thrust belt is underlain by basement-involved thrust sheets that fill the space between the basement décollement (Fig. 3i) and the minimum depth to the basement (Fig. 3j). Although the geometry of proposed basement thrusts varies widely, basement thrusts in cross sections through the Central Andes are used to provide minimum estimates in shortening values while accounting for the area between the main décollement and the elevation of sedimentary rocks exposed at the surface (Kley, 1996; Baby et al., 1997; Gil et al., 2001; Müller et al., 2002; Gil, 2002; McQuarrie, 2002a; Elger et al., 2005).

SHORTENING ESTIMATES AND UNCERTAINTIES

Restoration of the balanced cross section provides a preferred shortening estimate across the northern section of the Andean Plateau (12–14°S). The deformed cross section has a length of 185 km. The preferred shortening estimate yields a 308-km-long undeformed section and a shortening estimate of 123 km, or 40% (Fig. 3).

A detailed description of the methods used in calculating the variability in the shortening estimate is included in Appendix 1, and the end-member variations are depicted in Figure 6. Factors that influence the uncertainties include basement thrusts, depth to décollement, and fault displacement in the absence of matching footwall and hanging-wall cutoffs.

Basement Thrusts

The fundamental control of shortening estimates in any thin-skinned cross section is the amount of area between the mapped surface geology and the basal décollement and the way that area is filled. Using basement thrusts to fill the area decreases the shortening estimate (e.g., Kley, 1996; McQuarrie, 2002a). If the doubled stratigraphic section in our cross section (Figs. 3d and e) is replaced by a basement thrust, the amount of shortening in the cross section would decrease by ~35 km, yielding a shortening estimate of 88 km or ~30% (Fig. 6; Table 1). If on the other hand, all the area above the basal décollement was filled by doubling the sedimentary section, and no basement thrusts were involved, then the amount of shortening could be increased by 120 km, yielding ~240 km of shortening or ~55%. The geometric difficulties with the minimum and maximum shortening estimates are annotated on Figure 6. Our preferred estimate includes a basement thrust in the hinterland and duplexed sedimentary rocks toward the foreland. The basement thrust fills space in the hinterland, while the duplexed sedimentary rocks are kinematically more compatible with both the style of deformation in the overlying Paleozoic section as well as the amount of deformation in the deformed Upper Devonian to Tertiary section in the Subandean zone.

Dip of Décollement

A change in the dip of the décollement also affects the amount of area that needs to be filled in a cross section. An increase in the dip of the décollement generates more area that needs to be filled by shortening. For example, an increase in the dip of the décollement of 3° from a 2° dip to 5° dip over a 185-km-long cross section would result in the addition of 900 km^2 of material will increase the amount, correlating to an increase in shortening of ~90 km (Table 1). A decrease in the dip of the décollement by 1° would result in the loss of 300 km^2, correlating to a decrease in shortening of ~30 km. Our preferred estimate has a 2° dip of the décollement in the foreland and a 4° dip of the basement in the hinterland. The 2° increase may be due to the weight of the basement fault on the décollement. This increase in dip toward the hinterland has been inferred in other cross sections through the Andes (Müller et al., 2002; McQuarrie, 2002a), and it has been imaged in different orogens (e.g., Hauck et al., 1998).

Displacement

An increase in the amount of displacement on faults and accompanying erosion of hanging-wall material increase the amount of shortening. Mapped strata, orientations of strata, and the preservation of hanging-wall and footwall stratigraphic cutoffs constrain the amount of displacement possible on faults. The value for increased displacement on faults was not mathematically calculated, but assumptions were made based on the existing length of thrust sheets with unpreserved cutoff angles and kinematic viability. A reasonable estimate of the amount of increased shortening that could result from increased displacement on faults in this cross section is ~10 km (Fig. 6).

Minimum and Maximum Shortening Estimates

If we combine the uncertainties due to basement-thrust involvement, décollement dip, and displacement, the minimum shortening estimate for the cross section would be ~58 km or 25%, and the maximum shortening estimate would be ~333 km or 65% (Table 2).

Comparisons to Previous Shortening Estimates

Previous shortening estimates through the Peru portion of the Central Andean fold-and-thrust belt are sparse and have been concentrated almost entirely in the Subandean zone, proximal to the Eastern Cordillera (Sébrier et al., 1988; Gil et al., 2001; Gil, 2002), and in the Altiplano (Carlotto, 2002). A challenge in comparing a series of shortening estimates is identifying the final length (present-day width) of each area being compared. Sébrier et al. (1988) proposed the first cross section through the Subandean zone of Peru. A line length balance of this section indicates 20 km of shortening (Kley and Monaldi, 1998; Kley et al., 1999) (Table 3). A more recent study by Gil (2002) looked at a series of cross sections through the Subandean portion of the Peruvian Andes. Gil's sections include the Subandean zone and a portion of the Eastern Cordillera referred to as the Internal Subandean zone. The total estimate of shortening at ~12.5°S is 121 km; 40 km of this is in the Subandean zone proper, and 81 km of this is in the Eastern Cordillera (Table 3). Of the 40 km estimate in the Subandean zone, 22 km of that is subsurface duplexing of a 1.5-km-thick package of Ordovician and Silurian strata; the Devonian through Tertiary equivalent is assumed to be eroded. A minimum shortening estimate for the Subandean portion of Gil's section would be 18 km. Although the main focus of Carlotto's study was on the Altiplano, he used both balanced cross sections and line length balance of structures to determine a total shortening estimate across the northern margin of the Andean Plateau in Peru. He proposed 20 km of shortening in the Subandean zone, 33 km of shortening in the Eastern Cordillera, and 13 km of shortening of the Altiplano. By comparing only minimum shortening estimates across the Subandean zone, all four estimates are very consistent, and they are significantly less than the 70–80 km of shortening in the

