LATITUINAL VARIATION OF DENUDATION IN THE EVOLUTION OF THE BOLIVIAN ANDES

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ABSTRACT. Latitudinal gradients in topography, relief, climate, and deformation have been used to suggest that climate-driven erosion has exerted a first order control on the development of the central Andes. We synthesize the spatial and temporal variations in denudation across the eastern Bolivian Andes (14-22°S) from new and existing estimates to test whether physical evidence exists to support the hypothesis that erosion influences thrust belt evolution. Basin-morphometry, channel network indices, climate, and longitudinal river profiles indicate a northward increase in relative relief, fluvial incision, and denudation. Short-term denudation-rate averages from landslide mapping and sediment-flux data range from 1 to 9 mm/yr in the north compared to 0.3 to 0.4 mm/yr in the south. Long-term denudation-rate estimates from thermochronology, cosmogenic radionuclides, foreland basin sediment volumes, stratigraphy, paleoerosion surface degradation, and balanced cross sections range from 0.04 to 1.6 mm/yr with rates up to more than twice as fast in the north when comparing estimates from the same method. The shorter-term denudation rates exhibit the greatest variance. Our denudation synthesis shows that an along-strike disparity in denudation (greater in the north) has existed throughout the Holocene and perhaps existed since as early as the late Miocene. Our denudation synthesis also suggests that the disparity and denudation rates have increased to the present. Correlations between the thrust belt geology, geometry, geomorphology, climate, and kinematics of the orogenic wedge provides a case study in observing the regional scale interactions between uplift, climate, and erosion. We conclude that the denudation history, uplift history, and tectonic-geomorphic correlations suggest that models of the evolution of the Bolivian Andes should incorporate a latitudinal erosion gradient for the last 10 kyrs to perhaps 10 Myrs.

INTRODUCTION

The interactions between uplift, climate, and erosion shape mountain belts. The most fundamental relationship is that tectonic uplift builds mountain topography while climate-driven erosion wears it down. Mountain ranges can be idealized to that of a doorstop wedge (Coulomb wedge) which theoretically describes how erosion can influence the distribution of uplift and deformation within the wedge (Dahlen and Suppe, 1988; Dahlen, 1990; Koons, 1994; DeCelles and Mitra, 1995; Hilley and Strecker, 2004; Whipple and Meade, 2004). Furthermore, numerical modeling studies have been used to investigate the dynamics of uplift-climate-erosion feedbacks in orogenic evolution (Beaumont and others, 1992; Willett and others, 1993; Avouac and Burov, 1996; Willett, 1999; Beaumont and others, 2001). These modeling studies provide important inferences about the co-evolution of topography, climate, erosion, and deformation that can be applied to real mountain belts (Hoffman and Grotzinger,

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1993; Reiners and others, 2003; Theide and others, 2004, 2005). However, the complex variability of the controls on denudation processes has limited the application of these models of mountain evolution to real mountain belts (Beaumont and others, 2000; Hovius, 2000). Therefore, quantifying mountain denudation histories remains an essential goal for understanding the role of erosion in orogenesis.

An alternative approach to numerical modeling is to constrain the spatial and temporal variability of denudation rates and identify their forcing factors using physical evidence from the mountain geomorphology and geology. For example, recent databases and digital topography provide new perspectives on morphometric controls on basin denudation at variable spatial scales (Hovius, 1998). Additionally, denudation rates can be quantified at different spatial and temporal scales using thermochronology, cosmogenic and radiometric isotopes, volume estimates from sedimentary basins, balanced cross sections, fluvial incision models, stratigraphic and geomorphic field observations, and modern sediment-flux data.

The Andes are an exceptional natural laboratory for evaluating the role of erosion in the evolution of a mountain belt. The Andes possess significant along-strike contrasts in styles of deformation, uplift, climate, and presumably denudation. For example, the Andes span 60° degrees of latitude, resulting in a climatic gradient that ranges from tropical to arid. A continental-scale analysis of Andean topography and mean annual precipitation by Montgomery and others (2001) suggested that the distribution of crustal mass is controlled by the style and intensity of denudation in addition to tectonic shortening. Recently, Lamb and Davis (2003) proposed a climatic mechanism for the uplift of the central Andes whereby Mio-Pleistocene cooling and aridification caused the South American trench to become sediment starved, thus focusing the plate boundary shear stresses.

Previous studies have documented strong correlations between the topography, thrust belt geometry, precipitation, orogenic wedge behavior, and erosion at a regional scale in the central Andes of Bolivia (fig. 1) (Masek and others, 1994; Mugnier and others, 1997; Horton, 1999; Leturmy and others, 2000; McQuarrie, 2002b). Differences in the subduction geometry do not appear to be responsible for the along-strike contrasts in deformation because the top of the slab dips uniformly at ~30° throughout this region (Tinker and others, 1996). The general consensus amongst several critical taper, analogue, and numerical modeling studies of the Bolivian Andes is that erosion has played a significant role in the kinematic and topographic evolution of the thrust belt (Mugnier and others, 1997; Horton, 1999; Leturmy and others, 2000).

Our objective is to build upon the previous studies of Bolivian tectonics, climate, and erosion by producing a relative and absolute denudation database that spans multiple spatial and temporal scales. This approach will allow us to evaluate the hypothesis that erosion controls the evolution of the Bolivian Andes, as well as provide quantification of erosion rates that may be used to calibrate numerical models of mountain-belt evolution. To achieve this objective, we synthesized: (1) basin-morphometry, channel network indices, and river profile data as a relative index of basin denudation; (2) sediment-flux and landslide mapping data to estimate short-term denudation rates; and (3) mineral cooling ages, cosmogenic radionuclide concentrations in river sediments, balanced cross sections, basin fill volumes, stratigraphic sections, and paleoerosion surface degradation to estimate long-term denudation rates. Finally, we evaluate this synthesis of denudation rates within the context of Bolivian orogenic-wedge behavior to discuss the implications for the role of climate-driven erosion in thrust belt evolution.

**PHYSICAL SETTING: TOPOGRAPHY, TECTONICS, AND CLIMATE**

The Andes are the product of Cenozoic crustal shortening and thickening associated with Nazca plate subduction beneath the western margin of South America.
Fig. 1. Tectonics, topography, and climate of the eastern Bolivian Andes. (A) Major provinces: the Altiplano (AL), Eastern Cordillera (EC), Interandean zone (IA), and Subandean zone (SA). Inset (simplified from Horton and others, 2001) shows location of the Bolivian Andes in west-central South America. DEM base is the USGS GTOPO30 at 1 km resolution. (B) Topography and precipitation profiles (from Horton, 1999). Topography data is based on a 100-km wide bin along each profile. Maximum-minimum elevation is shaded, average is black line. Width of shaded zone is relief. Mean annual precipitation data (circles) are projected from sites within 200 km of section line. High relief, narrow thrust belt, and wet climate north of the orocline contrasts with the low relief, wide thrust belt, and semi-arid climate to the south.
(for example Isacks, 1988). The Andes exhibit great variation in geomorphology and style of deformation along their length of more than several thousand kilometers (Kley and others, 1999; Kennan, 2000; Montgomery and others, 2001; McQuarrie, 2002a). The Bolivian Andes (14-22°S) exemplify this physical variability on a regional scale because they contain a dramatic contrast in topographic relief, structural geometry, kinematics, and climate between areas north and south of the oroclinal bend at 17.5°S (fig. 1). As discussed by Horton (1999), the north has a wet climate, high relief, and narrow thrust belt that may be characterized by out-of-sequence deformation. In contrast, the south has a semi-arid climate, lower relief, wide thrust belt, and may be characterized by more in-sequence thrusting.

The Bolivian Andes step downward to the east over variable horizontal distances of 200 to 400 km (fig. 1). Figure 1 highlights the four morphological-structural provinces of the thrust belt (after Kley, 1999; McQuarrie, 2002b) and the latitudinal contrast in topography and precipitation (after Horton, 1999). The Altiplano (AL) is an internally-drained plateau of ~3.8 km in average elevation covered by Neogene to Quaternary sedimentary and volcanic rocks and is bounded on the west by a modern volcanic arc and to the east by the high peaks (up to >6 km) of the Eastern Cordillera (EC). The EC consists of mostly lower Paleozoic rocks unconformably overlain by Mesozoic to Neogene sedimentary rocks and is ~3.5 km in average elevation decreasing eastward from the 6-km high peaks to less than 3 km at its eastern edge. The Interandean zone (IA) contains Silurian to Devonian rocks disconformably covered by late Paleozoic to Cretaceous rocks and is ~2 km in average elevation. The Subandean zone (SA) contains mostly Carboniferous to Neogene rocks and is ~1 km in average elevation decreasing eastward to the modern foreland. Through detailed structural cross-section balancing and restoration, Kley (1999) and McQuarrie (2002b) have inferred the presence of two large basement thrusts that fed slip into the overlying cover rocks. These thrusts produce large structural steps controlled by their ramps and terminations. The 6-km high structural steps at the terminations of these thrusts manifest themselves at the surface as 1-km high topographic steps (McQuarrie, 2002b). These basement highs and topographic steps are what mark the EC-IA and IA-SA boundaries (Kley, 1996, 1999). The regional surface faults coincident with the structural and topographic steps at the EC-IA and IA-SA transitions are the Main Andean Thrust and the Main Frontal Thrust, respectively (Sempere and others, 1988). However, correlation between mean topography and the structural steps breaks down in the compacted northern portion of the thrust belt (Kley, 1999).

The chronology of mountain building in the central Andes is disputed. Contraction began in the Eastern Cordillera somewhere between Eocene (McQuarrie and others, 2005) and late Oligocene time (Isacks, 1988; Gubbels and others, 1993). Eocene mountain building in the EC is constrained by thermochronology at c. 40 ± 5 Ma (Benjamin and others, 1987; Masek and others, 1994). Late Oligocene contraction in the EC is bracketed by ages (20-25 Ma) of adjacent foreland basin deposits (Sempere and others, 1990; Jordan and others, 1997). Deformation reached the SA as early as ~20 Ma as evident by growth structures and preliminary thermochronometer cooling ages (McQuarrie and others, 2005), with undeformed paleosurface remnants in the EC suggesting deformation to be exclusively concentrated in the SA since ~10 Ma (Gubbels and others, 1993).

