Plateau-style accumulation of deformation: Southern Altiplano

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[1] Employing surface mapping of syntectonic sediments, interpretation of industry reflection-seismic profiles, gravity data, and isotopic age dating, we reconstruct the tectonic evolution of the southern Altiplano (~20–22°S) between the cordilleras defining its margins. The southern Altiplano crust was deformed between the late Oligocene and the late Miocene with two main shortening stages in the Oligocene (33–27 Ma) and middle/late Miocene (19–8 Ma) that succeeded Eocene onset of shortening at the protoplateau margins. Shortening rates in the southern Altiplano ranged between 1 and 4.7 mm/yr with maximum rates in the late Miocene. Summing rates for the southern Altiplano and the Eastern Cordillera, we observe an increase from Eocene times to the late Oligocene to some 8 mm/yr, followed by fluctuation around this value during the Miocene prior to shutoff of deformation at 7–8 Ma and transfer of active shortening to the sub-Andean fold and thrust belt. Shorthening inverted early Tertiary graben and half graben systems and was partitioned in three fault systems in the western, central, and eastern Altiplano, respectively. The east vergent fault systems of the western and central Altiplano were synchronously active with the west vergent Altiplano west flank fault system. From these data and from section balancing, we infer a kinematically linked western Altiplano thrust belt that accumulated a minimum of 65 km shortening. Evolution of this belt contrasts with the Eastern Cordillera, which reached peak shortening rates (8 mm/yr) in between the above two stages. Hence local shortening rates fluctuated across the plateau superimposed on a general trend of increasing bulk rate with no trend of lateral propagation. This observation is repeated at the shorter length and time scales of individual growth structures that show evidence for periods of enhanced local rates at a timescale of 1–3 Myr. We interpret this irregular pattern of deformation to reflect a plateau-style of shortening related to a self-organized state of a weak crust in the central South American back arc with a fault network that fluctuated around the critical state of mechanical failure. Tuning of this state may have occurred by changes in plate kinematics, during the Paleogene, initially reactivating crustal weak zones and by thermal weakening of the crust with active magmatism mainly in the Neogene stage.


1. Introduction

[2] Although considerable advance has been made in recent years in understanding the processes involved in the formation of orogenic plateaus, the precise temporal and spatial patterns of uplift and lateral progradation of deformation in plateaus beyond the last few million years is still shrouded in darkness. This situation is largely due to a lack of systematic data with the potential to date the timing of deformation and uplift, as well as of detailed structural data that reveal the internal architecture of the world’s large orogenic plateaus, Tibet and Altiplano [e.g., Dewey et al., 1988; Isacks, 1988; England and Molnar, 1997; Allmendinger et al., 1997; Shen et al., 2001]. Young, syntectonic and posttectonic sediments largely cover both plateaus and restrict access to the subsurface. Yet, models of large-scale orogeny and plateau formation heavily rely on assumptions that relate to the early stages of plateau evolution and the conditions governing these. This is particularly true for the Andean Altiplano, which, due to its setting in an ocean-continent convergence system, has been called a plate tectonic paradox [e.g., Allmendinger et al., 1997]. Because of the unsatisfactory data situation, published ideas on accumulation of deformation in the Altiplano range from various inferred stages in plateau evolution [e.g., James, 1971; Pardo-Casas and Molnar, 1987; Isacks, 1988; Sempere et al., 1990; Allmendinger and Gubbels, 1996; Kley, 1996; Allmendinger et al., 1997; Lamb and Hoke, 1997; Lamb et al., 1997; Lamb, 2000; Hindle and Kley, 2002] to suggestions of continuous eastward progradation of deformation, starting at the Western Cordillera, with a propagation style much like that observed in classic foreland fold and thrust belts [e.g., Horton et al., 2001; McQuarrie, 2002]. To date, the available data in the Altiplano do not provide a complete quantitative record of the spatial and temporal pattern of strain accumulation during plateau formation that also sufficiently resolves the earlier stages.

[3] In the present paper we attempt to provide a first nearly complete record of magnitude and timing of shortening across the southern Altiplano plateau (compare Figure 1) by the mapping of syntectonic sediments, by identifying structures, their relative ages and magnitudes of
Figure 1
deformation from the interpretation of industry reflection-seismic profiles, and by isotopic age dating of deformation, including a reevaluation of published ages. Our analysis focuses on the southern Altiplano at 20–22°S. This segment is characterized by the highest density of geophysical and geological data in the central Andes, all provided by a variety of recent projects [e.g., Wigger et al., 1994; Beck et al., 1996; Swenson et al., 2000; Yuan et al., 2000; Haberland and Rietbrock, 2001; Brasse et al., 2002; Götte and Krause, 2002; ANCORP Working Group, 2003]. In addition, high-resolution industry reflection seismic sections covering the entire plateau were made available to this project by the Bolivian national oil industry (courtesy of YPFB).

2. Geological Framework

[4] In spite of ongoing subduction for more than 200 Myr, large-scale deformation of the upper plate and formation of the Andes high plateau only commenced during early to mid-Tertiary times [Sempere et al., 1990; Lamb and Hoke, 1997; Allmendinger et al., 1997; Horton et al., 2001]. The resulting central Andean mountain belt is built from a series of morphotectonic units (Figure 1d): the Longitudinal Valley in the forearc with an average elevation of some 1000 m and thin sedimentary infill; the 4500–5500 m high Chilean Precordillera with Paleozoic and Mesozoic basement, overlain by a thin Neogene cover; the Western Cordillera built by the Neogene to recent magmatic arc with peak elevations of up to 6000 m; the flat Altiplano Plateau at some 3800 m average elevation forming a 200-km-wide intramontane basin; the Eastern Cordillera reaching 5000–6000 m altitude, a doubly vergent thick-skinned thrust system; the inter-Andean and sub-Andean belts accumulated most shortening since the Miocene [e.g., Baby et al., 1997; Kley and Mondali, 1998] and still shows active growth at its eastward deformation front as evidenced from active seismicity and GPS data [Lamb, 2000; Hindle and Kley, 2002; Bevis et al., 1999].

[5] In contrast to the plateau flanks, the southern Altiplano evolution is only recorded by few structural and age data [e.g., Baby et al., 1997; Rochat et al., 1999; Welsink et al., 1995] (see Figure 1a). Most of the area is covered by Quaternary fluvial and lacustrine sediments, including several large salt pans (e.g., Salar de Uyuni, Figure 1), and by posttectonic volcanic rocks from the active magmatic arc of the Western Cordillera. Most of the exposed pre-Cenozoic basement is made up of mainly unmetamorphic and mildly folded Ordovician to Devonian clastic sediments while the western margin of the plateau is built of deformed Late Paleozoic and Mesozoic igneous and sedimentary rocks. In southern Bolivia, Cretaceous sediments unconformably overlie this basement. These sediments were partly deposited in a shallow marine environment [Fiedler et al., 2003] and mark the last marine ingression prior to plateau formation. This postrift sequence is related to the Potosi basin [Fiedler et al., 2003], a branch of a major rift system in the south (“Salta Rift”), the site of which largely matches the extent of the later Eastern Cordillera.

[7] The establishment of internal drainage early in the Andean evolution with two cordilleras confining the proto-plateau area was responsible for storing sediment that reflects the entire deformation history (compare Vilque well [Welsink et al., 1995]; see Figure 1a). In the southern Altiplano, the late Paleocene Santa Lucia formation and the Eocene/Oligocene Potoco formation [Jordan and Alonso, 1987; Sempere et al., 1997; Horton et al., 2001], a series of coarse to fine-grained clastics and red beds conformably overlie the Cretaceous/Paleocene deposits (Figure 2). Rochat et al. [1999] and Jordan et al. [2001] suggested that these sediments were partly deposited in local extensional basins. This sequence is in turn overlain by the Miocene continental clastic San Vicente Formation [Martínez et al., 1994; Mertmann et al., 2003]. To the east this formation is delimited by the San Vicente Thrust zone (Figure 1a) marking the western topographic limit of the Eastern Cordillera. The San Vicente Formation is characterized by significant lateral variations in thickness, grain size, and detrital composition as well as by unconformable
onlaps on the Paleozoic basement. Isotopic ages from intercalated volcanic rocks of the southern Altiplano range between some 27.4 and 5 Ma, interpreted to bracket the timing of deposition of the San Vicente Formation and the supposedly contemporaneous deformation [Grant et al., 1979; Sempere et al., 1990; Soler et al., 1993; Lamb and Hoke, 1997; Horton et al., 2002]. Late Miocene to recent volcano-sedimentary deposits unconformably overlie all preexisting units. The above sequences, summarized in Figure 2, show significant similarities in depositional style and age with those identified by Lamb and Hoke [1997] farther north in the Altiplano.