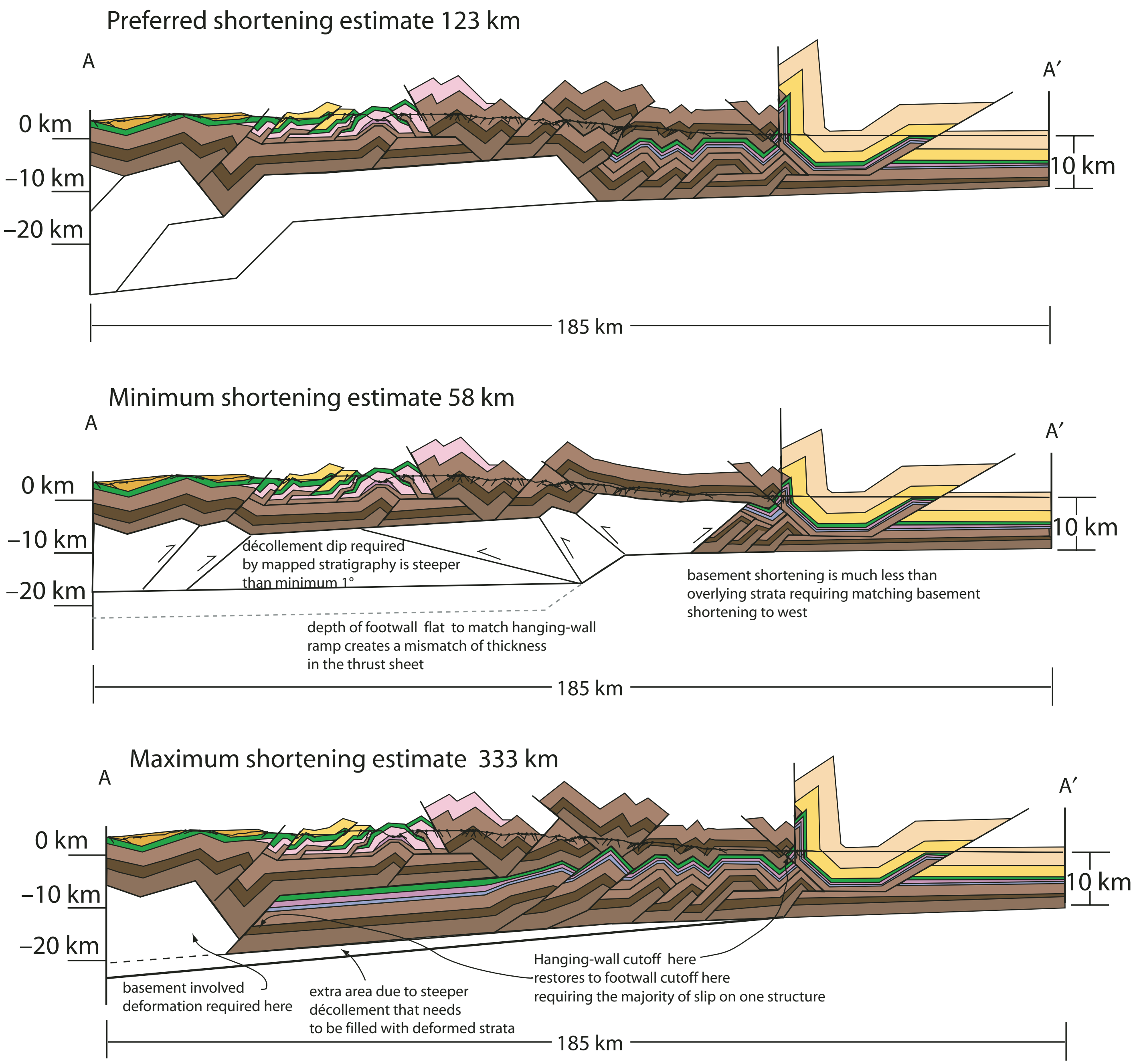


Figure 6. Variations in the way shortening is accommodated in our preferred, minimum, and maximum shortening estimates. Annotations indicate problems with the kinematics in the maximum and minimum shortening estimates. Stratigraphic color key is given in Figure 2.

Subandean zone of Bolivia (Dunn et al., 1995; Baby et al., 1997; McQuarrie, 2002a; Barke and Lamb, 2006; McQuarrie et al., 2008).

CRUSTAL THICKENING BUDGET

To determine if the shortening in our cross section accounts for the crustal area of the orogen, we predict the amount of shortening that would be needed to account for the crustal area in Airy isostatic equilibrium and then compare the shortening predictions to our measured values. We made predictions assuming initial crustal thicknesses of 35, 40, and 45 km. Due to the history of Permian-Triassic rifting in the region, an average global crustal thickness of ~35–40 km (Bassin, et al., 2000), and a crustal thickness of ~35–40 km in the foreland of the Andean Plateau, we assume that a 35–40-km-thick crust best represents the starting conditions. Previous work, however, has assumed a 40–45 km initial crustal thickness, so we include calculations using this initial condition for comparison (e.g., Beck et al., 1996; Kley and Monaldi 1998).