The Bolivian orogenic wedge varies dramatically in geometry and topography across the orocline. In the north, the thrust belt is <250 km in length from the Altiplano margin to the foreland, whereas in the south it is ~400 km long (fig. 1). The topographic relief (compare vertical width of shaded regions in fig. 1B) reaches values twice as large in the northern EC compared to the southern EC. The orogenic-wedge decollement slope is twice as steep in the north compared to the south (4° vs. 2°), and average wedge surface slopes are more than three times (3° vs. 0.8°) greater in the
north than in the south (Horton, 1999). In the north, the topographic break between the EC and the SA has receded to the west by 20 to 30 km from the Main Andean Thrust suggesting significant erosion (Masek and others, 1994). The southern EC possesses remnants of a late Miocene erosion surface, the San Juan del Oro surface, which suggests no upper-crustal deformation has occurred there since ~10 Ma (Gubbels and others, 1993). The north is devoid of any similar features (Horton, 1999). However, the magnitudes of shortening are very similar across the orocline where the northern portion has 300 km of shortening (40%) and the south has 326 km of shortening (37%) (McQuarrie, 2002b). The bend (at 17.5°S), which separates the two regions, is a geologically complicated transition zone where the thrust belt changes width abruptly, possibly involving vertical axis rotations, strike-slip and tear faulting, as well as locales of variable shortening (for example, Sheffels, 1995).

Throughout the Cenozoic, the Bolivian Andes have been transverse to the major Hadley-cell-driven wind currents that control global precipitation distribution at approximately 20°S latitude (Gordon and Jurdy, 1986; Montgomery and others, 2001; Garrison, 2002). The Bolivian Andes are located between the equatorial intertropical convergence zone (ITCZ) where low pressure and precipitation is abundant, and the horse latitudes (30° N and S) where dry conditions persist. Easterly winds bring precipitation from across the Brazilian shield such that locally, the northern portion of the Bolivian Andes receive wetter, northeasterly prevailing winds and the southern Bolivian Andes receive drier, southeasterly prevailing winds (fig. 2) (Garreaud and others, 2003). The inverse correlation between the width of the thrust belt and mean annual precipitation is suggestive of the notion that erosion has taken a large bite out of the northern segment (figs. 1 and 2) (Isacks, 1988).
The modern climate has strong correlations with latitude and topography in the eastern Bolivian Andes (figs. 1 and 2). Mean annual precipitation is roughly twice as large in the north compared across areas of similar elevation to the south. In addition, the average stream discharge is much greater in the north (Horton, 1999; Montgomery and others, 2001). Precipitation decreases by half or more from the lowlands of the foreland westward up onto the Altiplano. Masek and others (1994) attributed these climatic gradients to be associated with latitudinal and orographic effects of the thrust belt and the Andean plateau.

**Denudation**

In this section we present relative and absolute denudation information and rates for the eastern Bolivian Andes from basin-morphometry, drainage network analyses, and a synthesis of previous and new estimates from numerous sources. Where possible, we also identify forcing factors associated with each denudation estimate. The basin-morphometry and channel network analyses were carried out at 1 km resolution with the USGS GTOPO30 DEM using RiverTools 2.4 (developed by S.D. Peckham).

**Relative Denudation**

**Basin-morphometry.**—Many studies have correlated controls of basin denudation with basin-morphometry and climatic settings to quantify spatial variations in denudation. Anhert (1970) demonstrated that mean denudation rate in mid-latitude basins is directly proportional to mean basin relief for low relief areas. Other studies showed that denudation is directly proportional to drainage-basin area, maximum basin elevation, mean-trunk-channel gradient, mean annual precipitation, basin-relief ratio (basin relief/maximum basin length), and local relief (Pinet and Souriau, 1988; Milliman and Syvitski, 1992; Summerfield and Hulton, 1994).

The morphology of the three major basins of the eastern Bolivian Andes—the Beni, Grande, and Pilcomayo—and selected basin-morphometric and channel network indices are shown in figure 3 and table 1. The Beni basin exclusively occupies the northern portion of the thrust belt, whereas the Pilcomayo occupies the southern portion. The Grande basin occupies mostly the southern portion, but its headwaters reach northward across the orocline. The northern Beni basin possesses greater maximum basin elevation, relief, and relief ratio than either the Grande or Pilcomayo basins (table 1). These morphologic indices suggest greater relative denudation in the north.

Hypsometric analysis (area vs. elevation) has been used to estimate relative denudation between the Beni and Pilcomayo basins by examining the amount and distribution of basin area relative to its maximum and minimum elevation (Masek and others, 1994). The lower the hypsometric integral, defined as the area beneath the hypsometric curve, the greater the relative basin denudation (Strahler, 1952). Hypsometry of the three large basins—the Beni, Grande, and Pilcomayo—suggest a northward increase in denudation (fig. 4). All of the southern—Grande, Parapeti, and Pilcomayo—basin lowlands possess large fluvial megafans that shift the hypsometry to significant basin area at lower elevations (see ~20% of basin area with uniform hypsometric curve slopes at ≥10% elevation in fig. 4) (Horton and DeCelles, 2001). The northern Beni basin uplands are well dissected, and the lowlands possess the thrust belt valleys and ridges of the SA that show up as steps in the hypsometric curve at lower elevations. In contrast, the Grande and Pilcomayo uplands are less dissected showing significantly more basin area at higher elevations.

**Channel network and river profiles.**—There are several correlations between the basin channel network indices, river profiles, latitude, and thrust belt geology. Slightly higher drainage density and lower channel sinuosity characterize the northern Beni basin compared to the southern basins (table 1). Higher drainage density indicates a
greater amount of river channels per basin area suggesting the existence of a more efficient sediment and water evacuation system in the north. Sinuous channels tend to possess lower gradients than straight channels, also suggesting higher fluvial denudation in the north compared to the south.

The morphology of longitudinal river profiles provides additional assessment of a relative gradient in fluvial denudation that decreases southward. Longitudinal profiles of the major rivers that traverse the thrust belt were calculated by digitizing points where the river channel intersects the topographic contours (20 m contour interval) on 1:50,000 scale topographic maps and computing the straight-line distance between these points. Local channel gradients were calculated using nearest neighboring points along the elevation profiles. Rock units, faults, and other geologic structures along the river courses were collected from geologic maps (Geobol, 1992; Dunn and others, 1995; Geobol, 1996a and 1996b; Geobol-YPFB, 1996; McQuarrie, ms, 2001).

Fig. 3. Drainage basins of the eastern Bolivian Andes. The Beni, Pirai, and Grande basins drain north into the Beni plain, whereas the Parapeti, Pilcomayo, and Bermejo basins drain south into the Chaco plain. The major river basins are outlined with solid black lines (from Rivertools analysis and Guyot and others, 1990, 1994). The gray lines indicate the locations of the river profiles in figure 5.
Table 1
Basin indices of the eastern Bolivian Andes.
Locations of basins are in figure 3

<table>
<thead>
<tr>
<th>Feature</th>
<th>Beni</th>
<th>Grande</th>
<th>Pilcomayo</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area (km²)</td>
<td>85336</td>
<td>65837</td>
<td>87980</td>
</tr>
<tr>
<td>Relief (km)</td>
<td>5.60</td>
<td>4.70</td>
<td>5.23</td>
</tr>
<tr>
<td>Relief ratio</td>
<td>1.57</td>
<td>1.44</td>
<td>1.46</td>
</tr>
<tr>
<td>Drainage density (km⁻³)</td>
<td>1.290</td>
<td>1.282</td>
<td>1.285</td>
</tr>
<tr>
<td>Sinuosity</td>
<td>1.14</td>
<td>1.25</td>
<td>1.23</td>
</tr>
</tbody>
</table>

Figure 5 shows the river profiles and local channel gradients of the six major transverse rivers—Rio Coroico, Rio Grande, Rio Parapeti, Rio Pilcomayo, Rio Pilaya, and Rio San Juan del Oro—draining the eastern Bolivian Andes. Overall, river gradients tend to increase northward. For example, the southernmost river, Rio San Juan del Oro, possesses no gradients greater than 2.5 percent, and the adjacent Rio Pilaya possesses gradients that rarely exceed 5 percent. In contrast, a major river in the Beni basin, the Rio Coroico, possesses dramatically higher local gradients of up to 60 percent. In general, steeper river reaches coincide with across structural-strike orientations whereas reaches that parallel strike are significantly lower in gradient. Stream gradient generally decreases downstream, but knickpoints (slope breaks or interruptions) frequently disrupt this trend. The largest river profile knickpoints are concurrent with the Main Andean and Frontal Thrust faults that mark the large topographic steps and become more pronounced southward from the Rio Grande to the Rio Pilaya (fig. 5). The Rio Coroico profile ends before reaching these steps, but is graded to less than 1 percent slope for the remainder of its course to the Beni plain. The southward increase in knickpoint morphology (most notable when comparing the Rio Pilcomayo and Rio Pilaya profiles in fig. 5) indicates that there is a similar decrease in fluvial
erosion because large rivers tend to grade their profiles quite rapidly (Bull, 2002). This combination of erosion indicators also suggests greater denudation in the north. All of the river profiles cross Paleozoic marine siliciclastic rocks, Permian and Mesozoic nonmarine rocks, and Tertiary synorogenic sedimentary rocks (fig. 5) (McQuarrie, 2002b). Dominant lithofacies include quartzites, sandstones, siltstones, shales, and conglomerates with a general decrease in erodibility with age and concurrent metamorphism (N. McQuarrie, personal communication, 2002). Generalizations about the mechanical properties of the various stratigraphic units identify the Silurian and Cretaceous rocks as relatively weak (McQuarrie and Davis, 2002). Along these river profiles, steeper river reaches generally coincide with older, stronger rocks. However, some reaches over more resistant rocks, such as the Ordovician, are rather low in relative gradient when coincident with significant, large-scale (1-10 km) surface deformation (fig. 5).