3. Plateau Architecture

3.1. Structural Style

[8] In outcrop, major folds often associated with thrust faults characterize the southern Altiplano structural style. The folds have subvertical axial planes, a parallel geometry, and are nearly cylindrical with NNE-SSW oriented fold axes plunging less than 5° north or south (see Figure 1a). Bedding surfaces on the fold limbs have well developed slickensides usually oriented downdip (flexural-slip folds). Sediments at the limbs of folds often exhibit growth structures. Apart from folding, most of the shortening is recorded by thrusts rooting below the anticlines. Strike-slip faults are of very minor importance. Major extensional structures are not observed on the southern Altiplano, except for rare small-scale faults, which usually show displacements of no more than a few decimeters to meters.

[9] At surface, the moderately eastward dipping thrust system of the San Vicente Fault Zone (SVFZ) separates the Eastern Cordillera in the east from the sediment-covered Altiplano in the west (Figure 1a). In the plateau area to the west, we subdivide the southern Altiplano into three tectono-stratigraphic domains based on its internal architecture: the eastern, the central, and the western southern Altiplano domain. These are also particularly well imaged by the high-resolution Bouguer map (Figure 1b). In the areas exposing deformed Tertiary rocks positive Bouguer anomalies perfectly trace anticlines and fault hanging walls. Hence the Bouguer map can be seen to highlight the structural trend of folds and faults beneath the younger cover.

[10] The eastern domain, which also includes the Tertiary/Quaternary Lípez Basin, is the buried tip of the Eastern Cordillera's west vergent, thin-skinned western margin with widely spaced open anticlines. The central Altiplano domain is formed by a thrust system separated from the Eastern Cordillera by the undeformed Lípez basin: The Khennayani-Uyuni Fault Zone (KUFZ) represents its key part and is the most important structural element of the southern Altiplano [Sempere et al., 1990; Martinez et al., 1994; Welsink et al., 1995; Rochat et al., 1999]. It represents an east vergent leading imbricate fan of up to four major, N-S to NNE-SSW trending, emergent thrusts (dips are 20–30° at frontal thrust to subvertical at westernmost thrust). Along the thrusts, Paleozoic to Oligocene sediments were displaced eastward over the Cenozoic fill of the Lípez Basin (Figure 3). The hanging walls show folded and faulted Paleozoic sedimentary basement with remnant Late Mesozoic, and a thinned to incomplete Cenozoic cover (Figure 3a). The footwalls of the individual thrusts are characterized by thrust front synclines containing growth structures with steep western limbs and up to 2 km long, gently dipping eastern limbs. Because thrust hanging wall cutoffs are eroded, we assess displacement from down-plunge projection of cutoff geometries mapped in the southern parts of the area (Figure 1a). The western parts of the central Altiplano system is built from west vergent
fault propagation folds with 6–10 km wavelength overlying blind thrusts. The folds have associated small thrust top basins filled with wedge-shaped, onlapping deposits. This entire central Altiplano system is continued to the north in the exposed Tambo-Tambillo area (Figure 1a) [cf. Lamb and Hoke, 1997].

[11] The largely sediment-covered western Altiplano domain only exposes deformed rocks at the Yazo´n Peninsula in the southern Salar de Uyuni (~21°S, 68°W), where folded strata of the San Vicente Formation crop out (Figure 4). The structure of the Yazo´n Peninsula is a west vergent, NNE-SSW trending fold with a long eastern back limb, a steep west facing forelimb and an associated thrust front basin filled with growth strata. In spite of the isolated nature of this exposure, the Bouguer map (Figure 1d) shows that the Yazo´n structure is part of a major NNE-SSW trending system of folds and/or faulted and uplifted rocks under the volcanic and Salar cover (Figure 4). This nearly unknown structural unit in the western Altiplano significantly contributes to its deformation architecture and, from

3.2. Reflection Seismic Data

[12] Surface observations are complemented by industry reflection data that image the crust down to some 4 s two-way time (TWT) depth. The definition of seismic facies and their correlation with mappable stratigraphic formations and well data mainly is shown in Figure 9 in the auxiliary material1.

[13] Seismic lines 10007 and 10010 (Foldout 1) illustrate the lateral (north-south) variability of the Lı ´pez Basin and deformation front of the Eastern Cordillera. Toward depth (line 10007 in Foldout 1), the Potoco Formation shows a westward thinning wedge with conformable top lap reflector geometries at both the upper and lower sequence boundaries. The overlying San Vicente Formation is built from three subsequences, defined here as the OMsv1, OMsv2 (Oligo-Miocene San Vicente Formation), and Pilkhaua Subsequence. The OMsv1 and OMsv2 Subsequences are characterized by relative uniform thickness across the entire section. The Pilkhaua Subsequence is absent in line 10007, most probably due to erosion (compare, however, its occurrence on the west flank of Vilque Anticline in line 10010). In the syncline east of the Vilque Well small normal faults are local features, related to the collapse of the forelimb of the eastern Anticline above evaporite layers within the OMsv2 Subsequence (600 m of evaporites drilled in the Vilque Well). The most prominent unconformity of line 10007 is the truncation at the base of the OMsv1 Subsequence near the deformation front (DF, see Foldout 1e, 1f) above eroded top lap terminations of the Potoco and El Molino Formations. A second structural disconformity in the western part of the Vilque Anticline between the OMsv1 and OMsv2 Subsequences (Foldout 1e) is related to the blind tip of the hanging wall ramp of the Vilque fault bend fold.

[14] At its eastern border, the southern part of the Lı ´pez basin (line 10010, Foldout 1f) exhibits thinning and progressive onlaps of reflectors of the Potoco Formation and OMsv1 Subsequence against a basement high. As in line 10007, the unconformity at the base of the OMsv1 Subsequence overlies well-developed top lap terminations from the Potoco Formation With the onset of deposition of the OMsv2 Subsequence the downlapping reflectors in the western part of the basin indicate a westward prograding system of foresets. Locally developed top lap terminations below the eastern OMsv2 Subsequence show that this sequence boundary was also partly erosive (Foldout 1e, 1f).

[15] The most obvious difference between the northern section (10007) and the southern section (10010, compare Foldout 1) is the smaller basin width in the southern section. A second difference between lines 10007 and 10010 is a

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southward decrease of thrust displacement related to formation of the Vilque Anticline, a classical fault-bend anticline, in the north. The most significant observation of Foldout 1, however, is the strong similarity between both sections in terms of the characteristics of seismic facies, similar positions of unconformities, and the structural style allowing correlation between lines.

[16] In the central Altiplano domain (lines 10023 and 2593, see Foldout 1) surface outcrops at sites of poor quality of the seismic image correlate with steeply inclined strata or with footwall units buried beneath deformed imbricates. The strongly folded and imbricated sequences of the central Altiplano can be observed in nearly continuous outcrop 5–10 km south of the seismic section (Figure 1a). Line 10023 shows an east vergent imbricate fan building the eastern part of the system; west vergent fault bend folds dominate the western part. The analysis of seismic facies and their correlation with surface geology exhibits the same, albeit
Foldout 1. Cross section of the southern Altiplano based on seismic lines 2553 (stacked section), 10023, 2593, and 10007 (migrated sections), courtesy of YPFB (see Figure 1 for location) and their time-depth conversion using average interval velocities of the lines as shown in inset (bold line, average; thin lines, variation along line). Western Altiplano profile is projected into the common section. Structures are shown by name or numbers (eastern Altiplano) with nearby isotopic ages projected along strike. Restored section shows late Oligocene stage prior to onset of sedimentation of San Vicente Formation. Details show (a) western deformation front of western Altiplano thrust system; (b) wedge-shaped westward thinning of Potoco Formation and youngest fold of central Altiplano thrust system folding the onlapping Pilkhaua Subsequence; (c) fault-propagation folds of San Vicente Formation with piggyback basins filled with Pilkhaua Subsequence that onlap neighboring folds; (d) thrust units bounded by Corregidores (right) and San Cristóbal Faults (left) with Potoco beds truncated by San Vicente Formation and growth strata of the Pilkhaua Subsequence; (e) buried western deformation front of the Eastern Cordilleran Thrust System truncated by the base of San Vicente Formation (OMsv1); (f) western deformation front of the Eastern Cordilleran Thrust System along the southern line 10010 (see Figure 1 for location) with truncation of older structures by the basal San Vicente Formation. The latter shows internal truncations in the east (between OMsv1 and OMsv2), westward prograding sedimentary foresets in the west (OMsv2), and growth structures at the top of the sequence (Pilkhaua Subsequence). Numbers indicate sequence of fault activation. (See enlarged version of this figure in the HTML.)
more deformed and dissected, seismic sequences as detected in the eastern Altiplano (Foldout 1 and Figure 9 in the auxiliary material). This includes the prominent double reflector of the El Molino Formation, the seismically transparent facies of the Potoco Formation, and the continuous high-amplitude reflectors of the San Vicente Formation. In the central Altiplano, subdivision of the San Vicente Formation into the OMsv1 and OMsv2 Subsequences is not possible any more. The overlying Pilkhaua Subsequence is most distinct as infill of local thrust top basins (e.g., Foldout 1c and Figure 5).