Isostasy Crustal Area Calculations

South of Peru, at ~20°S, broadband seismic studies suggest that the topography of the Andean Plateau is in Airy isostatic equilibrium

TABLE 1. MAXIMUM AND MINIMUM SHORTENING ESTIMATE VARIATIONS FROM PREFERRED SHORTENING ESTIMATE

	Decreased shortening (km)	Increased shortening (km)
Basement thrusts	–35	+120
Décollement dip	–30	+90
Displacement	0	+10
Total	**–65**	**+210**

Note: Values subtracted or added to preferred shortening estimate.

TABLE 2. MAXIMUM AND MINIMUM SHORTENING ESTIMATES

	Difference (km)	Shortening estimate (km)	Percent shortening (%)
Preferred	0	123	40
Minimum	123–65	~58	25
Maximum	123 + 210	~333	65

TABLE 3. COMPARISON OF SHORTENING ESTIMATES THROUGH PERU

	Altiplano	Eastern Cordillera (west)	Eastern Cordillera (east), Interandean zone, internal Subandean zone	SAZ	Total
Sébrier et al., 1988				20 km	20 km
Gil, 2002			81 km	40 km	121 km
Carlotto, 2002	13 km	13 km	20 km	20 km	66 km
This study	15 km	31 km	60 km	17 km	123 km

with a 60–70-km-thick crust (Beck et al., 1996; Beck and Zandt, 2002). Seismic studies in Peru are limited; however, a local study in the Western Cordillera of southern Peru is also consistent with an ~70-km-thick crust (Cunningham et al., 1986). Assuming that the Andean Plateau of southern Peru is in isostatic equilibrium at 3.8 km elevation, and the original undeformed crust was at 0.5 km elevation, we can estimate the total crustal cross-sectional area within the boundaries of the cross section. Using a crustal density of 2.8 g/cm^3, mantle density of 3.3 g/cm^3, and an initial crustal thickness of 35 km, the total crustal cross-sectional area within the boundaries of the cross section needed to produce the 3–4-km-high topography is 9080 km^2. The cross-sectional crustal area of the entire plateau in this region is 24,300 km^2 (Fig. 7; Table 4). If we instead assume a 40-km-thick undeformed crust, the area of the cross section is 10,005 km^2 and the area of the plateau is 26,550 km^2, whereas for a 45-km-thick initial crust, the cross-section area is 10,930 km^2, and the plateau area is 28,800 km^2 (Table 4).

Amount of Erosion

From our balanced cross section, we can predict the maximum, minimum, and most likely amount of material removed by erosion. We calculated this value by counting the 1 km^2 blocks of material projected above the topographic surface shown on the cross section (Fig. 3). This is a robust estimate of amount of material removed, because the slip on all but three faults is constrained by preserved hanging-wall cutoffs. When faults are not constrained, we chose the minimum amount of displacement to honor the observed stratigraphic contacts, allow kinematic compatibility, and balance. In map view, only one fault with unconstrained cutoffs is continuous along strike. All others lose displacement along strike and/or die out in map-scale folds, limiting their total displacement. If we assume that no Carboniferous through Cretaceous sedimentary strata were deposited where they are not directly observed, we obtain a minimum erosion estimate of 385 km^2. Maximum estimates of erosion are calculated by assuming increased displacement on permissible faults (up to 10 km) as well as assuming that a thick section of Carboniferous-Cretaceous rocks was deposited across the entire extent of the cross section, even in the center where they are not observed. Variations in Carboniferous-Cretaceous rock thickness (removed via erosion) were calculated from 2 km, 3 km, or wedge-shaped 2–3-km-thick sections. If we assume that a 2-km-thick section of strata has been removed by erosion, we get an area estimate of 760 km^2, and a 3-km-thick section of strata removed results in an estimate of 950 km^2. If we instead assume a 2–3 km wedge-shaped section of the Carboniferous-Cretaceous was eroded in addition to the material projected above the topography line in Figure 3, then the amount of material removed by erosion is 850 km^2, correlating to an 8% loss of the total crustal area (Table 4).

Shortening Predictions

Shortening estimates were also calculated based on predictions from isostasy calculations and erosion. The amount of erosion was added to the crustal area in isostatic equilibrium and then divided by the undeformed crustal thickness in order to obtain estimates for initial length (Fig. 7; Table 4). For the entire length of the plateau, we assumed an 8% loss of material due to erosion (see previous section). Using a 35-km-thick crust, the amount of shortening across the entire plateau should be 300 km, which is ~180 km more shortening than our preferred cross-section shortening estimate of 123 km, but it is approximately equal to our maximum shortening estimate of 333 km. By looking only at the area under the cross section, a 35-km-thick crust with an area of 9930 km^2 yields a 283-km-long initial, undeformed length. This length is 25 km less than the 308 km undeformed length of our cross section (Fig. 7). The difference in area between the measured and predicted shortening (if the cross section is in isostatic equilibrium) correlates to an excess of 875 km^2 of material in our cross section. Thus, using our preferred estimate, only 875 km^2 of the 15,200 km^2 cross-sectional area of the Western Cordillera can be accounted for by shortening between the volcanic arc and undeformed foreland, leaving 14,325 km^2 of area unaccounted for.