Climate.—Mean annual precipitation (IGM, 2000) and measured river discharges (compiled from Guyot and others (1988, 1990, 1993, 1994)) generally decrease southward (figs. 2 and 6, table 2). The northern Rio Coroico traverses a region that
possesses 1000 to 2000 or more mm/yr of mean annual precipitation (figs. 2 and 3). The course of the central Rio Grande flows through a climate of roughly half as much rainfall. The southern Rio Pilcomayo, Rio Pilaya, and Rio San Juan del Oro occupy a region of less than 500 mm/yr of mean annual precipitation in the upper reaches of their watersheds, and almost never exceed 1000 mm/yr. Mean annual river discharge measurements for small basins of similar magnitude in size (10^2-10^3 km^2) range up to two orders of magnitude higher in the northern Beni basin (16-250 m^3/s) compared to the southern most Bermejo basin (1.5-9.0 m^3/s) (table 2). Medium size basin (10^3 km^2) river discharges range from 475–2340 m^3/s in the Beni basin compared to 49–260 m^3/s in the Pilcomayo basin. The outlets of the two largest montane basins north and south, the Beni and the Pilcomayo, measured discharges of 2340 m^3/s and 260 m^3/s, respectively (table 2). Mean annual rainfall decreases by more than half and river discharge can be more than an order of magnitude less from north to south in the thrust belt. Fluvial bedrock incision is widely approximated by a stream power model that is largely a function of channel gradient and river discharge (for example, Whipple and Tucker, 1999). Therefore, stream power and hence predicted fluvial
TABLE 2

Discharge, sediment flux, and denudation-rate estimates for basins of the eastern Bolivian Andes. See figures 3 and 6 for locations. Dates = dates data was collected, DA = drainage area, Q = mean annual discharge, TSS = total suspended solids.

<table>
<thead>
<tr>
<th>Station^</th>
<th>Dates</th>
<th>Elevation (m)</th>
<th>DA (km²)</th>
<th>Q (m³/s)</th>
<th>TSS (10⁶ t/yr)</th>
<th>Denudation* (mm/yr)</th>
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<tbody>
<tr>
<td>Beni Basin</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SI</td>
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<td>1800</td>
<td>272</td>
<td>16</td>
<td>1.4</td>
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<tr>
<td>UN</td>
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<td>1200</td>
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<td>21</td>
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<td>595</td>
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<tr>
<td>TM</td>
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<td>56</td>
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<tr>
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<tr>
<td>EL</td>
<td>1977-83</td>
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<td>23700</td>
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<td>31200</td>
<td>251</td>
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<td>42900</td>
<td>53</td>
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<td>47500</td>
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<td>31</td>
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<td>340</td>
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<td>260</td>
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<tr>
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<td>4.6</td>
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<tr>
<td>CM#</td>
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<td>2100</td>
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<tr>
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<td>TO</td>
<td>1977-82</td>
<td>1900</td>
<td>460</td>
<td>9.0</td>
<td>1.5</td>
<td>1.265</td>
</tr>
<tr>
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<td>1979-82</td>
<td>1900</td>
<td>920</td>
<td>3.8</td>
<td>0.4</td>
<td>0.170</td>
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#Station code changed from source to avoid duplication.
*Estimated TDS/TSS ratio is 13% for the Beni basin, 10% for the other basins when TDS data not available. Assumes a 65% denudation component for the dissolved load and an average rock density of 2.6 g/cm³ (after Knighton, 1998, p. 81).
erosion should decrease southward throughout the eastern Bolivian Andes following the decrease in river gradients, discharges, and precipitation (Horton, 1999).

**Denudation-rate Estimates**

Table 2 and figure 6 tabulate and illustrate the values and locations of short-term denudation-rate estimates calculated from sediment-flux data collected from Andean gauging stations as discussed below. Table 3 summarizes all of the denudation-rate estimates discussed in this paper including their range of values, time span over which they are averaged, method, type, associated uncertainties, interpretations, and sources from which they were compiled or derived. Figure 7 is companion to table 3 and shows the locations of all of the estimates except the sediment-flux gauging stations.

**Short-term denudation rates.**—Previous short-term ( < 100 yrs) denudation-rate estimates for the Bolivian Andes are quite sparse and vary over four orders of magnitude (fig. 7). Blodgett (ms, 1998) mapped landslides across two northern EC drainage basins (~10 km² each). Making assumptions about scar area–landslide volume relationships, sediment delivery to streams, and landslide age, he estimated erosion rates of 4–14 mm/yr. Landslide age was determined by estimating an average landslide scar revegetation time that ranged from 10 to 35 years depending on the elevation zone within which it was located. Elevation, climate, and season during which repeat aerial photographs were taken (and tree-ring cores) controlled the estimate of revegetation time and therefore landslide age. He assumed a 90 percent sediment delivery factor from the hillslopes to the rivers and immediate removal from the basin. However, the assumption of only 10 percent hillslope sediment storage estimated from a small number of landslides and compared across a few repeat aerial photographs taken at different seasons implies that the storage estimate may be lower limit and the resultant denudation rate an upper limit.

Blodgett (ms, 1998) extrapolated the 4 to 14 mm/yr erosion rates to an area approximately an order of magnitude larger than the two basins mapped by calibrating Landsat image reflectance spectrums. He concluded that the 4 to 14 mm/yr estimate was appropriate to the high and moderate relief zones of the canyons of the northern EC. Key factors identified as triggering landslides in these two zones were high relief, relatively moist conditions, and sufficient stream power (slope and discharge) to undermine the slopes (Blodgett, ms, 1998). In contrast, the dominant erosional processes in the highest, previously glaciated valleys of the northern EC are rockfalls and avalanching rock debris because they are transport-limited environments where stream power is minimal (Blodgett, ms, 1998).

Summerfield and Hulton (1994) reported total basin denudation-rate averages from sediment-flux data for the Amazon and Parana basins. The Beni and Grande basin rivers are tributaries to the Amazon River, and the other basins that drain into the Chaco plain are tributaries to the Parana River (fig. 3). Normalized to basin area, the estimates are 0.093 mm/yr for the Beni basin and 0.014 mm/yr for the Parana basin (fig. 7). In addition, Blodgett (ms, 1998) calculated denudation rates from sediment-flux data near the landslide study area. These estimates range from 0.0073 to 8.1 mm/yr (fig. 7). He noted that the two lowest-magnitude rates were from formerly glaciated valleys.

To first order, river sediment flux integrates the diverse affects of geomorphology, climate, lithology, and tectonics that influence erosion and sediment storage in a given basin source area and can be used to estimate a basin-averaged denudation rate (Hovius, 1998). Unfortunately, the short time involved in data collection and the stochastic nature of sediment delivery, transport, and climate, make denudation rates calculated from sediment flux measurements only approximations (Hovius and others, 2000; Fuller and others, 2003). Regardless, sediment flux derived erosion-rate

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estimates remain the dominant method for estimating annual to decadal time-averaged erosion rates (for example, Summerfield and Hulton, 1994).

The importance of bedload (usually ~10% of suspended load but can be up to 50%; Garde and others, 2004) and upstream areas of deposition ignored by sediment flux derived denudation rates might suggest that they are minimum rates (Gregory and Walling, 1973; Milliman and Meade, 1983; Hovius, 1998). However, factors such as

<table>
<thead>
<tr>
<th>Rate Range (mm/yr)</th>
<th>Time Span</th>
<th>Method</th>
<th>Type</th>
<th>Uncertainties</th>
<th>Interpretation</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.014 ± 0.093</td>
<td>10s yrs ago or less</td>
<td>sediment flux</td>
<td>basin area average</td>
<td>short time-scale</td>
<td>approximation</td>
<td>Summerfield and Hulton, 1994</td>
</tr>
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<td>0.0073 ± 0.1</td>
<td>10s yrs ago or less</td>
<td>sediment flux</td>
<td>basin area average</td>
<td>short time-scale</td>
<td>approximation</td>
<td>Blodgett, ms</td>
</tr>
<tr>
<td>0.049 ± 0.7</td>
<td>10s yrs ago or less</td>
<td>sediment flux</td>
<td>basin area average</td>
<td>short time-scale, assumed rock density and some TDS/TSS</td>
<td>approximation</td>
<td>Guyot and others, 1988, 1990, 1993, 1994</td>
</tr>
<tr>
<td>4.14</td>
<td>0-35 yrs ago</td>
<td>landslide mapping</td>
<td>area</td>
<td>assumed area-volume relationship, age, and 10% basin storage</td>
<td>upper limit</td>
<td>Blodgett, ms</td>
</tr>
<tr>
<td>0.04-1.35</td>
<td>400-15,000 yrs ago</td>
<td>cosmogenics</td>
<td>basin area average</td>
<td>assumed quartz fraction, ignored nuclns, and neglected shielding ~20%</td>
<td>over-estimate?</td>
<td>Safran and others, 2005</td>
</tr>
<tr>
<td>0.13-0.2</td>
<td>0.2 or 3 Ma</td>
<td>mass balance</td>
<td>area</td>
<td>timing of erosion surface incision error ~25%, fold-thrust belt area estimate</td>
<td>Gabbels, ms</td>
<td></td>
</tr>
<tr>
<td>0.13-0.27</td>
<td>0.2 or 3 Ma</td>
<td>seismic basin fill area</td>
<td>area</td>
<td>15% error in area, proximal source, age range 2-3 Ma</td>
<td>minimum?</td>
<td>Gabbels, ms</td>
</tr>
<tr>
<td>0.33</td>
<td>0-5 Ma</td>
<td>stratigraphic sections</td>
<td>point sources</td>
<td>proximal source only, averaged basin fill rate</td>
<td>minimum?</td>
<td>Flemings and Jordan, 1989</td>
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<tr>
<td>0.04-0.09</td>
<td>0.1-0 Ma</td>
<td>erosion surface/DEM analysis</td>
<td>area</td>
<td>time-transgressive (3-10 Ma) age used 10 Ma, includes upper bound of material area</td>
<td>minimum</td>
<td>Gabbels, ms</td>
</tr>
<tr>
<td>0.1-0.8</td>
<td>0.6 or 50+ Ma</td>
<td>AFT, ZFT cooling ages</td>
<td>point sources</td>
<td>15% frrm ages, AFT corrected for topography, ZFT assumed geotherm</td>
<td>AFT more accurate</td>
<td>Benjamin and others, 1987, Masek and others, 1994, Safran, ms, Anders and others, 2002</td>
</tr>
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<td>0.9-1.6</td>
<td>0-12 Ma</td>
<td>AFT cooling ages</td>
<td>point sources</td>
<td>20% from ages, assumed geothermal gradient</td>
<td>approximation</td>
<td>Ege and others, 2003</td>
</tr>
<tr>
<td>0.07-0.84</td>
<td>0.15 or 20 Ma</td>
<td>cross-sections</td>
<td>area</td>
<td>20%, averaged stratigraphic thicknesses, proximal source, present day thrust belt length scale</td>
<td>approximation</td>
<td>McQuarrie, 2002b</td>
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<tr>
<td>0.1-0.6</td>
<td>10-40 Ma</td>
<td>AFT cooling ages</td>
<td>point sources</td>
<td>20% from ages, assumed geothermal gradient, cooling stopped by 10 Ma</td>
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<td>Ege and others, 2003</td>
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<tr>
<td>0.08-0.27</td>
<td>0-40 Ma</td>
<td>cross-sections</td>
<td>area</td>
<td>20%, averaged stratigraphic thicknesses, proximal source, present day thrust belt length scale</td>
<td>approximation</td>
<td>McQuarrie, 2002b</td>
</tr>
</tbody>
</table>