[17] The Potoco Formation forms a continuously eastward thickening wedge that reaches its maximum thickness in the hanging wall of the San Cristóbal Fault and again exhibits conformable sequence boundaries. Its thickness is significantly smaller east of the San Cristóbal Fault. From seismic analysis, map interpretation, and field observations, the lower sequence boundary of the overlying San Vicente Formation evolves from conformable in the footwall of the individual faults to erosive in their hanging wall positions (top lap terminations of the Potoco reflectors, see arrow in Foldout 1d; see also Figure 3c).

[18] The main body of the San Vicente Formation has a more constant thickness, but also thins westward (line 10023 in Foldout 1) and is very thin on the Khenayani-Uyuni Fault Zone as compared to the immediately adjacent areas. Contacts with underlying strata are conformable in most cases except for the above erosional unconformity at in hanging wall positions (Corregidores Fault, Khenayani-Uyuni Fault, and Allka Orkho Fault). In the uppermost San Vicente Formation, thrust top basins filled with onlapping deposits show wedge-shaped sedimentary bodies defining the Pilkhaua Subsequence. These Pilkhaua-filled basins generally do not extend below 0.5–0.8 s TWT depth. West
of the westernmost anticline, the seismic profile was acquired over undeformed lacustrine sediments south of the Salar de Uyuni Subsequence at flanks of major folds (Foldout 1b).

In the western Altiplano, the structural architecture was only accessible at the Yazón peninsula and in a seismic stack section of poor quality crossing the Salar de Uyuni (example and tentative interpretation of the line drawing of line 2553 is shown in Foldout 1a). Interestingly, structural analysis exhibits a doubly vergent imbricate fan and fault bend fold system quite similar to the central Altiplano structure. In addition, this unit shows a similar style of Bouguer anomalies as the central Altiplano (Figure 1d). Judging from reflection styles, there is restricted evidence for the presence of the El Molino and Potoco Formations (see Figure 9 in the auxiliary material) or they are indistinguishable from the San Vicente Formation. In contrast, the Pilkhaua Subsequence is clearly identified by its reflector architecture. Both San Vicente Subsequences are also observed at surface (Yazón Peninsula) and show the same onlapping relationship as in the central and eastern Altiplano domains (Figure 4). All units are covered and onlapped by younger sediments and volcanics of the Salar de Uyuni basin.

### Section Balancing

Using surface data and the depth-converted interpreted line drawing (depth conversion based on averaged interval velocities of the seismic lines, see inset in Foldout 1), we iteratively constructed and restored a balanced section with the software packages GEOSEC® (Paradigm) and 2D-MOVE® (Midland Valley Exploration Ltd.) using flexural flow as the dominant deformation mechanism as indicated from field observations. We determined fault geometry at depth from seismic interpretation and depth-to-detachment calculations using the fold geometries of various fault-related folds [cf. Epard and Groshong, 1993].

The eastern Altiplano domain is shown to be part of a west vergent thin-skinned fold-and-thrust belt with widely spaced ramp anticlines and open fault-bend folds (Foldout 1, line 10007). Here, most shortening was accumulated prior to deposition of the San Vicente Formation in the western part of the section (4.5 km shortening). All thrusts join a gently eastward dipping detachment, which is characterized by a double flat ramp geometry, descending eastward in two steps from six to nine kilometers depth within lower Paleozoic sediments and probably linking with the San Vicente fault. Minor reactivation (~1.5 km shortening) during a later stage (post-OMsv2 deposition) was responsible for the formation of two wide ramp anticlines east of the earlier deformation front. This interpretation is in contrast to the extensional model of Welsink et al. [1995] but is similar to that of Baby et al. [1997] and Müller et al. [2002] showing, however, significantly more detail.

In the central Altiplano domain, the Khenayani-Uyuni Fault Zone (KUFZ) forming the eastern part of this doubly vergent thrust system is built from four westward dipping thrusts (KUF, CF, SCF, IPF, see Foldout 1) with steep to moderately dipping hanging wall strata underlain by a gently west dipping detachment (Foldout 1, line 10023). In comparison to earlier interpretations [e.g., Baby et al., 1997; Lamb and Hoke, 1997; Rochat et al., 1999; McQuarrie and DeCelles, 2001], this geometry, as deduced from surface as well as from seismic data (compare Foldout 1), accommodates significantly more shortening. Large parts of the hanging walls were eroded during thrusting. Eroded hanging wall cutoffs were therefore reconstructed from downplunge projection of the map. The less deformed, west vergent part of the central Altiplano Fault System exhibits four fault propagation folds above eastward dipping, mostly blind, thrusts (Foldout 1, line 10023). The folds have associated syntectonic basins filled with sediments of the Pilkhaua Subsequence. Initial shortening prior to deposition of the San Vicente Formation, as evidenced from its erosive basis, only affected the external parts of the doubly vergent thrust system, namely the Khenayani-Uyuni and the Corregidores imbricates in the east and the Allka Orkho Fault in the west. The eroded hanging wall cutoffs were reconstructed from extrapolating stratigraphic thickness from the footwall to the hanging wall. Where thickness changes are evidenced, like the reduced Potoco Formation thickness in the KUFZ area, we used erosion estimates derived from fission track modeling. Ege [2004] calculated some 0.9–2.4 km of erosion for the Potoco Formation prior to the onset of San Vicente deposition in this area. Restoration of the central Altiplano revealed a total of 40 km horizontal contraction (13 km of which were accumulated in the early stage) with a predominance of east directed thrusting (see Foldout 1).

The main body of the San Vicente Formation below the Pilkhaua Subsequence exhibits a uniform thickness unaffected by the west vergent structures. In contrast, in the dominantly east verging thrusts, the same formation is very thin or absent, or represented only by Pilkhaua-style deposits (compare Foldout 1, line 10023). This fact requires either continuous or repeated activity of the KUFZ, in order to maintain local relief, or very high relief formed during the initial stage that was not downgraded before the late stage of deposition of the San Vicente Formation. In both cases, local relief with respect to neighboring domains persisted until the sedimentary surface in these reached the surface elevation of the eroded KUFZ area and then started to transgress the latter. In summary, deformation in the central Altiplano locally started prior to deposition of the San Vicente Formation and continued throughout San Vicente deposition in the KUFZ area, while all of the central Altiplano structures were active during deposition of the Pilkhaua Subsequence.

For the western Altiplano thrust system, we only aimed at a semiquantitative model interpreting mainly the seismic section based on reflection styles and geometries. The doubly vergent style of deformation, again with a predominance of east directed thrusting, is similar to that of the central Altiplano system. However, for lack of resolution of structures, the measured shortening of 21 km (see Foldout 1) is a minimum estimate only. A separation of
shortening into two stages is not detectable due to the poor quality of the seismic line and lack of exposure.

[25] Balancing shows that all thrusts in the central and western Altiplano thrust systems merge into a basal detachment, which lies at depths between 9 and 12 km in the Paleozoic sediments (compare eastern Altiplano) and dips 1°–2° westward (Foldout 1). Apart from the dominant east directed thrusting, indications for this detachment geometry were derived from the following two observations:

[26] 1. The Late Cretaceous El Molino Formation forms a subhorizontal regional near sea level in the east rising to ~1500 m elevation in the western Altiplano. Exceptions to this are the strongly deformed external parts of the central Altiplano doubly vergent thrust system, i.e., the Khenuyani-Uyuni Fault Zone and the Allka Orko Fault. There, the El Molino Formation was uplifted along faults and locally crops out at an altitude of ~4000 m (Foldout 1).