BOLIVIAN OROCLINE

Isacks (1988) was the first to hypothesize that the Bolivian orocline formed as a result of along-strike changes in the magnitude of Neogene horizontal shortening in the Subandean zone, accompanied by tectonic rotations in the forearc. However, reconciliation of predicted and/or actual shortening estimates to predicted and/or actual vertical-axis rotations has proven complicated (Kley, 1999; Lamb, 2001; Roperch et al., 2006; Arriagada et al., 2008). To a first order, shortening estimates indicate a decrease in shortening to both the north and south away from the hinge of the orocline (Kley and Monaldi, 1998; McQuarrie, 2002b; McQuarrie et al., 2008; this study), which matches both the decreasing width of high topography away from the bend as well as the sense of measured vertical-axis rotation. Large-magnitude rotations are located in the Andean forearc and along the Eastern Cordillera–Altiplano boundary. On average these vertical-axis rotations, 37° ± 15° in southern Peru and 29° ± 19° in northern Chile (Taylor et al., 2005; Roperch et al., 2006; Arriagada et al., 2006, 2008, and references therein), surpass the predictions of Isacks (1988) by 20°–25°.

The timing of the forearc rotations is much older than the Neogene age initially proposed by Isacks (1988). The large observed forearc rotations decrease in magnitude from Paleocene to late Oligocene and are not found in Miocene ignimbrites, indicating the rotations occurred

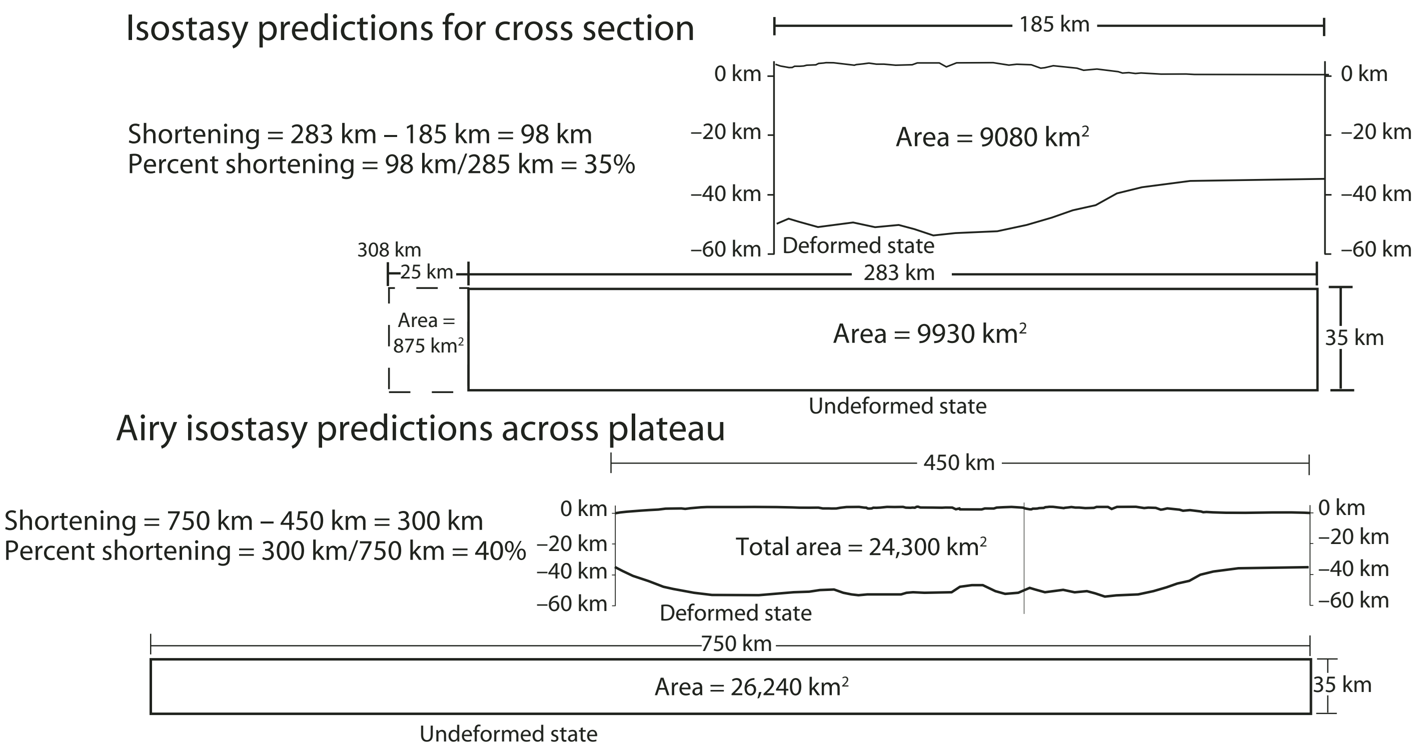


Figure 7. Calculations of crustal area for the length of the balanced cross-section line and across the entire plateau assuming Airy isostasy with a crustal density of 2.8 g/cm^3, mantle density of 3.3 g/cm^3, and a 35-km-thick initial undeformed crustal thickness. An estimate of the amount of material lost to erosion was added to the crustal area of the deformed section, and the entire area was divided by 35 km to get (1) the initial starting length and (2) to determine the amount of shortening. Along the cross-section line, crustal area, 9080 km^2, plus the area lost due to erosion, 850 km^2, equals an initial area of 9930 km^2. If we divide the area by 35 km (crustal thickness), we get an initial length of 283 km. This length is 25 km less than the original length from balanced cross section (308 km) and is equivalent to 875 km^2 less area. For the entire plateau, the current area of 24,300 km^2 plus an 8% loss of material due to erosion of 1940 km^2 equals an initial area of 26,240 km^2, which, when divided by 35 km, equals an initial length of 750 km.