*Reference identifies the source from which the estimate was reported or from which data was modified to create the estimate in this study.

Table 3

Denudation-rate estimates for the eastern Bolivian Andes. Symbols below rates correspond to figure 7. Listed in increasing order of time averaging. AFT, ZFT = apatite and zircon fission track.
anthropogenic disturbance of the landscape might have accelerated modern erosion rates directly or indirectly by increasing a landscapes’ sensitivity to climate (Milliman and Syvitski, 1992; Hooke, 2000; Hewawasam and others, 2003). Indeed, some parts of the Bolivian Andes, especially around the larger cities are cultivated and used for grazing. However, recent work that discusses sediment-yield data from most of the same gauging stations we consider in this study concluded that they represent the natural annual-decadal rate of sediment delivery from the hillslopes for several reasons; the remote, mountainous locations limit floodplain storage and minimize anthropogenic disturbance, and the modern climatic conditions have been relatively stable for the last 1500 years (Aalto and others, 2006).

We now describe a simple method for calculating short-term denudation-rate estimates from sediment-yield data and apply it to data collected from the montane basins of the eastern Bolivian Andes. Table 2 shows the compiled sediment-flux data (from Guyot and others, 1988, 1990, 1993, 1994) and our calculated denudation-rate estimates. The data are ordered by basin from north to south including station codes (locations shown in fig. 6), duration of data collection, upstream drainage area (DA), mean annual discharge (Q), total suspended solids (TSS), and calculated denudation rates including averages grouped by basins of similar order of magnitude in drainage area. We calculated denudation rates assuming total denudation for the total suspended solids (TSS), a 65 percent denudation rate component for the total dissolved solids (TDS), and a rock density of 2.6 g/cm³ after Meybeck (1979) and Knighton (1998) (table 2). Thirty-five percent of the TDS fraction is ignored because it is considered to represent the atmospheric inputs into rainfall and CO₂ in weathering reactions and not the result of denudation processes (Meybeck, 1979; Walling, 1987). TDSs were reported for approximately half of the Beni and Pilcomayo basin stations, none of the Grande basin stations, and all Bermejo basin stations. Of those stations that

Fig. 7. Denudation-rate estimates for the eastern Bolivian Andes. Map Key lists estimates from north to south. Method = method used for calculating the denudation rate, Time Span = time span over which the denudation rate is averaged, sed flux = sediment-flux data, landslides = landslide mapping, AFT = apatite fission-track thermochronology, cosmo = cosmogenic radionuclides, x-section = cross sections, mass bal = mass balance, ES/DEM = erosion surface and DEM analysis, seismic = seismic cross-sectional area, basin fill = basin fill rate. See table 3 for summary of additional information associated with each estimate and text for discussion.
reported TDS, the average TDS/TSS ratio was 13 percent for the Beni basin and 10 percent for the other basins. Therefore, we used the 13 percent TDS/TSS average to calculate the TDS for the remaining Beni basin stations and 10 percent TDS/TSS for the southern (Grande and Pilcomayo) stations for which the TDS was not reported. We chose to use these local averages from what little data was reported because TDS/TSS ratios can vary dramatically despite the average quoted ratio of 6:1 (~17%) assigned to the relative efficacy of mechanical (TSS and bedload, the latter of which is not included in our denudation rates) versus chemical erosion (Walling, 1987; Summerfield and Hulton, 1994). No corrections were made for anthropogenic components since they have not been quantified in any way for this region and are considered to be minimal (Aalto and others, 2006).

The Beni basin exhibits the greatest denudation at multiple basin scales (table 2). Denudation-rate averages are generally greater by a factor of two or three for the Beni basin in the north compared to the Pilcomayo and Bermejo basins in the south. Beni basin denudation rates are 0.2 to 2.5 mm/yr compared to 0.05 to 0.8 mm/yr for the Pilcomayo basin. The average denudation rate is 1.3 mm/yr for the Beni basin and 0.3 mm/yr for the Pilcomayo basin. The Pirai and Grande basins centered at the orocline bend range from 0.1 to 0.9 mm/yr and 0.2 to 4.7 mm/yr, respectively. The Pirai basin, located at the orocline, occupies a localized area of enhanced precipitation (1000-1500 mm/yr mean annual) relative to the Grande basin, but does not extend much higher than 1 km in elevation at its headwaters (fig. 3).

The Grande basin is unique for several reasons. It possesses headwaters that reach both north and south of the orocline and hence encompasses both wet and dry climate regimes (figs. 2 and 3). It drains to the south of the bend, shares morphometric features intermediate to the Beni and Pilcomayo basins as well as intermediate precipitation magnitudes (figs. 2-4, table 2). The Grande basin also possesses the largest average denudation rates, though the largest basin-derived rates are nearly identical to those of the Beni basin (1.6 vs. 1.5 mm/yr).

Basin size appears to influence denudation rate in the eastern Bolivian Andes (table 2). For example, small basins (area = $10^2$ km$^2$) within the Beni and Pirai watersheds average 0.7 to 1.0 mm/yr of denudation in comparison to 0.4 mm/yr for the small basins of the Bermejo watershed. Denudation-rate averages of medium-sized basins (area = $10^3$ km$^2$) within the Beni and Grande watersheds are 1.5 to 2.5 mm/yr compared to 0.4 mm/yr for medium-sized basins within the Pilcomayo watershed. Denudation-rate averages of the largest-sized basins (area = $10^4$ km$^2$) within the Beni and Pilcomayo watersheds are 1.5 mm/yr and 0.3 mm/yr, respectively. Similar to measurements in western Canada, the highest relative denudation-rate averages are in the medium-sized basins (Slaymaker, 1987). Some denudation rates decrease with increasing basin size probably reflecting the influence of sediment storage processes (Milliman and Meade, 1983). This is certainly true when comparing the total basin denudation-rate averages for the Amazon and Parana watersheds, which are an order of magnitude lower than their montane basin components (compare fig. 7 and table 2).

In summary, elevation, climate, relief, glaciation, slope, and basin size are identified as potential forcing factors in the short-term denudation-rate estimates synthesized here. Elevation zone and climate affect the revegetation time estimates for the landslide scars thus influencing their age estimate, and consequently the rate of denudation. Relief, climate, and slope influence landsliding as well. Previously glaciated valleys possess some of the lowest sediment fluxes in the north. The average denudation rates for the Pilcomayo and Bermejo watersheds in the south are 3 to 4 times slower and more uniform than in the northern Beni watershed.
Long-term denudation rates.—Previous long-term (>100 yrs) denudation-rate estimates come from low-temperature thermochronology and cosmogenic radionuclides. Since erosion is the primary mechanism of rock exhumation in thrust belts, low-temperature thermochronometers can constrain denudation rates at long time scales (1-10s Myrs) because they delimit rock cooling histories in the shallow (upper 2-5 km) crust (for example Gleadow and Brown, 2000). Thermochronometer cooling ages are based on the thermally sensitive retention of radiogenic decay products such as fission tracks and \(^{4}\)He in minerals below temperatures of less than \(\sim 250^\circ C\) (for example Gallagher and others, 1998; Farley, 2002). As a result, the apatite and zircon fission track and (U-Th)/He systems are established tools for reconstructing the exhumation histories of thrust belts (Ehlers and Farley, 2003).

Overall, denudation-rate estimates from thermochronology range from 0.1 to 0.8 mm/yr from the northern EC (fig. 7). Safran (ms, 1998) conducted apatite fission-track analyses on more than 20 granitoid samples from the EC that included the dataset of Benjamin and others (1987). Safran (ms, 1998) estimated 0.2 to 0.6 mm/yr of erosion from samples aged 6 to 20 Ma (fig. 7). These denudation-rate calculations incorporated topographic influences on subsurface temperature gradients improving their accuracy (Stuwe and others, 1994). Apatite and zircon fission-track data from the same location (hence not shown in fig. 7) recorded 4 to 8 km of material eroded over the last 10 to 15 Ma, which corresponds to 0.3 to 0.8 mm/yr of erosion (Benjamin and others, 1987; Masek and others, 1994). Recent work on the Benjamin and others (1987) dataset suggests that erosion rates have increased in the last 50 Ma from 0.1 mm/yr to as much as 0.75 mm/yr at present in either an exponential or step-wise fashion (Anders and others, 2002).