[27] 2. The floor of the basins filled with Pilkaua Subsequence has a constant elevation of some 3200 m west of the Khenuyani-Uyuni Fault dropping to 2100 m east of the fault. Because the Pilkaua Subsequence has a nearly identical age throughout the southern Altiplano (see below and Figure 2) and lacks lateral facies changes, this indicates (1) joint uplift of the latter basins in the west and (2) a planar decollement geometry below this part of the plateau without major variation of dip angle that would have triggered formation of ramp anticlines and associated differential uplift of the above units.

[28] The restored section (see Foldout 1) reveals that shortening was preceded by an extensional stage that led to the formation of wedge-shaped half graben fills of the Eocene/Early Oligocene Potoco Formation [cf. Jordan et al., 2001]. The eastern border of the central half graben is formed by a normal fault that was inverted as the San Cristóbal reverse fault. A similar half graben structure may build the Lípez Basin and also appears plausible for explaining the seismic architecture of the western Altiplano. Estimated extension of the proplateau domain was less than 10–15%.

4. Geochronology: The $^{40}$Ar/$^{39}$Ar and K-Ar Age Data

[29] To further decipher the Cenozoic deformation history of the central and western Altiplano, we sampled volcaniclastic intercalations at key locations along the cross section for $^{40}$Ar/$^{39}$Ar and K/Ar isotopic age determinations. Linking age data with structural constraints and stratigraphy aimed at providing time brackets for specific deformation increments. Samples have been selected to cover the entire sedimentation and deformation history. Sample material mostly consisted of felsic, rhyolitic tuffs with abundant indications for textural and physico-chemical disequilibria. An outline of data acquisition procedures and a detailed evaluation of analytical data are given in the auxiliary material.

[30] The sedimentation history of the San Vicente Formation is surveyed with five crystal tuff samples. Samples KE06-99 and KE10-99 were taken from the eastern flank of the Vilque Anticline in the eastern Altiplano, sample KE19-99 is from the eastern flank of the Ines Anticline in the central Altiplano, and samples KE11-99 and KE12-99 come from the northern part of the Yazo´n Peninsula in the western Altiplano (Figure 1).

[31] The history of three syntectonic basins of the Pilkaua Subsequence is constrained as follows: For the basin west of the Yazo´n Anticline (western Altiplano), a late pre-tectonic tephra sample is from the youngest predeformational strata of the San Vicente Formation, from the vertically inclined forelimb of the Yazo´n Anticline (KE28-00). KE24-00 was taken only 50 m farther west, in the basin part of the steeply inclined (48°) syntectonic Pilkaua sediments, ~5 m above the basal conglomerate (Figure 4). The samples KE01-00 and KE30-00 were interbedded in the syntectonic basin west of the Ines Anticline, central Altiplano (Figure 5). KE39-00, KE40-00, and KE41-00 were sampled from the syntectonic basin west of the central Anticline (Foldout 1). The crystal tuff KE40-00 is only slightly inclined, while KE39-00 lies horizontally. KE41-00 is a volcanic boulder in a volcaniclastic layer within the steeper parts of the basin. The latter two tuffs were deposited during or after the latest stages of thrusting. Sample KE13-00 is a post-tectonic ignimbrite of the southern central Altiplano (southwest of Julaca, Figure 1 and Foldout 1).

[32] The $^{40}$Ar/$^{39}$Ar analytical data are summarized in Table 1. An important observation is the presence of variable amounts of excess Ar in almost all samples, both in hornblende and biotite. We therefore evaluated the data using both the apparent age spectra and the information from Ar-Ar isochrons [cf. Miller et al., 1998]. In some samples, excess Ar is restricted to the first, low-temperature degassing steps, and high-temperature degassing steps constitute plateau ages, which we regard as valid extrusion ages for the respective tuffs (KE13-00 biotite, KE24-00 biotite, KE40-00 biotite, KE28-00 amphibole, KE39-00 amphibole, and KE41-00 amphibole; Table 2). Samples with persistence of excess Ar throughout the degassing spectra allowed determination of reasonably precise Ar-Ar isochron ages (samples KE01-00 biotite, KE39-00 biotite; Table 2), similarly interpreted as extrusion ages.

[33] The K/Ar analytical data are presented in Table 2. All K/Ar apparent ages for pre-Pilkaua tuffs of the San Vicente Formation are higher than ~18 Ma. The observation of abundant presence of excess Ar in the Ar/Ar data set sets important limitations on the geological significance of the K-Ar ages from similar rocks. The melts that formed the felsic tuffs and ignimbrites were rich in argon with elevated $^{40}$Ar/$^{36}$Ar ratios. Correspondingly, Ar solubility is known to be particularly high in felsic, rhyolitic melts [Kelley, 2002]. It is very likely that the presence of excess Ar is a general phenomenon in phenocrysts from Altiplano tuffs and ignimbrites that are petrologically and genetically similar to the rocks that were analyzed by the Ar/Ar method in the present study. The conventional K/Ar approach depends on the assumption that all Ar trapped during the magmatic stage is of atmospheric composition. The K-Ar dates are thus equivalent to the total gas ages calculated from Ar/Ar data.
Table 1. Sample Details and Summary of $^{40}\text{Ar}/^{39}\text{Ar}$ Data Collected by Incremental Heating

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock Type</th>
<th>Material</th>
<th>Sieve Fraction (mm)</th>
<th>UTM-x, UTM-y</th>
<th>Strike/ Dip</th>
<th>Tectonic Position</th>
<th>Plateau Age, Ma</th>
<th>TGA, Ma</th>
<th>Isochron Age, Ma</th>
<th>Initial $^{40}\text{Ar}/^{39}\text{Ar}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>KE01-00</td>
<td>tephra</td>
<td>biotite</td>
<td>&gt;0.25</td>
<td>665292, 7677470</td>
<td>290/60</td>
<td>syntectonic</td>
<td>29.7 ± 5.3</td>
<td></td>
<td></td>
<td>366.1 ± 20.6</td>
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<tr>
<td>KE13-00</td>
<td>ignimbrite</td>
<td>biotite</td>
<td>≥0.35</td>
<td>644433, 7669689</td>
<td>315/15</td>
<td>posttectonic</td>
<td>9.4 ± 0.3</td>
<td>10.1 ± 0.4</td>
<td>242.7 ± 47.2</td>
<td>CF: 0.9878</td>
</tr>
<tr>
<td>KE24-00</td>
<td>tephra</td>
<td>biotite</td>
<td>0.160–0.355</td>
<td>605882, 7731505</td>
<td>288/48</td>
<td>syntectonic</td>
<td>17.1 ± 0.7</td>
<td>17.1 ± 1.1</td>
<td>330.5 ± 47.8</td>
<td>CF: 0.9990</td>
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<tr>
<td>KE28-00</td>
<td>tephra</td>
<td>hbl</td>
<td>0.2–0.5</td>
<td>605967, 7730893</td>
<td>323/88</td>
<td>pre-tectonic</td>
<td>20.9 ± 2.1</td>
<td>24.0 ± 2.9</td>
<td>447.9 ± 31.8</td>
<td>CF: 0.9999</td>
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<tr>
<td>KE39-00</td>
<td>crystal tuff</td>
<td>biotite</td>
<td>0.355–0.5</td>
<td>675418, 7678968</td>
<td>158/65</td>
<td>horizontal</td>
<td>15.8 ± 1.2</td>
<td>8.5 ± 0.9</td>
<td>400.6 ± 13.2</td>
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<tr>
<td>KE39-00</td>
<td>crystal tuff</td>
<td>hbl</td>
<td>0.355–0.5</td>
<td>675418, 7678968</td>
<td>158/65</td>
<td>horizontal</td>
<td>8.3 ± 1.2</td>
<td>9.2 ± 1.4</td>
<td>307.0 ± 45.5</td>
<td>CF: 0.9991</td>
</tr>
<tr>
<td>KE40-00</td>
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<td>biotite</td>
<td>0.5–1</td>
<td>679815, 7684634</td>
<td>272/15</td>
<td>syntectonic</td>
<td>7.9 ± 0.2</td>
<td>8.1 ± 0.3</td>
<td>279.1 ± 20.9</td>
<td>CF: 0.9997</td>
</tr>
<tr>
<td>KE41-00</td>
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<td>hbl</td>
<td>0.125–0.25</td>
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<td>257/60</td>
<td>syntectonic</td>
<td>29.3 ± 2.0</td>
<td>32.2 ± 2.2</td>
<td>317.6 ± 75.2</td>
<td>CF: 0.9994</td>
</tr>
</tbody>
</table>

Abbreviation is fsp, feldspar; $^{40}\text{Ar}$ refers to in situ radiogenic Ar. Constants used are $\lambda_3 = 4.962 \times 10^{-10}/\text{yr}$, $\lambda_0 + \lambda_0' = 0.581 \times 10^{-10}/\text{yr}$, $^{40}\text{K}/^{40}\text{Ar}_{\text{total}} = 1.193 \times 10^{-4} \text{ g/g}$. Errors given at the 1$\sigma$ level. 