in the Eocene-Oligocene, perhaps as a consequence of differential shortening focused in the Eastern Cordillera (Roperch et al., 2006). The timing of vertical-axis rotation directly overlaps with the 45–25 Ma age of shortening in the Eastern Cordillera of Bolivia (Barnes et al., 2006, 2008; Ege et al., 2007; McQuarrie et al., 2008). If along-strike variations in crustal shortening have occurred systematically during Andean deformation, the amount of shortening along different profiles across the Central Andes can be used to calculate the associated magnitude of vertical-axis rotation. An increasingly robust data set on the ages and magnitudes of rotations along the Bolivian orocline (Richards et al., 2004; Rousse et al., 2005; Taylor et al., 2005, 2007; Arriagada et al., 2006, 2008; Roperch et al., 2006; Barke et al., 2007,) along with new estimates of shortening along the northern limb of the Bolivian orocline (McQuarrie, 2002a; McQuarrie et al., 2008; this study), allows us to compare the actual and predicted shortening variations and vertical-axis rotations along the northern limb of the Bolivian orocline.

TABLE 4. CRUSTAL AREA BUDGET

Shortening estimate	Erosion (km^2)	Initial thickness (km)	Cross-section area (km^2)	Erosion + cross-section area (km^2)	Undeformed length (km)	Shortening (km)	Percent shortening (%)
Preferred	850	35	9080	9930	283	98	35
		40	10,005	10,855	271	86	32
		45	10,930	11,780	262	77	29
Minimum	385	35	9080	9465	270	85	31
		40	10,005	10,390	260	75	29
		45	10,930	11,325	251	66	26
Maximum	1010	35	9080	10,090	288	103	36
		40	10,005	11,015	275	90	33
		45	10,930	11,940	265	80	30

Differential Shortening Estimates and Predicted Rotations

Table 5 lists the shortening estimates and the difference between those estimates for three cross sections through the northern limb of the Bolivian orocline. Early (45–20 Ma) shortening in the Eastern Cordillera and Interandean zone of Bolivia produced a 10 km difference in shortening between the cross section located at the hinge of the Bolivian orocline (18°S–17°S), and the cross section located between 15°S and 17°S, which would result in a vertical-axis rotation of only 2°. Between the cross sections at 12–14°S and 15–17°S, an 80 km difference in shortening in the Eastern Cordillera and Interandean zone suggests rotations as high as 9.5°. The combined ~11°–12° of vertical-axis rotation that the shortening estimates predict from 45 to 20 Ma is still significantly less than the 37° ± 15° measured by paleomagnetism (Roperch et al., 2006). However, the difference in shortening needed to

TABLE 5. DIFFERENTIAL SHORTENING AND PREDICTED ROTATIONS FOR THE NORTHERN LIMB OF THE BOLIVIAN OROCLINE

Region	Latitude (°S)	Shortening (km)	Difference (km)	Rotation (°)	Total (°)	Age
Eastern Cordillera and Interandean zone	12–14*	92				
			79	9.5		
	15–17†	171			11.5	45–20 Ma
			10	2		
	17–18†	181				
Subandes	12–14*	17				
			50	6		
	15–17†	66			7	< 15 Ma
			6	1		
	17–18†	72				
Altiplano	12–14*	15				
			25	3		
	15–17†	40			4.5	? <15 Ma
			7	1.5		
	17–18†	47				
Total (preferred shortening)			177 km	23°		45–0 Ma
Total (maximum shortening)			–131 km	11.5° cw		45–0 Ma
Total (minimum shortening)			219 km	30°		45–0 Ma

Note: cw—clockwise rotations; all other rotations counterclockwise.
*This study.
†McQuarrie (2002a).

produce 37° of vertical-axis rotation is ~370 km, highlighting the importance of other tectonic processes contributing to the rotation of the Andean forearc and along the western boundary of the Eastern Cordillera. Predicted young (younger than 15 Ma) vertical-axis rotations due to shortening in the Subandean zone are 1°, due to differential shortening between the cross sections at 17–18°S and 15–17°S, and 6°, due to shortening between the cross sections at 15–17°S and 12–14°S. Potential north-south shortening variations between young (younger than 15 Ma) deformation in the Altiplano region suggest an additional 4.5° of vertical-axis rotation (Table 5). The young predicted rotation of 7° in the Subandean zone and 11.5° total (Subandean zone and Altiplano) are compatible with the young, uniform vertical-axis rotations measured by Rousse et al. (2005) using paleomagnetic data. They calculated ~11° of vertical-axis rotation in basins within the Altiplano and rotations of 7.8° ± 4.8° in the Subandean basin since ca. 12 Ma. Although large cross structures (such as tear faults) may accommodate along-strike differential shortening, geologic maps at a variety of scales through the Central Andes do not show obvious candidates.