Ege and others (2003) analyzed apatite fission-track cooling ages from \(\sim 30\) samples of Paleozoic to Tertiary meta-sedimentary and sedimentary rocks across the entire thrust belt at 21°S (fig. 7). They estimated denudation rates from samples aged \(\sim 7\) to 75 Ma. Assuming deformation stopped in the plateau region (AL and western EC of fig. 1) by 10 Ma (after Gubbels and others, 1993), they calculated denudation rates of 0.1 to 0.6 mm/yr in the AL and EC. Furthermore, they estimated 0.4 mm/yr of erosion in the IA, and 0.9 to 1.6 mm/yr in the SA.

Safran and others (2005) estimated upstream basin-averaged erosion rates in the northern EC and IA by measuring \textit{in situ} \(^{10}\)Be from quartz in river alluvium (fig. 7). They estimated 0.04 to 1.35 mm/yr of erosion averaged over time scales of 400 to 15,000 yrs from basins ranging in size from 1 to 70,000 km\(^2\). They observed that the erosion rates correlate with a metric of channel gradient which appears to be primarily influenced by the tectonics. Lack of correlation between lithology, climate, and the erosion rates led them to conclude that both factors play a secondary role at best in influencing the erosion rates.

We estimated long-term denudation-rate estimates for the eastern Bolivian Andes from a wide range of source materials including an erosion surface and its subsequent incision, sediment volume in the foreland basin, and balanced cross sections (fig. 7). We now describe the methods used to estimate these long-term denudation rates.

The San Juan del Oro surface is a regional paleoerosion surface identified by a combination of features including remnants of low-relief uplands, coalesced pediments, and a prominent Tertiary unconformity in the stratigraphic record (Gubbels and others, 1993). Gubbels (ms, 1993) estimated the amount of material removed by erosion from below the San Juan del Oro surface in the southern EC since its formation \(\sim 10\) Ma. From DEM analysis along three transects, Gubbels concluded that 120 to 240 km\(^2\) of material was removed from below the surface. He approximated the surface elevation as the average between the maximum and average modern topographic envelopes, based on surveyed locations of the erosion surface remnants. The
120 to 240 km\(^2\) estimate range includes an upper bound in which the maximum topographic envelope was used to approximate the erosion surface elevation. Though only presently occupying one third of the EC east of the plateau (defined roughly as the 3 km elevation contour, Isacks, 1988) margin, the San Juan del Oro surface presumably once covered a large region of the southern thrust belt spanning ~300 km across-strike and ~450 km along-strike (Gubbels, ms, 1993). Assuming a 10 Myr time period and across-strike distances of 315, 285, and 260 km (measured from Gubbels, ms, 1993, fig. 4.13, p. 155), we converted these areas into denudation rates ranging from 0.04 to 0.09 mm/yr (fig. 7). Erosion surface remnants are bracketed by ~3 to 10 Ma ages (Kennan and others, 1997). This suggests that the San Juan del Oro surface formation was time-transgressive. Therefore, using the maximum age in our rate calculation implies that these denudation-rate estimates are a minimum even though they incorporate the upper bounds of material area removed. Constrained by K-Ar dated tuff deposits near the top of the erosion surface, further mass balance calculations involving the entire, regional erosion surface inferred 400 m of denudation since surface incision began at 2 to 3 Ma (Gubbels, ms, 1993). We calculated this denudation-rate estimate as 0.13 to 0.2 mm/yr over the last 2 to 3 Ma (fig. 7). However, actual timing of incision into the erosion surface is more spatially variable and bracketed by tuffs ranging in age from 1.5 to 3.5 Ma (Kennan and others, 1995, 1997).

We calculated additional denudation-rate estimates from sediment volume data in the southern foreland basin. Gubbels (ms, 1993) estimated the amount of material denuded over the last 2 to 3 Ma as 140 \(\pm\) 20 km\(^2\) from Tertiary strata imaged by seismic data in the foreland basin along a single profile at 20°S. Assuming that the Tertiary sediment was eroded from the most proximal part of the thrust belt, which is 300 km wide, we calculated a denudation rate range of 0.13 to 0.27 mm/yr (fig. 7). This estimate incorporates an age range of 2 to 3 Ma and the error in the material area (\(\pm\)20 km\(^2\)). Flemings and Jordan (1989) reported a Subandean basin fill rate of 82 m\(^2\)/yr over the last 5 Myr at 22°S. With a proximal thrust belt width of 250 km, we estimated 0.33 mm/yr of denudation over the last 5 Myrs (fig. 7). Mass-balance calculations that suggest the Andean foreland basin has been overfilled imply that these basin fill denudation estimates may be minimum rates (Gubbels, ms, 1993). Furthermore, recent estimates from the northern EC suggest that about half of the sediment shed from the Andes is stored in the foreland (Aalto and others, 2006).

We calculated denudation rates from balanced cross sections and their restorations for both the north and south by estimating the material area removed from above the present day topographic surface of the thrust belt (lines B-A’ & X-Y’ in fig. 7) (cross sections from McQuarrie, 2002b). We measured line lengths of the oldest continuous contacts above the present topography, added the appropriate average thickness of pre-Tertiary rocks for each physiographic province estimated from the restored sections, and added that to additional areas of older rocks above the topographic surface. Finally, the estimated area of material removed by erosion was divided by the most proximal modern length scale of the thrust belt and its age along each section to produce the denudation-rate estimate. Figure 8 illustrates and outlines an example of this procedure applied to the eastern portion of the southern Subandean zone showing ~100 km\(^2\) of rock material estimated to be removed by erosion. For the south at ~20°S, we used average thicknesses of 1.5 km for the Mesozoic section in the EC, and 1 km in the IA and SA. For the north at ~17°S, we used average thicknesses of 1.3 km for the Silurian, 800 m for the Devonian, and 800 m for the Mesozoic sections, respectively. We estimate 425 km\(^2\) of material removed from the 50 km wide SA in the north compared to 192 km\(^2\) from the 130 km wide SA in the south. In total, we estimate 1297 km\(^2\) of material was removed from the 210 km wide thrust belt in the north, and 1264 km\(^2\) from the 380 km wide thrust belt in the south.
DeCelles and Horton (2003) traced the early to middle Tertiary migration of the Andean foreland basin to its present location. This suggests Tertiary foreland basin deposits were present across the entire thrust belt and that in some regions, such as the IA, have been completely removed by erosion. Assuming a 3 km thick succession of Tertiary foreland basin deposits across the entire thrust belt (add 3 km thickness to step 2 in fig. 8), our estimates of material area removed increases to 2256 km$^2$ in the north, and 2357 km$^2$ in the south. This adjusts our SA estimates to 627 km$^2$ and 432 km$^2$ in the north and south, respectively. Favoring the most recent estimates of Andean chronology, we assume the thrust belt in the EC began deforming at 40 Ma, and the Subandean zone at 15 to 20 Ma (McQuarrie and others, 2005). We estimate 0.08 to 0.16 mm/yr of denudation in the south and 0.15 to 0.27 mm/yr in the north. The lower values were calculated for just the Paleozoic and Mesozoic rock areas and the higher values reflect areas that include the additional 3 km of Tertiary foreland basin deposits. For the SA, we estimate 0.07 to 0.22 mm/yr of denudation in the south compared to 0.43 to 0.84 mm/yr in the north. These SA estimates similarly reflect the differences in material estimates as well as an age range of 15 to 20 Ma. The error in material area is associated with errors in stratigraphic thickness of the various rock units involved which is ~20 percent (N. McQuarrie, personal communication, 2005). Adjustment in age by 5 Myrs or less has negligible affect on these denudation-rate estimates.

The most important factors that influence the long-term denudation-rate estimates calculated from balanced sections are the time and length scales over which they are averaged. Theoretically, the denudation processes that removed these materials were influenced by all of the geomorphic, geologic, and climatic factors that affect erosion. The modern thrust belt width is well defined, but the assumption of using the most proximal width as the length scale may not be appropriate. Perhaps material volume or cross-sectional area alone is a more appropriate unit in which to compare them. The values are 140 km$^2$ over the last 2 to 3 Ma from seismic data (Gubbels, ms, 1993), 82 m$^2$/yr over the last 5 Ma from basin fill rate (Flemings and Jordan, 1989), and ~120 km$^2$ over the last 10 Ma from the paleoerosion surface and DEM analysis (Gubbels, ms, 1993). Average rates derived from these estimates are ~58 m$^2$/yr, 82 m$^2$/yr, and ~12 m$^2$/yr. For these cases, the differences in areas are equivalent to the differences in magnitude of the rates. For the balanced sections, the areas of material removed are 1297 to 2256 km$^2$ and 425 to 627 km$^2$ for the total thrust belt and SA in

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**Fig. 8.** Diagram illustrating methodology for material removed by erosion from above the present-day topography of a balanced section. Example section is from the eastern portion of the southern Subandean zone (from McQuarrie, 2002b). Method: (1) Measure line length of Mesozoic (M)-Carboniferous (C) boundary above the topography (dashed contact), (2) add average thicknesses of units above the Carboniferous (in this case: 1 km Mesozoic strata), (3) measure area above surface, but below K-C boundary (grid pattern), (4) add steps 2 and 3 results for total estimate of material removed by erosion.
the north, respectively. For the south, the total estimate is 1264 to 2357 km$^2$ and for the SA it is 192 to 432 km$^2$. In this case, the areas are nearly equal for the entire thrust belt, but approximately a third larger for the SA in the north. This shows a strong dependence of the thrust belt geometry on the denudation estimate. The southern denudation rates are equal to half those in the north (0.08-0.16 vs. 0.15-0.27 mm/yr). For the SA, denudation rates in the south are a quarter of those in the north (0.07-0.22 vs. 0.43-0.84 mm/yr). The material area estimates imply more similar amounts of material removed over the entire thrust belt than the denudation rates. However, both perspectives display the same relative trends to different degrees: greater denudation in the north, denudation increasing to the present, and the disparity in denudation between the north and south increasing to the present.