Table 2. Sample Details and Summary of K-Ar Data

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock Type</th>
<th>Material</th>
<th>UTM-x, UTM-y</th>
<th>Strike/ Dip</th>
<th>Tectonic Position</th>
<th>$^{40}\text{Ar}/^{39}\text{Ar}$ ppm</th>
<th>$^{40}\text{Ar}/^{39}\text{Ar}_{\text{total}}$ Average Ar, ppm</th>
<th>K, %</th>
<th>Average K, %</th>
<th>Age, Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>KE06-99</td>
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<td>biotite</td>
<td>750589, 7664545</td>
<td>108/20</td>
<td>pre-tectonic</td>
<td>0.007841, 0.008093</td>
<td>0.302, 0.112</td>
<td>6.242, 6.291</td>
<td>6.266</td>
<td>18.2 ± 0.5</td>
</tr>
<tr>
<td>KE10-99</td>
<td>crystal tuff</td>
<td>biotite</td>
<td>751833, 7668609</td>
<td>069/03</td>
<td>pre-tectonic</td>
<td>0.005046, 0.005200</td>
<td>0.227, 0.222</td>
<td>4.124, 4.137</td>
<td>4.130</td>
<td>17.8 ± 0.5</td>
</tr>
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<td>KE11-99</td>
<td>crystal tuff</td>
<td>biotite</td>
<td>607111, 7757827</td>
<td>270/55</td>
<td>pre-tectonic</td>
<td>0.010510, 0.010850</td>
<td>0.274, 0.298</td>
<td>6.298, 6.065</td>
<td>6.182</td>
<td>23.6 ± 0.6</td>
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<tr>
<td>KE12-99</td>
<td>crystal tuff</td>
<td>fsp</td>
<td>607111, 7757827</td>
<td>275/55</td>
<td>pre-tectonic</td>
<td>0.000626, 0.000550</td>
<td>0.066, 0.077</td>
<td>0.343, 0.353</td>
<td>0.348</td>
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<td>KE19-99</td>
<td>crystal glass</td>
<td>tuff</td>
<td>666540, 7669542</td>
<td>120/55</td>
<td>pre-tectonic</td>
<td>0.005949, 0.005676</td>
<td>0.325, 0.281</td>
<td>3.348, 3.380</td>
<td>3.367</td>
<td>24.7 ± 0.6</td>
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</tbody>
</table>

Abbreviation is fsp, feldspar; $^{40}\text{Ar}$ refers to in situ radiogenic Ar. Constants used are $\lambda_3 = 4.962 \times 10^{-10}/\text{yr}$, $\lambda_0 + \lambda_0' = 0.581 \times 10^{-10}/\text{yr}$, $^{40}\text{K}/^{40}\text{Ar}_{\text{total}} = 1.193 \times 10^{-4} \text{ g/g}$. Errors given at the 1$\sigma$ level.

5. Age and Pattern of Deformation

5.1. Constraints on Deformation Stages

[34] The youngest structures observed are small normal faults that dissect the syntectonic sediments of the Pilkhaua Subsequence in the eastern Altiplano (to the east of the Vilque Anticline, Foldout 1) but are covered by undeformed Quaternary sediments. These faults are local features related to underlying evaporates (see chapter on seismic interpretation) and were only observed in the eastern part of the main cross section (line 10007).

[35] The formation of structures related to deposition of the Pilkhaua Subsequence occurred between ~19 and ~7 Ma. In the western Altiplano, the earliest onset of shortening was found at the Yazón Anticline (youngest pre-tectonic tuff, 20.6 ± 0.1 Ma, Ar-Ar; sample ke24-00, Figure 4). However, interpretation of the seismic section farther north indicates that these samples may be underlain by as much as several 100 m of onlapping Pilkhaua Subsequence that records an earlier onset of deformation. The oldest undeformed andesitic lava covering the western Altiplano thrust system (at northern margin of Salar de Uyuni) was dated at 11.1 ± 0.4 Ma [Baker and Francis, 1978] marking a minimum age for the end of shortening here (compare similar observations in Western Cordillera farther north at 18–19°S by Charrier et al. [2002] and García et al. [2002]).

[36] Syntectonic sedimentation in the doubly vergent thrust system of the central Altiplano began at a similar
time: The oldest syntectonic age (15.8 ± 1.2 Ma, Ar-Ar, ke01-00) was obtained from a 60° WNW dipping tephra in the thrust front basin west of the west vergent Ines Anticline (see Figure 5, note, however, that the sample position is underlain by 200 m of Pilkhaua deposits at depth). In the east vergent part, the oldest K-Ar maximum age was derived from a vertically inclined tuff in the footwall of the San Cristóbal Fault (13.7 ± 0.4 Ma, sample ps99-17 [Silva-González, 2004]; note again underlying Pilkhaua deposits of 400–600 m thickness). The end of deformation is indicated by undeformed dacitic lava, which unconformably overlies steeply inclined San Vicente strata in the hanging wall of the westernmost fault of the KUFZ system, dated at 11.0 ± 0.5 Ma (K-Ar, sr99-10 [Silva-González, 2004]). Subhorizontal to horizontal tuffs in growth sequences yielded 7.9 ± 0.2 and 8.3 ± 1.2 Ma (ke41-00, Ar-Ar, bedding 272/15; ke39-00, Ar-Ar, horizontal bedding) in the central parts of the doubly vergent system, and 11.0 ± 0.3 Ma (K-Ar maximum age, ps99-26 [Silva-González, 2004]) in the footwall of the San Cristóbal Fault. Subhorizontal tuffs sealing the Khenayani-Uyuni Fault (10 ± 0.3; sample SR99-15 [Silva-González, 2004]) indicate that deformation at the KUFZ was shut off between 10 and 11 Ma (compare similar observations on Tambo-Tambillo structure toward the north [Lamb and Hoke, 1997]). The progressively tilted deposits of growth strata in the hanging wall of the KUFZ indicate continuous thrusting since ≥14 Ma (oldest dated Pilkhaua deposit). However, the parallel and folded reflectors of several 100 m of San Vicente deposits underlying this sample may indicate slow or absent deformation earlier. On the other hand, the absence of San Vicente deposits older than 15–18 Ma in the KUFZ structure, their unconformable basis covering all deeper units in the hanging wall anticlines, and the thickness of 3–3.5 km for the San Vicente deposits in the footwall require continuous or repeated fault activation during San Vicente deposition in order to maintain local relief and to provide accommodation space (see above).

[37] In the eastern Altiplano domain, the laterally uniform thickness and overall conformable boundaries of the OMsv1 and OMsv2 Subsequences indicate a late Miocene age of deformation for the Vilque and the eastern Anticlines (see Foldout 1). The youngest OMsv2 age in the eastern Altiplano (16–17 Ma, extrapolated from samples ke10/99, 17.8 ± 0.5 Ma, and ke 06/99, 18.2 ± 0.5, ~200 m below top OMsv2) is a maximum age of formation of these structures. Onlaps and divergent reflectors to the west of the Vilque Anticline (line 10010, see Foldout 1) indicate that its growth was coeval with the deposition of the Pilkhaua Subsequence. The youngest mildly tilted tuff in the Pilkhaua Subsequence north of the Vilque anticline has an age of 7.6 ± 0.2 Ma [Silva-González, 2004].