To calculate the predicted vertical-axis rotations, we used our preferred estimate of shortening of 123 km across the northern plateau margin. When compared to central Bolivia, the shortening difference is 177 km. This estimate predicts a total rotation amount of 23° (Table 5). If instead we use the maximum possible shortening (333 km), shortening in the Eastern Cordillera in Peru is greater than Bolivia by 131 km, producing 15° of *clockwise* rotation (all measured paleomagnetic vertical-axis rotations are counterclockwise in this region). Using our minimum estimate of 58 km, the differential shortening would be 219 km, with 30° of counterclockwise rotation. As emphasized by Pueyo et al. (2004), thrust slip paths may be oblique to structural trend in areas with rotation, and this has implications for estimating thrust shortening. In the Central Andes, the rotations measured in the fold-and-thrust belt are all small (6°–8°), while the shortening estimates are large (125–300 km). These combine to make the errors associated with not accounting for rotation less than 1 km (well below what we can begin to resolve). Even using the largest possible rotations (~30°), shortening errors associated with not accounting for rotations are on the order of 10%.

DISCUSSION

Mechanisms for Rotation

As has been suggested previously (Richards et al., 2004), along-strike variation of shortening can account for a part of the observed rotations, but not the large 37° ± 15° vertical-axis rotations measured in either the forearc or along the northern Eastern Cordillera–Altiplano boundary in Peru. We propose that as much as half of the vertical-axis rotations measured in the forearc and eastern Altiplano are not a result of differential shortening but rather the way shortening was accommodated in the inner bends of two developing oroclines. The first orocline is the pronounced bend in the South American continent and Andean mountain chain termed the Bolivian orocline (Carey, 1958). The second is the small orocline delineated by the curving of stratigraphy and structures around the northern margin of the Andean Plateau, referred to as the Abancay deflection (Sempere et al., 2002). We propose that the along-strike variations in horizontal shortening induced initial bending of the continental margin. As the bend developed, block rotations within the forearc region facilitated deformation in a tightening core and allowed for orogen-parallel transport of material toward the symmetry axis of the orocline (Arriagada et al., 2008). Thus, the forearc rotations reflect both bending due to differential shortening, ~12°, as well as block rotations (~25°–30°) that accommodate bending and eastward translation of the forearc. In addition to Bolivian orocline-induced rotations, the Abancay deflection also induces counterclockwise rotations. The role of the Abancay deflection in both the plateau and orocline development has been noted previously. The northern margin of the plateau acts as a hinge, decoupling rotation to the south from that to the north, as proposed by Lamb (2001) because of the proposed marked reduction in shortening magnitude at that boundary. Roperch et al. (2006) proposed that the Abancay deflection marks the northern boundary of both the plateau and the orocline because it is the northern boundary of large early rotations in the forearc region. Although the data necessary to evaluate a pronounce change in shortening, particularly between 45 and 25 Ma, do not exist, we can evaluate the magnitude of rotation induced by differential propagation of the fold-and-thrust belt north and south of the Abancay deflection. Over a distance of ~145 km, the Andean fold-and-thrust belt has propagated 130 km further eastward in the south than in the north. This differential propagation has the potential to induce rotations up to 40°, remarkably similar to the 50° ± 5° rotations measured by paleomagnetism in this region (Roperch et al., 2006). The rotations induced by a dramatic reduction in shortening at the northern margin of the plateau can be seen in the map-view reconstruction of the plateau produced by Arriagada et al. (2008).

Implications for Crustal Thickness

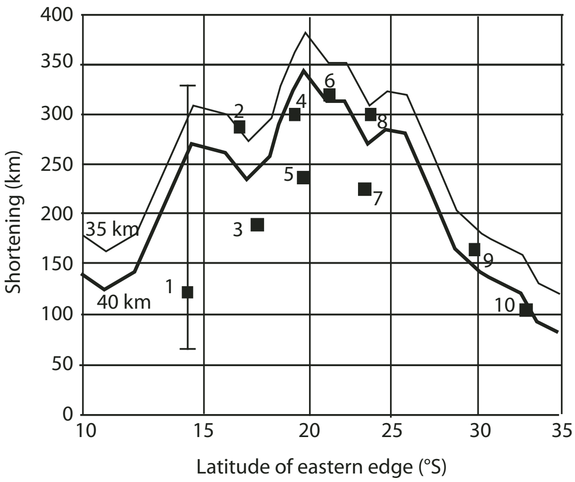
Our preferred shortening estimate from our balanced cross section is insufficient to account for the entire crustal area of the Andean Plateau. Assuming our maximum estimate (333 km) is correct, shortening in the cross section accounts for all of the crustal thickness. However, since it exceeds shortening in the Eastern Cordillera to the south, our maximum estimate would predict clockwise vertical-axis rotations between the hinge of the Bolivian orocline and its northern margin. The multitude of measurements of vertical-axis rotations in this region show >20° counterclockwise rotation (e.g., Arriagada et al., 2008, and references therein), making our maximum shortening estimate unlikely. If our preferred shortening estimate is correct, it requires additional material, either from unmeasured shortening in the Western Cordillera or flow from the south. Larger magnitudes of shortening in the center of the Bolivian orocline will increase crustal thickness and gravitational potential energy in this region, providing a mechanism for crustal flow from the axis of the orocline to the limbs (McQuarrie, 2002b; Yang and Liu, 2003; Ouimet and Cook, 2006). In the Central Andean Plateau in Bolivia, even the highest shortening estimates do not provide the amount of material needed to account for the crustal area of the northern Andean Plateau by lateral flow of material from the south, assuming a 35-km-thick crust (Fig. 8)(Allmendinger et al., 1997; Kley and Monaldi, 1998; McQuarrie, 2002a; McQuarrie et al., 2005). However, assuming an initial 40-km-thick crust in Bolivia and Peru, along-strike flow of material could provide for 3420 km^2 of extra material (McQuarrie, 2002a). Including the additional material, our preferred shortening estimate of 123 km is still ~60 km short of what is required to account for an isostatically balanced crustal thicknesses. This suggests that either there is a significant amount of hidden shortening (~60–150 km) in the Western Cordillera of the Andean Plateau, which has been covered by recent volcanism, or else a process other than shortening contributes to the 60–70-km-thick crust predicted from isostatic equilibrium and seismic studies.