**Discussion**

The contrasts in basin-morphometry, channel network features, river profiles, and climate described above provide a semi-quantitative assessment of relative denudation across the Bolivian Andes correlative with latitude. The denudation-rate estimates presented here constrain how erosion rates have varied spatially and temporally across the thrust belt. Here we discuss the trends of our denudation-rate database, integrate them with our basin-morphometry and river profile observations, and place them in the context of the history of uplift, climate, and wedge dynamics in Bolivia.

**Denudation Rates: Trends Across Time and Space**

Despite the uncertainties in the various denudation-rate estimates we have calculated and synthesized, we observe that a disparity in denudation (greater in the north) has persisted throughout the Holocene and perhaps since the late Miocene. Taking the data at face value, with one exception, rates averaged over the last $\sim$10 Ma are higher in the north though the disparity is not large. Rates averaged over the Holocene are higher in the north with a more significant disparity. Additional, secondary trends suggest that the north-south disparity and denudation rates may have increased to the present. Figure 9 illustrates these observations by plotting the ranges of denudation rates versus the time scales over which they were averaged from all the estimates listed in table 3.

Direct comparison of the only denudation-rate estimates derived from the same method and averaged over the longest-term (>15 Myrs) shows higher values in the north compared to the south (fig. 9). These estimates derived from the balanced sections range from twice as much in the north (0.08-0.16 mm/yr vs. 0.15-0.27 mm/yr) across the entire thrust belt to four times as much in the north for the SA (0.07-0.22 mm/yr vs. 0.43-0.84 mm/yr). Given our assumptions of the age of the entire thrust belt (40 Ma) and deformation starting in the SA at 15 to 20 Ma, this implies that (1) erosion was twice as great in the northern part of the thrust belt as early as 40 Ma and (2) over the more recent past (last $\sim$10 Ma), the disparity in denudation has grown to perhaps as much as 4 times greater in the north (fig. 9). The absolute denudation rates estimated for the south are similar in magnitude for both the entire thrust belt as well as the younger SA, whereas the absolute rates in the north increased 4 times from the 40 Myr averaged rates to the 15 to 20 Myr averaged rates for the SA.

Comparison of the apatite fission-track derived estimates shows similar rates north and south (figs. 7 and 9). In the Eastern Cordillera, the range of rates is almost identical north and south (0.1-0.8 mm/yr vs. 0.1-0.6 mm/yr). As previously mentioned, Anders and others (2002) reported up to a seven-fold increase in erosion rate to the present (0.1-0.75 mm/yr) in the north from re-evaluating Benjamin and others (1987) apatite and zircon cooling ages and concluded that denudation rates have doubled every 20 to 35 Ma over the last 50 Ma. Safran’s (ms, 1998) estimates from the northern EC also showed rates accelerating to the present, though they were averaged over ages
from 6 to 20 Ma. It is interesting to note that the sparse fission-track data from the southern SA estimate the most rapid of the long-term averaged denudation rates (0.9-1.6 mm/yr) (figs. 7 and 9) (Ege and others, 2003). Unfortunately, there are no estimates from the northern SA to make any comparison. In general, the thermochronology studies seem to corroborate the temporal part of the cross section story that suggests long-term averaged erosion rates have increased to the present from those averaged over 30 Ma to those averaged over 20 Ma. However, unlike the estimates from the cross sections, the fission-track derived erosion rates do not show any significant difference between the northern and southern portions of the EC.

Almost all of the intermediate time-scale averaged (1-10 Myrs) denudation-rate estimates are located in the southern region and range from 0.04 to 0.35 mm/yr (figs. 7 and 9; table 3). The DEM analysis using the San Juan del Oro erosion surface as a datum produced the lowest erosion rates, possibly because we averaged them over the maximum age of the surface that is time-transgressive from 3 to 10 Ma (Kennan and others, 1997). The other estimates from mass balance, seismic, and basin fill rates all range from 0.13 to 0.33 mm/yr averaged over the last 2 to 5 Ma. These estimates are nearly identical to those estimated for the south averaged over the last 10 to 12 Ma (~0.1-0.3 mm/yr). From this corroboration, we can infer that average rates of erosion
across the southern Bolivian Andes have been on the order of 0.1 to 0.3 mm/yr over the last ~5 to 15 Ma. In comparison, although there are no reported sediment volume data for the north, estimates of modern Tertiary foreland basin thicknesses range from ~3 km in the south to more than twice as much, ~4 to 7 km in the north (Horton and DeCelles, 1997). This suggests that denudation-rate estimates from sediment volumes in the north might be greater than the south by a factor of two or more, especially since recent data suggest only half of the sediment removed from the northern region is trapped in the Beni foreland (Aalto and others, 2006). This is consistent with the disparity in erosion in the north increasing from near double to more than double when averaged over the last ~10 Ma as inferred from the long-term estimates as discussed above.

The shorter-term (400-15,000 yrs) averaged denudation rates from the cosmogenic radionuclides in the northern EC range from 0.04 to 1.35 mm/yr, have a mean value of 0.42 mm/yr, and a mode of 0.2 to 0.4 mm/yr (fig. 9) (Safran and others, 2005). Safran and others (2005) noted that this mode is similar to the fission-track ages (0.2-0.6 mm/yr) suggesting no significant variation in erosion rates over the last several million years. However, the only direct comparison they can make is from samples in the Zongo River valley headwaters where these shorter-term rates (0.4-0.55 mm/yr) exceed the fission track rates by ~15 to 100 percent implying at most a moderate increase in erosion rate to the present (fig. 9) (Safran and others, 2005, p. 1013). The local modern denudation-rate averages calculated from landslide mapping and sediment-flux data range from 1.0 to 9.0 mm/yr further suggesting an increase in erosion rate to the present (fig. 9).

The shorter-term (1-35 yrs) denudation-rate averages from sediment-flux and landslide mapping data in the northern Beni basin (1.0-9.0 mm/yr) are larger than those in the southern Pilcomayo basin by up to more than an order of magnitude (0.3-0.4 mm/yr) (figs. 7 and 9; table 2). For example, the denudation estimate of 4 to 14 mm/yr from landslide mapping in the northern EC is the single largest denudation estimate. Although the estimate is an approximation, it is similar to estimates from landslide mapping in the Southern Alps of New Zealand (Hovius and others, 1997). Even though the northern estimates are larger, their variance encompasses the entire range of the southern rates (fig. 9). The small Pirai basin located directly at the oroclinal bend possesses denudation-rate averages (0.4-0.7 mm/yr) similar to intermediate values from the Beni, Pilcomayo, and Bermejo basins, but are lower than the Grande basin to the south. The only denudation rate for the Parapeti basin (1.0 mm/yr) is intermediate in comparison to the regions directly north and south. The denudation rates for the Grande basin stray from the gross latitudinal erosion gradient of the thrust belt (fig. 6, table 2). These unexpectedly high rates of the Grande basin could be related to an unusually wet period from ~1969 to 1974 when almost all of the basin stations were collecting data yet no others in the region were operative. Perhaps encompassing an area with a particularly wide range of annual rainfall (fig. 3) combined with the more complicated compressional and extensional deformation associated with the transition zone at the bend and the urban Cochabamba area is somehow responsible for the higher erosion.

The short-term (1-35 yrs) denudation rates (~0.01-14 mm/yr) reach larger values than the long-term denudation rates by as much as an order of magnitude (~0.05-1.6 mm/yr) (fig. 9). Similarly, Clapp and others (2000) found that long-term sediment production rates estimated from cosmogenic isotopes were more than 50 percent lower than those estimated from short-term sediment-flux data in an arid basin of southern Israel. They interpreted their results as representative of sediment evacuation following a period of greater weathering and lower sediment transport. This may be the case for the Bolivian Andes as well. In contrast, Kirchner and others (2001) showed
that long-term average denudation rates from cosmogenic isotopes and fission-track thermochronology were 17 times greater than sediment-flux data collected over 10 to 84 years from non-glaciated, montane drainages in Idaho. They concluded that 70+\% of the total long-term sediment delivery from mountain catchments is not recorded by the decadal-averaged rates from sediment-flux data and is therefore highly episodic. Opposite trends in denudation rates averaged over multiple time scales suggest that local conditions, such as climate, influence erosion-rate magnitudes measured with the same methodology as Hallet and others (1996) demonstrated for glaciated regions. Climate has been recognized to play a large role in controlling 1 to 10 Myr time scale terrestrial erosion for northern South America by climate and sedimentation rate correlations in the Amazon fan (Mikkelsen and others, 1997; Harris and Mix, 2002).

The larger variation of the modern denudation-rate estimates (fig. 9) is not surprising because of the stochastic nature of the processes that influence sediment flux and the fact that they are averaged over short time intervals. The increase in denudation rates to the present in the north may reflect the combined affects of the currently hypothesized positive feedbacks associated with uplift, climate, and erosion in the thrust belt and increasing anthropogenic influence. If the anthropogenic affect on the land has been negligible (Aalto and others, 2006), then the north-south erosion disparity and how it has changed through time could, to first-order, result from the combined influences of the latitudinal climate gradient (Montgomery and others, 2001) and the more local positive feedbacks associated with uplift, climate, and erosion (Masek and others, 1994).

**Basin-Morphometry, River Profiles, and Climate: North versus South**

The contrasts in basin-morphometry and longitudinal river profiles across the northern and southern portions of the thrust belt provide evidence suggestive of greater relative denudation in the north (figs. 1 and 5, table 1). Synthesizing these morphologic contrasts with the temporal and spatial trends in the denudation rates provides a framework for the evolution of the thrust belt. Figure 10 schematically illustrates all of the north-south contrasts mentioned in the remaining discussion.