[38] Except for the observations in the KUFZ area, undisturbed deposition of the San Vicente Formation in the entire southern Altiplano indicates a stage of local tectonic quiescence. In contrast, its more complex internal sequence architecture toward the margin of the Eastern Cordillera (with westward prograding foresets) must have witnessed nearby uplift east of the study area. From the sedimentary record in the San Vicente Formation with material provenance from the east [Horton et al., 2002], deformation in the adjoining parts of the Eastern Cordillera (i.e., San Vicente thrust and hanging wall to the east) must have peaked between 25 and 17 Ma, during deposition of the westward prograding OMsv2. According to Müller et al. [2002], the nearby San Vicente thrust was sealed by volcanics with an age of ~17 Ma marking the end of the main stage of deformation and uplift of the margin of the Eastern Cordillera. This matches the end of deposition of the OMsv2 Subsequence of the San Vicente Formation.

[39] The locally developed truncation of tilted hanging wall sequences below the San Vicente/OMsv1 across major parts of the southern Altiplano is evidence for pre-San Vicente deformation. The almost complete lack of volcanic intercalations in the Potoco Formation, which was deposited prior to this deformation increment, prevents deriving precise age constraints from syntectonic deposits. A single tuff age within intensely folded Potoco sediments in the footwall of the Corregidores Fault yielded a K-Ar maximum age of 40.4 ± 1.1 Ma (sr99-03 [Silva-González, 2004]). The oldest dated volcanic rock of the San Vicente Formation in the central Altiplano gave 29.3 ± 1.9 Ma (volcanic boulder in volcanioclastic deposit; sample ke41-00, amphibole Ar-Ar). Moreover, near sites of early shortening (eastern margin of KUFZ) the basal conglomerate overlying Eocene to Paleozoic sediments contains Silurian clasts from the adjacent eroded hanging walls and is overlain by a basaltic-andesitic tuff with a maximum age of 27.4 ± 0.6 Ma (K-Ar, sample sr99-12 [Silva-González, 2004]). Finally, Ege [2004] reports apatite fission track ages from the uplifted Paleozoic basement rocks of the KUFZ faults, ranging between 28 and 33 Ma. In conclusion, this early shortening stage in the central Altiplano is bracketed between ~35 and 27 Ma.

[40] In the eastern Altiplano domain, the anticlinal hinges formed during this stage were eroded prior to the deposition of the lowermost OMsv1 sediments. Some 700 m above the base of the latter, a K/Ar age of 25.3 ± 1 Ma was determined (sample an14-2000 [Silva-González, 2004]). Considering average sedimentation rates (compare Figure 2b) the base of the San Vicente Formation, and hence the period of this early deformation increment, is constrained to between 28 and 32 Ma at the western margin of the Eastern Cordillera.

5.2. Evolution of Shortening

[41] Using the data on horizontal shortening as derived from section construction and the newly reported as well as published ages from synkinematic deposits, we analyze the evolution of shortening in more detail with the strategy outlined by Victor et al. [2004]: Shortening related to each fault is distributed over the period of its activity as constrained from the sedimentary record in the immediately adjoining syntectonic deposits. The age estimate for the oldest beds recording the onset of deformation and of the youngest deformed are based on the above isotopic data or were estimated from extrapolation using average sedimentation rates (the latter are 100 m/Myr ± 20 m/Myr as derived for the Pilkhaua Subsequence across the Altiplano with
local maximum of ~200 m/Myr in the center of the Lipez basin) and some 150–250 m/Myr or 150–200 m/Myr for the deeper members of the San Vicente Formation in the eastern and central Altiplano domain, respectively; compare Figure 2). To assess the maximum or minimum period of fault activity we included the respective errors in isotopic age dating and the sedimentation rate estimate used to calculate the ages of the base and top of syntectonic deposits. Potential error in determining heave is estimated at ±10% (+20% in less well constrained sections). Using these estimates, we assessed the highest and lowest possible shortening rates for each fault over its activity period. Summing these numbers for all kinematically linked faults provides minimum shortening rates (minimum shortening and maximum activity period) and maximum shortening rates (maximum shortening and minimum activity period) for a thrust system.

[43] Figure 6 shows the summed results merged with equivalent data from the literature for the western plateau flank and the Eastern Cordillera at 21°S. As mentioned above, two main stages of shortening can be identified across the Altiplano up to the western border of the Eastern Cordillera. An earlier Oligocene stage (between 35 and 27 Ma) has averaged local shortening rates between 0.1 and some 3 mm/yr (not resolved in western Altiplano for lack of exposure). This stage is followed by an early Miocene lull in deformation. Shortening resumed with a distinct local acceleration to some 1.7–3 mm/yr (west and center) in the middle to late Miocene (20–10 Ma). In most of the Altiplano, rates were higher during the later stage while the eastern Altiplano shows the opposite relationship. At 7–8 Ma, deformation ceased everywhere. From these figures, bulk Cenozoic strain rates in the southern Altiplano were at an average of \(3.5 \times 10^{-16} \text{ s}^{-1}\) with a middle Miocene maximum of \(2 \times 10^{-14} \text{ s}^{-1}\) (Figure 7c).

5.3. Deformation Partitioning

[43] Inspection of the data at higher temporal and spatial resolution underlines the complex mode of strain accumulation discussed in the following with two examples. In the symmetrical syntectonic basin (Pilkhaau Subsequence) separating the adjacent Ines and Allka Orkho faults (Figure 5), growth sediments record alternating and contemporaneous activity of both faults: stage 1, activity of Allka Orkho Fault and minor uplift of Ines fold (ST1); stage 2, growth of Ines fold, quiescent Allka Orkho Fault (ST2); stage 3, growth of Allka Orkho Fault dominates over Ines fold (ST3); and stage 4, the Ines Fold again takes over. At the scale of the entire central Altiplano unit, we observe Miocene initiation of first the east vergent KUFZ structures, succeeded by formation of new thrusts (14–11 Ma) in all parts of this thrust system. Finally, active deformation between 11 and 8 Ma prograded westward with minor shortening in the internal parts of the system and a quiescent eastern deformation front.

[44] The eastern Altiplano shows a similar evolution for the deformation front of the Eastern Cordillera (section 10010 in Foldout 1f). Shortly before or during initial deposition of OMs1, shortening propagated westward into the Lipez basin and, subsequently, retreated back east to the San Vicente thrust. The latter accommodated subsequent shortening at the Eastern Cordillera front between 27 and 17 Ma (i.e., during OMs1 and OMs2 [Müller et al., 2002]). Following shutoff of the San Vicente Thrust, deformation again propagated westward, forming the Vilque Anticline, until its end at ~7 Ma.

[45] Across the entire southern Altiplano, the Pilkhaau-filled basins exhibit 3 to 5 internal subsequences based on observation of reflector terminations and reflector packages (compare Foldout 1 and Figures 5c and 5d). These packages show a trend toward long high-amplitude reflections at their respective tops. This trend in reflection signal reflects a progressive change in grain size, porosity, and/or average bed thickness. In the Vilque well, Welsink et al. [1995] describe the San Vicente Formation drilled to consist of two major fining upward sequences matching the two seismic sequences in line 10007 that crosses the well site (compare Foldout 1). Mertmann et al. [2003] also describe sequences of several 10 to 500 m thickness showing internal fining upward cycles in the San Vicente Formation. While these sequences are very probably also influenced by climatic fluctuations the nature of the sediments as syntectonic deposits next to tectonically created relief indicates that changes of the balance between relief formation and relief degradation have the most important impact on the local sedimentation.

[46] From the age span recorded in the Pilkhaau Subsequence (~10 Ma, Figure 2), the above number of internal sequences amounts to substages of locally enhanced rates of formation of relief at recurrence intervals of 2 ± 1 Ma for an individual structure. The San Vicente Formation shows a laterally varying number of 4–7 reflector packages of similar properties as mentioned above spanning the time between 29 and 19 Ma (Figure 5a and Foldouts 1c, 1e, and 1f). This time span yields the same recurrence interval as for the Pilkhaau Subsequence. Hence the major deformation periods in the Altiplano identified in Figures 6c and 6d that typically encompass some 5–15 Myr are broken down locally into shortening rate peaks at a typical timescale of 1 to 3 Myr for individual structures. Because of the variable number of these internal sequences between neighboring structures, the structures within a coupled fault system may at best exhibit a partial synchronization. More probably, they record random, mostly unsynchronized distribution of these fluctuations of shortening rate.

