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The topographic imprint of a transient climate episode: the western Andean flank between 15.5° and 41.5°S

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Abstract

Mountain-range topography is determined by the complex interplay between tectonics and climate. However, often it is not clear to what extent climate forces topographic evolution and how past climatic episodes are reflected in relief.

The Andes are a tectonically active mountain belt encompassing various climatic zones with pronounced differences in rainfall and glacier extent under similar plate-boundary conditions. In the Central to South-Western Andes climatic zones range from hyperarid desert with mean annual rainfall of 5 mm/a (22.5°S) to year-round humidity with 2500 mm/a (40°S). The Andes thus provide a unique setting for investigating the relationship between climate and topography.

We present the analysis of 120 catchments along the western Andean watersheds between 15.5° and 41.5°S, which is based on SRTMV3-90m data and the new medium-resolution rainfall (TRMM) dataset. For each basin, we extracted geometry, relief, and climate parameters to test whether Andean topography shows a climatic imprint and to analyze how climate influences relief. Our data document that elevation and relief decrease with increasing rainfall and descending snowline elevation. Furthermore, we show that local relief reaches high values of 750 m in a zone between 28° to 35°S. During Pleistocene glacial stages this region was affected by the northward shifting southern hemisphere Westerlies, which provided moisture for valley-glacier formation and extended glacial coverage as well as glacial erosion. In contrast, the southern regions between 35° to 40°S receive higher rainfall and have a lower local relief of 200 m, probably related to an increased drainage density. We distinguish two different, climatically-controlled mechanisms shaping topography: (1) fluvial erosion by prolonged channel-hillslope coupling, which smoothes relief, and (2) erosion by valley glaciers that generates relief. Finally, our results suggests that the catchment-scale relief of the Andes between 28° to 35°S is characterized by a pronounced transient component reflecting past climatic conditions.

1. Introduction

Landscapes are sensitive recorders of climate-driven surface processes. The topographic signature of a landscape can thus provide important insights into the complex interaction between climatic and tectonic processes. Climate undoubtedly plays a major role in creating relief and shaping topography (Büdel, 1982; Molnar and England, 1990). Modeling studies suggest that increased precipitation might cause a reduction of mountain-range topography and strongly influence drainage-basin morphology (Tucker and Slingerland, 1997; Whipple and Tucker, 1999; Roe et al., 2003). Roe et al. (2003) suggest that in rainfall-dominated environments mountain ranges can be lowered by more than half due to enhanced rainfall. Also river-profile concavity (Roe et al., 2002; Zaprowski et al., 2005) and mountain-range width (Whipple and Meade, 2006) appear to be coupled to rainfall. It has been suggested that a reduction of topography and concavity could be controlled by an increase in drainage density, which might lower hillslope, tributary, and trunk channel relief, as well as threshold hillslope angles (Tucker and Bras, 1998; Whipple et al., 1999; Reiners et al., 2003; Gabet et al., 2004). These inferred feedbacks are generally corroborated by field studies (Montgomery et al., 2001; Gabet et al., 2004). For example, in the tectonically active Himalaya Gabet et al. (2004) document that a doubling of annual rainfall from 2000 to 4000 mm results in a 33 %-decrease in total relief.

Conversely, glacial erosion appears to be one of the main relief-generating processes focusing the removal of material near the peaks, and thereby maintaining steep topography above the equilibrium-line altitude (ELA) and limiting range height (Hallet et al., 1996; Brozovic et al., 1997; Small and Anderson, 1998; Whipple et al., 1999; Montgomery et al., 2001; Tomkin and Braun, 2002; Mitchell and Montgomery, 2006). However, the role of climate-induced relief production has been challenged and it has been argued that active mountain belts are already characterized by threshold relief (Whipple et al., 1999; Montgomery and Brandon, 2002). The controversies emphasize that fundamental key questions concerning the interactions between topography and climate have not been fully answered yet. In this context, it is important to investigate the influence of climate on mountain-range morphology and topography. In addition to modeling efforts these interactions should be studied in real orogens.

A good starting point to analyze the imprint of climate on topography constitutes the comparison of drainage basins, which are located in different climatic settings where all other dependent variables, such as lithology and general tectonic boundary conditions, are similar. Ideally, the overall characteristics of climate should have been sustained over longer time scales in a well constrained tectonic system. These requirements are reasonably fulfilled along the west coast of South America. Indeed, the Andes provide a unique setting for investigating the relationship between climate and topography. More than 7000 km long, the Andes are a meridionally oriented, tectonically active orogen spanning large climatic gradients from hyperarid to humid, controlled by stable, hemisphere-scale atmospheric circulation patterns that persisted at least during

Quaternary and possibly over the Neogene (Schwerdtfeger, 1976; Bice et al., 2000; Haselton et al., 2002; Hartley and Chong, 2002).