In the Western Cordillera, Tertiary volcanics obscure much of the evidence for earlier deformation, but in the locations where Mesozoic sedimentary rocks are exposed, they are folded, faulted, and compressionally deformed, suggesting that they may have been significantly shortened prior to Tertiary volcanism. The young volcanic rocks make it impossible to construct balanced cross sections across the entire plateau to get a total shortening estimate. It is commonly assumed that the majority of shortening occurred in the Eastern Cordillera and Subandean zone (Kley, 1996; Kley and Monaldi, 1998; Müller et al., 2002; McQuarrie, 2002a). Kley and Monaldi (1998) suggested that deformation in the Western Cordillera is thick-skinned and only contains a small amount of shortening, less than 30%. If 30% shortening existed in the Western Cordillera of southern Peru, it would correspond to ~110 km of shortening. When added to our preferred estimate of shortening (123 km) in the Eastern Cordillera, the total shortening across the entire plateau would be 233 km, only ~70 km short of the required 300 km of shortening with a 35-km-thick undeformed crust, and ~34 km short of the required 267 km of shortening for a 40-km-thick undeformed crust. The combination of shortening in the Eastern Cordillera and Subandean zone of southern Peru, 30% shortening in the Western Cordillera, and along-strike addition of material from Bolivia (3000–3500 km^2) can potentially account for the entire crustal area of the northern Andean Plateau.

The extensive Tertiary volcanic cover present in the northern plateau suggests that magmatic processes may provide additional crustal material. Magmatic underplating can add mafic material to the base of the crust and increase crustal thickness, but in order to generate the felsic material consistent with the results of geophysical studies in the Andean Plateau (Beck and Zandt, 2002; Beck et al., 1996), differentiation of the underplated material into more and less mafic layers and subsequent loss of the more mafic layers, likely by delamination, would be required (e.g., Ducea and Saleeby, 1998). The wide (~270 km) volcanic arc that is unique to this portion of the Andes may be an indication that magmatic processes have a bigger role in accounting for crustal thicknesses in this region of the Andean Plateau than farther south in Bolivia.

Implications for Material Flow

Map-view restorations of the Central Andes show a convergence of motion vectors toward the axis of the orocline (Kley 1999, Hindle et al., 2002, 2005), with the magnitude of orogen parallel motion increasing away from the bend (Kley, 1999). Part of this orogen-parallel motion is a function of map-view modeling, which requires blocks to be internally rigid; thus, to combine shortening with rotation, or even to shorten an originally bent margin, requires either motion of material to the axis of the orocline, or extension along the outer arc of the bend (Kley, 1999) (Fig. 9). More detailed map-view reconstructions that take into account rotations of the forearc require less transfer of material into the axis of the orocline by about half (10%–15% versus 35%–40%; Arriagada et al., 2008). Figure 9B highlights the fact that the solution to the 3-D space problem created by shortening an orocline, without pronounced arc-parallel extension, requires counterclockwise rotation of material in the north and clockwise rotation of material in the south. This rotation is independent of shortening variations and



1 This study
2 McQuarrie et al. (2008)
3 Baby et al. (1997)
4 McQuarrie (2002a)
5 Sheffels (1990); Lamb and Hoke (1997)
6 McQuarrie (2002a)
7 Baby et al. (1997); Dunn et al. (1995); Kley (1996)
8 Kley (1996); Muller et al. (2002); Elger et al. (2005)
9 Grier et al. (1991); Coutand et al. (2001)
10 Allmendinger et al. (1990); Zapata and Allmendinger (1996)

Figure 8. Shortening estimates across the plateau from the volcanic arc to foreland between 10°S and 35°S. Thin black line on graph represents the amount of shortening needed to accommodate crustal area assuming Airy isostatic equilibrium and an initial undeformed thickness of 35 km. Thick black line represents amount of shortening assuming an undeformed thickness of 40 km. Many of the shortening estimates (1, 3, 4, 5, 7) fall short of the amount needed to account for the crustal area, including this study (1). Assuming a 40-km-thick initial undeformed crust, four of the cross sections provide more than enough shortening (2, 6, 8, 9), and only one lies on the predicted shortening line (10), which is located south of the plateau (modified from Kley and Monaldi, 1998). Numbers refer to cross-section lines in Figure 1.