Higher basin relief, drainage density, and high denudation rates characterize the northern Beni basin (fig. 10). High slopes and river profile curvature, large river discharges, and therefore inferred high stream power characterizes the Beni basin rivers such as the Rio Coroico (figs. 5 and 10) (Horton, 1999). Safran (ms, 1998) concluded that the incision of bedrock rivers control the morphology and relief of the northeastern Bolivian Andes. In contrast, lower slopes and river profile curvature, drainage density, denudation rates, and higher sinuosity characterize the southern Pilcomayo basin. Lower slopes, river discharges, and inferred stream power characterize the southern rivers such as the Rio Pilcomayo, Rio Pilaya, and Rio San Juan del Oro. A few localized areas of potentially high stream power at knickpoints are diagnostic of the southern rivers (figs. 5 and 10). The large topographic steps that are controlled by the basement thrust terminations correlate with the large river knickpoints in the south. Faults, folds, and lithologic boundaries correlate with many of the smaller knickpoints along all of the rivers (fig. 5). Reach-scale slopes are influenced by rock type, erodibility, 1 to 10 km scale deformation, as well as reach orientation with respect to structural strike, and tributary addition at lower elevations (fig. 5). As noted by others (Miller, 1990; Lecce, 1997; Knighton, 1999), these various, identifiable knickpoint controls create non-linear changes in stream power, and hence predicted incision along a river profile. Simulations of sediment flux from thrust belts have also demonstrated the importance of rock strength and erodibility (Tucker and Slingerland, 1996). In addition, Aalto and others (2006) showed that lithology and average catchment slope accounted for 90 percent of the variance in sediment yield measured from the Bolivian Andes.
The major morphological-structural provinces bounded by topographic steps in the southern region could influence the basin-averaged denudation rates from the sediment-flux data. Small basins are most immediately graded to the trunk-rivers for which they are tributaries. Basins located on the benches above the topographic steps in the south (three outlined basins of the southern region in fig. 10) are graded to the semi-regional flats and hence subject to less denudation as a result of low local stream power and relief, in comparison to basins located elsewhere. In other words, areas in between or above the major knickpoints in the southern EC may have yet to “feel” the potentially greater incision downstream. This can be seen best in the rate estimates from stations such as EP, CH, SJ, ER, and PJ in the Pilcomayo and Bermejo basins (table 2). The existence of the large knickpoints in the southern rivers, where stream power should be most effective at degradation because of their large discharge and
slope, is additional testament to lower erosion in the south. Furthermore, larger thrust belt width and greater spacing between anticlines is diagnostic of the southern SA (McQuarrie, 2002b). The greater spacing of folds in the southern SA facilitates more strike-parallel river orientations in synclinal valleys where slopes are low, fluvial erosion is reduced, and more sediment is stored. This synthesis of basin denudation and morphometry observations schematically suggests correlations between uplift, kinematics, climate, and geomorphology on fluvial erosion in the Bolivian Andes (fig. 10). Although it is difficult to demonstrate how long the current drainage basins and fluvial networks have persisted back in time, evidence from ancient fluvial megafan deposits suggests that the southern drainage systems may be inherited from the mid-Tertiary (Horton and DeCelles, 2001).

**History of Uplift and Climate in the Bolivian Andes**

The history of uplift and climate change in the central Andes provides a broad context within which to evaluate the spatial and temporal variation in erosion through time. Evidence from fossil floras, the San Juan del Oro paleoerosion surface, and crustal shortening suggests there was significant uplift since the late Miocene in the Altiplano and Eastern Cordillera, a shift to aridity at 15 Ma in the Altiplano, and an increase in erosion at 10 to 15 Ma (Gregory-Wodzicki, 2000). Gregory-Wodzicki (2000) concluded that since there was global evidence for drying at 15 Ma, the aridity shift was probably the result of climate change and not an uplift-induced rain shadow. More recent work has recognized rapid surface uplift of the Altiplano in the late Miocene measured with oxygen isotope compositions of carbonate deposits, presumably resulting from delamination of the mantle lithosphere (Garzione and others, 2006). Significant late Miocene uplift certainly could explain the concomitant increase in erosion in the late Miocene inferred from the fission-track data, our erosion synthesis, and an increase in terrigenous flux to the Amazon fan (Gregory-Wodzicki, 2000; Anders and others, 2002; Harris and Mix, 2002). Regardless of whether it was climate change or uplift or both, it would follow that average denudation rates calculated in this study that integrate over more than the last ~10 Ma should be lower than estimates that do not. This phenomenon is most evident for the southern region data where more long-term denudation-rate estimates exist (fig. 9). Furthermore, Zhang and others (2001) proposed that a global increase in sedimentation rates in Pliocene time was the result of increased erosion forced by increased climatic variability. Though they report no evidence from South America, recognition that global Plio-Quaternary climate possesses more variability then previous time is consistent with the trend of our results of increased erosion rates in more recent time.

Glaciers presently occupy only the highest peaks of the northern EC. To a limited extent, glaciation may have enhanced relief in the north and, by inference, relative denudation (Tomkin and Braun, 2002). Since the late Pleistocene glacial maximum, aridity has resulted in limited glaciation of even the 6 km summits in southern Bolivia (Fox, ms, 1993). Aridity of only the highest mountain regions may be part of the reason why Quaternary glaciation was more extensive in the north relative to the south though still limited to only peaks at elevations of above ~5 km (Clapperton, 1984). The influence of glaciers on mountain relief, erosion, and exhumation has certainly played a large role in other mountain belts (Hallet and others, 1996; Brozovic and others, 1997; Meigs and Sauber, 2000), but appears to have played a more minor role in the eastern Bolivian Andes. The few reported sediment-flux rates from previously glaciated valleys in the north (Blodgett, ms, 1998) possess the lowest denudations suggesting that the sediment liberated by earlier glaciations may have since been largely removed prior to the sediment-flux data acquired by the gauges in the mid-1970s and 1980s.
The Bolivian Orogenic Wedge

In this section, we integrate the denudation story presented in this study with that of previous studies of the external (subduction, global climate patterns), and internal forcing factors on the orogenic wedge, as well as critical taper and kinematic modeling. Figure 10 illustrates the contrasting features of the north and south thrust belt wedge.

**External controls.**—One external tectonic control on the wedge is the subduction process. As noted above, today the Nazca plate subducts uniformly at ~30° below Bolivia (Tinker and others, 1996). Although the segments of flat-slab subduction to the north and south evolved from more steeply dipping geometries illustrating the dynamic aspect of subduction, we assume that the present geometry is representative since the mid-Tertiary (Isacks, 1988). Of additional concern are the convergence and continental margin dynamics since deformation began in the Eastern Cordillera at ~40 Ma. In general, the Nazca plate convergence direction has been consistently oriented northeast-southwest (Pardo-Casas and Molnar, 1987). If the present Bolivian margin morphology has been consistent back through time, it describes more oblique subduction to the south and more orthogonal convergence to the north. This implies more convergence in the north. Although this would suggest more shortening north of the bend, the percent shortening is almost equal across the orocline (McQuarrie, 2002b). In the Miocene, the convergence direction shifted clockwise to a more east-west orientation such that convergence has since been more uniform across Bolivia. McQuarrie (2002a) used restored amounts of shortening to reconstruct the western continental margin back through time to show that the modern orocline only began to take shape ~20 Ma. Previous to that time, the Bolivian continental margin was much straighter, which might have implications for Miocene versus older stages of central Andean orogenesis.

**Global atmospheric circulation.**—The other important external control on the wedge is the global atmospheric circulation patterns that influence precipitation distribution. As mentioned above, the Bolivian Andes have been at approximately the same latitude since deformation began in the Eastern Cordillera (Gordon and Jurdy, 1986). This establishes the current atmospheric circulation patterns and their influence on precipitation as initial conditions for deformation east of the Altiplano (Montgomery and others, 2001). Therefore, the orographic influence on precipitation aside, the north has always received more precipitation from the wetter northeasterly prevailing winds. We might assume global climate change (climatic cooling and increased variability) occurs uniformly across the entire thrust belt, but their affects on denudation might vary locally. Since precipitation is enhanced in the north and reduced in the south both by global atmospheric circulation patterns and by orographic affects it is difficult to separate them. Instead, we conclude that both play a role in influencing the disparity in denudation across the orocline.

**Critical taper.**—Greater relief and larger topographic front and decollement angles should facilitate more critical wedge conditions in the north favoring propagation (Dahlen and Suppe, 1988). However, GPS measurements suggest the north is in a subcritical condition (internally deforming) and the south is in a critical condition (forward propagation) (fig. 10) (Horton, 1999). More involvement of crystalline basement rocks and higher-grade and stronger Paleozoic meta-sedimentary rocks in the southern wedge have been proposed to explain the lower threshold for critical taper in the south (Horton, 1999). If this is true, the weaker rocks in the northern wedge could facilitate more erosion because of their reduced strength. Rock strength and erodibility may be important for orogenic wedge scale erosion and kinematics as well as at the river profile scale as we have observed as have others (Aalto and others, 2006). Reduced rock strength of similar lithologies in the north resulting from lower metamorphic grade (weaker) might contribute to the enhancement of denudation in
the north. However, Safran and others (2005) concluded that lithology played a secondary role to tectonically driven channel gradients in forcing the erosion rates measured from cosmogenic radionuclides in the northern EC.

We can quantitatively compare the northern and southern Bolivian orogenic wedges in terms of denudation across time and space and identify correlations with the wedge geometry, morphology, and kinematics (for example with Horton, 1999). If we assume that erosion is important, then we can estimate that longer-term (>15 Myrs) sustained erosion rates ranging from 0.15 to 0.84 mm/yr (fig. 7) are enough to retard the propagation of wedges of $7^\circ$ taper (product of decollement angle and topographic slope, fig. 10) of relatively low strength. However, long-term sustained erosion rates ranging from 0.04 to 1.6 mm/yr (fig. 7) are not enough to retard the propagation of a strong wedge of $2.8^\circ$ taper. These numbers may be useful in calibrating models of orogenic wedge evolution that incorporate erosion.