6. Discussion

6.1. Deformation Age and Pattern

[47] The new data in this study together with published data complete a cross section of the central Andean Plateau and its margins at 20–22°S, except for a ~40-km-long segment of the Western Cordillera (the Recent magmatic arc), where Miocene to Quaternary volcanic deposits prevent access to the subsurface and no high-resolution reflection seismic data are available. However, observations from
exposed Neogene deposits of the Western Cordillera area farther north (18–19°S [Garcia et al., 2002; Charrier et al., 2002]) show similar age properties as the Precordillera and the western Altiplano at 21°S and may be projected to the unexposed part of the section at 21°S. The structural style of the entire southern Altiplano Plateau is here identified as a series of up to four doubly vergent thrust systems, which vary in size and in amount of shortening (Figure 6). Most of these resulted from inversion of earlier extensional structures. Inversion commenced in the Eocene in the west figure 6. Distribution of deformation ages, shortening, and magmatism across the southern central Andes (21°S) based on published and own data. (a) Compilation of deformation ages: Western Flank [Victor et al., 2004], Precordillera [Haschke and Guenther, 2003], Altiplano [this study; Ege, 2004; Silva-González, 2004], Eastern Cordillera [Gubbels et al., 1993; Müller et al., 2002], inter-Andean [Kley, 1996; Ege, 2004], and sub-Andean [Kley, 1996]. (b) Balanced cross section at 21°S compiled from Victor et al. [2004] for Altiplano West Flank; own data (Altiplano), and Müller et al. [2002] for Eastern Cordillera and sub-Andean; Moho and Andean Low-Velocity Zone (ALVZ) from receiver function data [Yuan et al., 2000]. Line drawing in the middle crust indicates locations of strong reflectivity in ANCORP seismic line [cf. ANCORP Working Group, 2003]. The western Altiplano cross section was projected along strike of gravity anomalies (Figure 1) 50 km southward. (c) Estimates of maximum and minimum periods of active faulting and related folding in the units building the Altiplano (see Foldout 1 for names and location of structures). (d) Cumulative shortening rates of the western, central, and eastern Altiplano thrust systems, calculated from fault activity periods and heave (see text for explanation; result for western Altiplano is extrapolated from observation of ages at Yázon peninsula to all faults observed in the seismic line considering the maximum possible error). Bold line delineates average shortening rates based on sliding average of 3 Myr sampling width. Shortening rate of the Precordillera from Victor et al. [2004] (complemented by Haschke and Guenther [2003]). For the Eastern Cordillera we estimated average shortening rates for individual subunits based on sections and age data summarized by Müller et al. [2002]. Because of the lack of resolution, there are both very few preserved piggyback basins and incomplete age dating of individual structures; the uncertainty here is higher than in the plateau area. See color version of this figure at back of this issue.
Accumulation of shortening in the past 45 Myr (see Figures 6 and 7) initiated at the protoplateau margins, i.e., the Eastern Cordillera and Chilean Precordillera at around 40–45 Ma ago [e.g., Sempere et al., 1997; Lamb and Hoke, 1997; Müller et al., 2002; Haschke and Guenther, 2003] followed by a break in the west. The subsequent evolution shows peak activity to repeatedly shift between the Eastern Cordillera and the more central and western parts of the evolving plateau area. In the final stage at ~11–8 Ma, contraction affected the entire southern Altiplano plateau area dying out everywhere at around 7–8 Ma. Shortening then migrated to the eastern flank of the plateau where eastward migrating deformation led to the formation of the broad, thin-skinned thrust belt of the inter-Andean and sub-Andean Ranges [Kley, 1996].

Synchronization of peak shortening periods in the western systems (Altiplano west flank, western and central Altiplano) and their anticorrelation with those from the Eastern Cordillera, exhibiting repeated advance and retreat of its western deformation front that is unrelated to the remaining Altiplano, indicates kinematic coupling of various thrust systems: One is rooted in the Chilean Precordillera/western Altiplano and the other in the Eastern Cordillera. The existence of the here newly defined western Altiplano Thrust Belt (see Figures 8 and 6b), coupling the three western thrust units, is also corroborated by the here identified gently westward dipping detachment underlying the systems. These findings make the more eastern parts of this belt (western and central Altiplano thrust systems) a broad east facing, low-tapered prowedge accumulating most shortening (~3–4 km shortening [cf. Victor et al., 2004]).

Although the data imply significant synchronicity of deformation stages across parts of the plateau at time frames >5 Myr, more detailed inspection of the seismic sequence architecture of the syntectonic basins reveals significant short-term complexity with repeated retreat of the deformation front (Eastern Cordillera) or repeated alternate thrust activation in the individual doubly vergent systems, and last, but not least, variations in local fault slip rates at local timescales of 1–3 Myr. Hence tectonic deformation rates would be seen to peak at various timescales between the isotopically resolved scales of several million years down to at least the subisotopic geological record exhibiting discon-

Figure 7. Cumulative shortening and shortening rate evolution of the southern Altiplano (sub-Andean plateau flank not included). (a) Summed minimum and maximum shortening rates across the southern Altiplano. Bold line depicts average based on sliding 3 Myr average. (b) Stacked average shortening rates of individual thrust systems building the southern Altiplano (from Figure 6). (c) Cumulative shortening.
here reported details along the section at 21°S may well be representative for most of the Altiplano to the north (compare to observations by Lamb and Hoke [1997]). Toward the south, the Puna plateau currently exhibits seismic activity across its entire area, bulk GPS-based shortening at some 6 mm/yr distributed over the entire Puna [Klotz et al., 2001], an increase of shortening rate since the Pliocene [Ramos et al., 2002], and the growth of three to four regularly spaced thrust systems across strike that define intervening basins. This situation is highly similar to the late Miocene stage of synchronous plateau-wide shortening in the southern Altiplano prior to eastward expansion with an associated acceleration of rates.

[53] Last, but not least, it is interesting to note that this evolution of shortening rates has a complex relation with the evolution of plate convergence rates between the South American and the oceanic Nazca plates [cf. Pardo-Casas and Molnar, 1987; Somoza, 1998]. Only the early, Eocene to Oligocene, evolution of upper plate shortening may be seen to correlate with two stages of enhanced plate convergence rates at 50 to 44 Ma and 25 to 20 Ma. In contrast, the decrease of Neogene convergence rate from some 15 cm/yr at 20 Ma to the present value of 6.5 cm/yr [Klotz et al., 2001] is anticorrelated with a shortening rate increase [cf. also Hindle and Kley, 2002]. We therefore speculate that plate convergence rate was only a relevant factor controlling upper plate shortening rate in the initial stages.

6.2. Shortening, Crustal Thickening, and Surface Uplift

[54] The distribution of upper crustal shortening concluded from the two thrust systems building the plateau (western Altiplano thrust system, ≥65 km; Eastern Cordillera, ~85 km [cf. Müller et al., 2002]) does not match an equivalent thickness distribution of the deeper crust beneath these (compare Figure 6b). While there is slightly thicker crust underlying the plateau margins, various geophysical data (see Figure 6) [Wigger et al., 1994; Beck et al., 1996; Yuan et al., 2000; ANCORP Working Group, 2003] indicate that the plateau crust everywhere exceeds 60 km thickness. Moreover, plateau surface uplift as recorded by Gregory-Wodzicki [2002] with enhanced rates since the late Miocene is succeeding the main stage of upper crustal shortening in the plateau domain. Hence upper crustal shortening as discussed here is geometrically as well as temporally decoupled from deeper crustal and mantle lithosphere deformation. This is in line with late Neogene to Recent underthrusting of the plateau domain from the east and deep crustal flow as suggested by various authors [i.e., Allmendinger and Gubbels, 1996, and references therein]. From the lack of a significant near surface deformation response to this stage, we conclude a near complete mechanical decoupling of the upper crust from the deeper crust since at least the late Miocene.

6.3. Magmatism and Deformation

[55] The evolution of the Altiplano was paralleled by arc and back arc volcanic activity since the late Oligocene, largely coeval therefore to the period of shortening and largely matching the lateral extent of the plateau. This link

Figure 8. Outline of western Altiplano thrust belt and of the western deformation front of the Eastern Cordillera thrust belt drawn on shaded relief map of Altiplano Plateau. Note narrow intervening basin between both belts (López Basin (LB) in the south and Toledo Basin (TB) in the north) as well as restricted outcrop of prowedge of western Altiplano Thrust Belt in Corque (C), Tambo-Tambillo (TT), and the here analyzed San Cristóbal (SC) and Yazón (Y) areas.
has been suggested to reflect a general genetic relationship between magmatic activity and plateau-forming processes in terms of thermally controlled weakening of the upper plate [e.g., Isacks, 1988; Allmendinger et al., 1997; Kay et al., 1999; Babeyko et al., 2002; ANCORP Working Group, 2003].