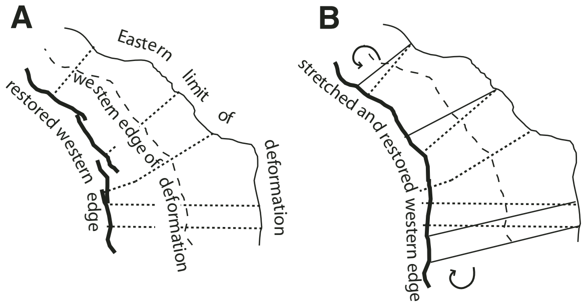


Figure 9. Problems in accommodating shortening along the Bolivian orocline. In both images, the thin line on the right is the eastern limit of deformation in the Central Andes. Curved dashed line is western edge of deformation, and westernmost line is the western edge of deformation, once shortening has been restored. Dotted lines are balanced cross-section locations. (A) Restoring the western edge of deformation in the Central Andes westward highlights the amount of along-strike overlap (thick black lines). (B) Overlaps in the restored western edge (A) have been removed, resulting in convergent flow lines. Solid cross-section lines connect same points in B as in A. To maintain length of arc, in initial and final states, motion of material into core and clockwise rotation in the south and counterclockwise rotation in the north are required; these rotations are in addition to rotations due to differential shortening.

requires the magnitude of rotation to increase toward the forearc. As emphasized by previous studies, shortening and bending an orocline forces upper-crustal material into the core of the orogen (Kley, 1999; Hindle et al., 2002, 2005). However, increased crustal thickness in the center of the Bolivian orocline may induce north- and south-directed crustal flow from the axis of the orocline to the limbs (McQuarrie, 2002b; Yang and Liu, 2003; Ouimet and Cook, 2006). This gravitationally induced flow of weak mid- to lower crust is in the opposite sense of that predicted by kinematic models of the upper crust. Even though the general motion of upper-crustal material is toward the axis of the bend, this study highlights the deficit of upper-crustal shortening in Peru, suggesting middle- to lower-crustal lateral flow from the axis to the limbs.

CONCLUSIONS

The preferred estimate of shortening from the volcanic arc to foreland basin in southern Peru is 123 km. This estimate is a result of rigorously testing the variables (the way space is filled, the dip of décollement, and displacement along faults) that have the largest effect on shortening amounts. Although we present both the minimum and maximum shortening estimates, we argue that our preferred estimate is the most kinematically viable. This amount of shortening is sufficient to account for the area of the cross section itself, assuming isostatic equilibrium. However, it is not enough shortening to account for the crustal area of the entire plateau. Three possible solutions include: (1) a 40% shortening amount maintained across the entire plateau (including the Western Cordillera), (2) an original crustal thickness of 40–45 km, and/or (3) addition of crustal material, by middle- to lower-crustal material flowing from the axis of the Bolivian orocline to the limbs or magmatic addition.

Differential shortening between the hinge and northern margin of the Bolivian orocline can account for the young rotations, younger than 15 Ma, observed in the Altiplano and Subandean zones but cannot account for the total rotation. This suggests that forearc rotations reflect both bending due to differential shortening, ~12°, as well as block rotations (~25°–30°), which facilitated deformation in a tightening core and transported material toward the center of the orocline.

APPENDIX

Variability in Shortening Estimate and Erosion Calculations

Basement Thrusts

We calculated the variations in shortening due to changes in the amount of area filled by basement thrust sheets versus duplication of the sedimentary section by thrusting. To obtain the amount of shortening by maximizing the area filled by basement thrust sheets, we took our restored, undeformed preferred cross section and measured the length of the section representing the doubled strata (see Fig. 3e). The length of this section is 35 km, so we subtracted 35 km from our undeformed section length for an initial undeformed length of 273 km (308 km – 35 km = 273 km). We then took 273 km and subtracted the deformed length of 185 km to get 88 km (273 km – 185 km = 88 km). We took the shortening estimate of 88 km and divided by the undeformed length of 273 km to get 32.2% shortening, or approximately 30%. To obtain the amount of shortening if the section currently filled with basement thrusts were instead filled with sedimentary strata, we measured the length of the basement thrust sheet in the undeformed section of our preferred shortening estimate (120 km). We then added 120 km to our undeformed length in our preferred shortening estimate to obtain an undeformed section of length 428 km (308 km + 120 km = 428 km). We took the difference between the deformed length and undeformed length to get a shortening estimate of 243 km or ~240 km (428 km – 185 km = 243 km), and then we calculated the percent shortening: 243 km/428 km = 56.8% shortening, or ~55%.

Dip of Décollement

In order to obtain shortening estimate variations based on changing the dip of the décollement, we took the length of our deformed cross section of 185 km and calculated the increase and decrease in area associated with adding 3° and subtracting 1° to the décollement dip:

$$185 \text{ km} \times \text{Tan}(3) = \\ 9.7 \text{ km } (9.7 \text{ km} \times^* 185 \text{ km})/2 = \\ 897 \text{ km}^2 = \sim 900 \text{ km}^2, \text{ and}$$

$$185 \text{ km} \times \text{Tan}(1) = \\ 3.2 \text{ km } (3.2 \text{ km} \times 185 \text{ km})/2 = \\ 296 \text{ km}^2 = \sim 300 \text{ km}^2.$$

We then used the average stratigraphic thickness of 10 km to get the undeformed length associated with increasing the dip by 3°:

$$900 \text{ km}^2/10 \text{ km} = 90 \text{ km of additional shortening,}$$

and decreasing the décollement dip by 1°,

$$300 \text{ km}^2/10 \text{ km} = 30 \text{ km less shortening.}$$

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