Kinematic models and erosion.—Best-fit analogue models of the southern SA involved sedimentation on the wedge, but no erosion (Mugnier and others, 1997). Combined analogue and numerical models of the northern and southern parts of the thrust belt suggest that (1) erosion restrains wedge propagation, and (2) high sedimentation rates associated with basin subsidence promotes piggyback basin formation that first induces a shift forward in the deformation front, followed by out-of-sequence thrusting (Leturmy and others, 2000). The northern wedge possesses large piggyback basins (Baby and others, 1992), corroborating the story of earlier high rates of erosion, sedimentation, and subsidence now characterized by high erosion and perhaps more internal deformation (Horton, 1999). The inference of a sediment evacuation phase following a period of sediment production by Clapp and others (2000) from similar denudation rate trends mentioned above is also consistent with this interpretation. The positive feedbacks associated with enhanced precipitation by both global and orographic affects, erosion, and relief may be contributing to the hindrance of wedge advance in the north. For the southern wedge, lower erosion, but high Tertiary synorogenic sedimentation rates demonstrated by growth strata and perhaps less out-of-sequence thrusting support a story of more continuous wedge propagation eastward (Baby and others, 1995; Horton and DeCelles, 1997).

Almost all of the denudation history and morphology synthesized here provides evidence of more sediments being evacuated from the north than the south, and larger regions of material mass at greater elevations in the south (fig. 10). The geomorphology and kinematics of the south are both consistent with more current sedimentation on the wedge contributing to its propagation. In contrast, little sediment storage is taking place on the northern wedge, but the larger amount of denudation is reflected in a much deeper foredeep basin outboard to the east (fig. 10). However, foredeep basin sediment accommodation is generally the result of flexural subsidence due to topographic, sedimentary, and dynamic (subduction) loads (DeCelles and Giles, 1996). The relatively rapid reduction in hinterland topographic load by upland denudation suggests that the sediment basin load may be controlling the foredeep subsidence in the north. In addition, the structural and topographic steps of the morpho-structural province boundaries are controlled by basement thrust terminations which, in turn, correlate with locales of high incision at the large river knickpoints (figs. 1 and 10). The lower stream power in the south detracts from fluvial efficiency at removing sediment from hillslope processes. In contrast, the province boundaries and knickpoint correlations are significantly subdued or even non-existent in the north. Furthermore, according to Blodgett (ms, 1998), 90 percent of landslide material is essentially moved immediately by the fluvial system due to its higher stream power, suggesting that in the north the fluvial system is efficient at flushing out sediment.
Rates of Uplift and Erosion in the Bolivian Andes

Longer-term (averaged over >10 Ma) denudation-rate estimates range from ~0.1 to 1.6 mm/yr (fig. 7). Estimates of rock and/or surface uplift over the last 25 Ma for the Altiplano and Eastern Cordillera are at the lower end of this range at 0.1 to 0.4 mm/yr (summarized by Gregory-Wodzicki, 2000). However, mantle delamination may have induced ~2.5 to 3.5 km of surface uplift of the Altiplano to its present elevation from ~10.3 to 6.8 Ma (Garzione and others, 2006). This suggests an uplift rate of ~0.7 to 1.0 mm/yr during this 3.5 Myr event in the late Miocene. This uplift event may be reflected in the higher denudation rates averaged over the last 10 Myr or less, but should not contribute to the north-south disparity as presumably the whole plateau region uplifted. It appears that denudation rates may have reached values equal to or greater than inferred uplift rates at times in the AL and EC. In the south at ~20°S, current rock uplift rates from space-geodetic data are 0.01 to 0.1 mm/yr in the EC and 0.5 to 2 mm/yr in the SA (Lamb, 2000). Our modern denudation-rate averages are ~0.3 to 0.4 mm/yr in the south and the fission-track derived rates of the southern SA are 0.1 to 1.6 mm/yr suggesting that locally erosion can almost keep pace with uplift in the southern SA.

Conclusions

Our denudation synthesis suggests one main conclusion for the Tertiary denudation history of the eastern Bolivian Andes: The northern portion has experienced greater rates of denudation than the south throughout the Holocene and perhaps since the late Miocene. Additional observations are that generally the short-term denudation-rate estimates exhibit the greatest variance, and denudation rates may be increasing to the present. Furthermore, knickpoints in longitudinal river profiles correlate with the thrust belt structure at the orogen and regional scale. In the south, large knickpoints at the EC-IA and IA-SA boundaries localize fluvial denudation, thereby potentially hindering basin-wide incision. In the north, more widespread, high gradient river slopes devoid of large knickpoints enhance fluvial denudation and relief. We conclude, given the combination of all existing geologic evidence to constrain the denudation and uplift history and observed tectonic-geomorphic correlations, that models of the evolution of the Bolivian Andes should incorporate a latitudinal erosion gradient for the last 10 kyrs to perhaps 10 Myrs.

The denudation synthesis presented here begs one question; to what greater precision can we extract denudation information from mountain morphology and geology spanning multiple spatial-temporal scales? The most obvious line of future research would be to produce a database of time-space denudation with multiple thermochronologic, cosmogenic, and isotopic techniques in an active orogenic setting where the structural and kinematic evolution is well-constrained, strath terraces are preserved to quantify incision rates, geomorphic features exist to provide regional constraints on the denudation, and comprehensive seismic, stratigraphic, climate, topography, river, lithologic, and sediment-flux data exist.

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Geobol [Servicio Geologico de Bolivia], 1992, Mapas Tematicos de Recursos Minerales de Bolivia, Tarija y Villazon (serie II-MTB-1B): Le Paz, Geobol, scale 1:250,000.

Geobol [Servicio Geologico de Bolivia], 1996a, Mapa Geologico de la Region de Vallegrande: Le Paz, Geobol, scale 1:250,000.

Geobol [Servicio Geologico de Bolivia], 1996b, Mapas Tematicos de Recursos Minerales de Bolivia, Sucre (serie II-MTB-8B): Le Paz, Geobol, scale 1:250,000.


Horton, B. K., 1999, Erosional control on the geometry and kinematics of thrust belt development in the central Andes: Tectonics, v. 18, p. 1292–1304.


Horton, B. K., and DeCelles, P. G., 1999, Erosional control on the geometry and kinematics of thrust belt development in the central Andes: Tectonics, v. 18, p. 1292–1304.


IGM (Instituto Geografico Militar) de Bolivia, 2000, Digital Atlas of Bolivia, CD-ROM.

Meybeck, M., 1979, Concentrations des eaux fluviales en éléments majeurs et apports en solution aux
Meigs, A., and Sauber, J., 2000, Southern Alaska as an example of the long-term consequences of mountain
McQuarrie, N., Horton, B. K., Zandt, G., Beck, S., and DeCelles, P. G., 2005, Lithospheric evolution of the
McQuarrie, N., and Davis, G. H., 2002, Crossing the several scales of strain-accomplishing mechanisms in the
—–2002b, The kinematic history of the central Andean fold-thrust belt, Bolivia: Implications for building
McQuarrie, N., ms, 2001, The making of a high-elevation plateau: Insights from the central Andean plateau,
McQuarrie, N., ms, 2001, The making of a high-elevation plateau: Insights from the central Andean plateau,
Leturmy, P., Mugnier, J. L., Baby, P., Colletta, B., and Chabron, E., 2000, Piggyback basin
development above a thin-skinned thrust belt with two detachment levels as a function of interactions
between tectonic and superficial mass transfer: The case of the Subandean zone (Bolivia): Tectonophys-
Masek, J. G., Isaaks, B. L., Gubbelts, T. L., and Fielding, E. J., 1994, Erosion and tectonics at the margins of
McQuarrie, N., ms, 2001, The making of a high-elevation plateau: Insights from the central Andean plateau,
—–2002a, Initial plate geometry, shortening variations, and evolution of the Bolivian orocline: Geology,
v. 30, p. 867–870.
—–2002b, The kinematic history of the central Andean fold-thrust belt, Bolivia: Implications for building
McQuarrie, N., and Davis, G. H., 2002, Crossing the several scales of strain-accomplishing mechanisms in the
hinterland of the central Andean fold-thrust belt, Bolivia: Journal of Structural Geology, v. 24,
p. 1587–1609.
McQuarrie, N., Horton, B. K., Zandt, G., Beck, S., and DeCelles, P. G., 2005, Lithospheric evolution of the
Andean fold-thrust belt, Bolivia, and the origin of the central Andean plateau: Tectonophysics, v. 399,
p. 15–37.
Meigs, A., and Sauber, J., 2000, Southern Alaska as an example of the long-term consequences of mountain
building under the influences of glaciers: Quaternary Science Reviews, v. 19, p. 1543–1562.
Meybeck, M., 1979, Concentrations des eaux fluviales en éléments majeurs et apports en solution aux
Mikkelsen, N., Maslin, M., Giraudau, J., and Showers, W., 1997, Biostatigraphy and sedimentation rates of
the Amazon fan, in Flood, R. D., Piper, D. J. W., Klaus, A., and Peterson, L. C., editors: Proceedings of the
Millet, J. R., 1990, The influence of bedrock geology on knickpoint development and channel-bed
Milliman, J. D., and Meade, R., 1983, World-wide delivery of sediment to the oceans: Journal of Geology,
v. 91, p. 1–21.
Milliman, J. D., and Syvitski, J. P. M., 1992, Geomorphic/tectonic control of sediment discharge to the ocean:
the importance of small mountain rivers: Journal of Geology, v. 100, p. 525–544.
Geology, v. 29, p. 579–582.
Pardo-Casas, F., and Molnar, P., 1987, Relative motion of the Nazca (Farallon) and South American plates
since Late Cretaceous time: Tectonics, v. 6, p. 233–248.
Safran, E. B., ms, 1998, Channel network incision and patterns of mountain geomorphology: Santa Barbara,


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