[56] In the southern Altiplano, the initiation of volcanic activity (27–29 Ma) slightly postdates the onset of the first shortening increment (35–27 Ma) and the early stage of deposition of the San Vicente Formation (compare Figures 6a and 7a) (compare similar observations in other parts of the plateau [e.g., Kay et al., 1999; Riller et al., 2001; Ramos et al., 2002]). Volcanism continued until 10–8 Ma in the back arc plateau with ongoing magmatism in the present arc [Everden et al., 1977; Fornari et al., 1993; Soler et al., 1993; Kennan et al., 1995], largely overlapping therefore with the main stage of shortening of the entire southern Altiplano domain. Hence it would appear that near surface magmatism in the southern Altiplano locally followed the onset of deformation with a minor time lag, but that it paralleled the general trend of increased shortening rates of the southern Altiplano plateau (compare Figure 7a).

[57] On the basis of these observations, it appears plausible that the thermal evolution of the lithosphere resulting in transient presence of melts may have sufficiently mechanically weakened the upper crust of the Altiplano to enhance global deformation rates through time as observed for the plateau (compare classical model by Isacks [1988]). Modeling results by Babeyko et al. [2002] showed that advective melt transfer through the crust with associated crustal melting requires some previous crustal shortening and thickening, which is equivalent to the correlation observed here.

6.4. Plateau-Style Deformation

[58] Along with the above observation of fluctuating shortening rates, the highly irregular spatial and temporal pattern of strain accumulation at all scales hints at a self-similar distribution of deformation during plateau growth since the Eocene with a lack of characteristic length and timescales in the processes shaping the plateau. Fault slip, or periods of enhanced local strain rates, on a fault network that is in failure equilibrium and has an approximately fractal size distribution (compare observations by Marrett and Allmendinger [1991] in the Puna) would be expected to generate the type of scale-independent response found in the syntectonic sedimentary record of the Altiplano. It is obvious that the number of small tectonic events, at least partly reflected in individual layers deposited in the syntectonic basins, is significantly larger than the number of reflection seismic sequences within each basin. Their number, in turn, is lower than the few larger “displacement events” building the Andes, like the rapid strong shortening of the Eastern Cordillera in the Miocene. These observations suggest that plateau deformation in the Andes is essentially accommodated by a series of interconnected faults, which are close to failure equilibrium helping to distribute strain.

[59] This conclusion and the above observations throw additional light on the dynamics of plateau formation. On the basis of numerical modeling, the mechanical evolution underlying plateau formation (see Wdowinski and Bock [1994], England and Molnar [1997], Pope and Willett [1998], Shen et al. [2001], and Vanderhaeghe et al. [2003] for details) involves heating and related weakening of the lower crust as a result of continuous crustal thickening and melting. However, most of these models invoke an initial domain of localized deformation, which expands laterally forming classical self-similar growing wedges until mechanical decoupling at the base entails surface flattening and the formation of a high plateau with two topographically distinct margins bounding a hot, viscous deeper lithosphere. The resulting plateau lacks a lateral mass gradient forcing lateral migration of deformation. Subsequent shortening within the plateau would be delocalized with an irregular pattern in time and space coincident with the here reported observations. In contrast to the model predictions, however, our results show that plateau initiation in the Altiplano area occurred by growth of two independent thrust systems at the site of the later plateau margins during the early stages. These do not show much evidence for lateral expansion until the sub-Andean ranges developed with concordant shutdown of deformation of the plateau. In the case of the central Andes, first-order anisotropies probably controlled this initial localization of deformation and, hence the subsequent evolution: the hot magmatic arc crust in the kinematic axis of the western Altiplano thrust system, and large-scale mechanical heterogeneity resulting from repeated rifting of the crust at the site of the later Eastern Cordillera.

7. Conclusions

[60] The southern Altiplano structure formed during a complex shortening history that peaked in the early Oligocene (~35–27 Ma) and middle/late Miocene (19–7 Ma). These stages were preceded by an extensional stage that resulted in formation of a series of grabens across the entire Altiplano, which later localized the formation of several thrust systems. The three western systems, the west facing Precordillera/Western Cordillera and the east facing western and central Altiplano systems probably form a coupled retropropyed system that is separate from the eastern Altiplano/Eastern Cordillera system. Both cordilleras delimiting the later plateau area started shortening earlier during the Eocene (40–45 Ma). Subsequently, their deformation history was decoupled with joint activity only in the middle Miocene. Horizontal contraction on the southern Altiplano gradually subsided between 11 and 7 Ma as indicated by the age of undeformed volcanic rocks sealing the contractional structures.

[61] Deformation of the Andean Plateau at 21°S is characterized by a spatially and temporally irregular fault activity pattern, at a generally increasing bulk rate of plateau shortening during the Oligocene/Miocene. Largely synchronized peaks at the 5–15 Myr timescale within a kinematically coupled thrust system were complemented by local
fluctuations of strain accumulation at individual faults at a typical timescale of 1 – 3 Myr that show no evidence of lateral synchronization. These observations preclude systematic propagating deformation. The spatial and temporal pattern of deformation indicates that the plateau probably was self-organized fluctuating about a critical state of failure throughout its evolution. This style of deformation is fundamentally different from smaller-scale orogenetic wedges that exhibit more continuous lateral growth. The Altiplano plateau has probably had low relief since its initiation and evolved from partly independent activity of two thrust belt systems rather than from a single initially doubly vergent orogen. Previously suggested thermal weakening of the plateau crust is reflected in a loose spatial and temporal relationship between deformation and, slightly later onset of, volcanism. Eocene to Oligocene crustal shortening with some relation to plate convergence rates...


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Figure 1
Figure 1. (a) Geological sketch map of the southern Altiplano (section of map of central Andes by Reutter et al. [1994] including our own mapping results) superimposed on shaded relief of the topography showing the distribution of pre-late Neogene rocks, lower-hemisphere equal-area stereonet projections with bedding and regional fold axes, and the location of reflection seismic lines. (b) Bouguer anomaly map (provided by H. J. Götze; see Götze and Krause [2002] for description of gravity field) of area shown in Figure 1a depicting positive anomalies closely related to faults and folds mapped at the surface and observed in the seismic sections. Note the chain of anomalies in the nearly completely sediment-covered western Altiplano with similar style as obvious from the exposed central Altiplano unit. (c) Location of the working area. (d) Major morpho-tectonic units of the southern central Andes.
Figure 6. Distribution of deformation ages, shortening, and magmatism across the southern central Andes (21°S) based on published and own data. (a) Compilation of deformation ages: Western Flank [Victor et al., 2004], Precordillera [Haschke and Guenther, 2003], Altiplano [this study; Ege, 2004; Silva-González, 2004], Eastern Cordillera [Gubbels et al., 1993; Müller et al., 2002], inter-Andean [Kley, 1996; Ege, 2004], and sub-Andean [Kley, 1996]. (b) Balanced cross section at 21°S compiled from Victor et al. [2004] for Altiplano West Flank; own data (Altiplano), and Müller et al. [2002] for Eastern Cordillera and sub-Andean; Moho and Andean Low-Velocity Zone (ALVZ) from receiver function data [Yuan et al., 2000]. Line drawing in the middle crust indicates locations of strong reflectivity in ANCORP seismic line [cf. ANCORP Working Group, 2003]. The western Altiplano cross section was projected along strike of gravity anomalies (Figure 1) 50 km southward. (c) Estimates of maximum and minimum periods of active faulting and related folding in the units building the Altiplano (see Foldout 1 for names and location of structures). (d) Cumulative shortening rates of the western, central, and eastern Altiplano thrust systems, calculated from fault activity periods and heave (see text for explanation; result for western Altiplano is extrapolated from observation of ages at Yázon peninsula to all faults observed in the seismic line considering the maximum possible error). Bold line delineates average shortening rates based on sliding average of 3 Myr sampling width. Shortening rate of the Precordillera from Victor et al. [2004] (complemented by Haschke and Guenther [2003]). For the Eastern Cordillera we estimated average shortening rates for individual subunits based on sections and age data summarized by Müller et al. [2002]. Because of the lack of resolution, there are both very few preserved piggyback basins and incomplete age dating of individual structures; the uncertainty here is higher than in the plateau area.