We conducted a detailed analysis of the western Andean flank between 15.5°S and 41.5°S. Using the new TRMM (Tropical Rainfall Measurement Mission) product 3B42 with 25 x 25 km spatial resolution and the topographic SRTMV3-90-m data, we extracted 19 geometry, relief, and climate parameters for 120 drainage basins. Therewith, we address the following questions: (1) How does rainfall affect the topography of drainage basins and why?, (2) What is the impact of glaciations on the topography of drainage basins?, and (3) Does climate also influence basin morphometry? Based on the potential interactions between climate and topography we expect to see a clear imprint of the different prevailing climatic zones on the relief of the western Andean flank. Additionally, landscapes might not only reflect prevailing climatic conditions, but could also show the imprint of past climatic episodes. Our data expands the rainfall range of other studies significantly (e.g. Zaprowski et al., 2005) and enables us to better understand the climatic controls of relief generation in mountain ranges, especially with regard to transience in landscapes and the role of rainfall on relief evolution.

2. Regional setting

2.1 Geologic setting

The Andean margin is characterized by the subduction of the Nazca plate beneath the South American continent at a rate of approximately 65 to 85 mm/a (Demets et al., 1994; Somoza, 1998; Angermann et al., 1999). The pre-Andean basement evolved during a period of already east-directed subduction (Pankhurst and Hervé, 2007). The basement rocks are subdivided into crustal segments, comprising exotic terranes, such as the Precordillera Terrane, as well as late Paleozoic to early Mesozoic igneous and metamorphic rocks, batholiths, and accreted marine sequences that constitute arc-forearc complexes (Hervé, 1988; Mpodozis and Ramos, 1989; Parada, 1990; Glodny et al., 2005; Hervé et al., 2007; Pankhurst and Hervé, 2007). This basement was strongly overprinted by subduction-related tectonic and magmatic processes, which have operated since the Jurassic period (Mpodozis and Ramos, 1989; Allmendinger et al., 1997).

The recent phase of contraction and shortening that initiated the uplift of the Andes was accompanied by extensive arc activity starting in the Eocene in the Central Andes; in the Northern Patagonian Andes these processes began in the early Miocene (Allmendinger et al., 1990; Jordan, 1993; Kley and Monaldi, 1998; Oncken et al., 2006). Total shortening decreases southward from ~ 300 km in the Central Andes to only ~ 15 km in the Northern Patagonian Andes (Isacks, 1988; Allmendinger et al., 1990; Diraison et al., 1998; Kley et al., 1999; Gregory-Wodzicki, 2000; Oncken et al., 2006; Vietor and Echtler, 2006) (Figure 1). GPS measurements corroborate this long-term pattern and document higher

shortening rates in the Central Andes (Dewey and Lamb, 1992; Klotz et al., 2001; Brooks et al., 2003). Based on shortening rates and tectonic activity, the Central to Southern Andes can be subdivided into four tectonic provinces (Figure 1). (1) The central part from 14° to 27°S comprises the Coastal Cordillera, the Central Depression, the Precordillera and the Main Cordillera with the internally drained intraorogenic plateaus of the Altiplano and Puna that display mean elevations of 4000 m. This sector displays pronounced crustal shortening in the Main Cordillera and the Subandean Ranges (Mpodozis and Ramos, 1989). (2) The region between 27° and 33°S is characterized by flat-slab subduction and a related absence of Quaternary volcanism and a Central Depression (Jordan et al., 1983; Mpodozis and Ramos, 1989; Kay and Mpodozis, 2002; Ramos et al., 2002). However, this region comprises the broken-foreland province of the Sierras Pampeanas that experiences active deformation and destructive earthquakes (Jordan and Allmendinger, 1986; Allmendinger et al., 1990). (3) South of 31°S, total shortening is significantly reduced and south of 37° shortening has stopped at the end of Miocene (Vietor and Echtler, 2006). South of 33°S the western onshore margin shows a pronounced morphotectonic segmentation integrating the forearc Coastal Cordillera, the Central Depression, and the Main Cordillera (Muñoz et al., 2000). (4) South of ~37°S the Main Cordillera is called Patagonian Cordillera. Here, deformation is partitioned in intra-arc strike-slip tectonics along the Liquiñe-Ofqui fault zone and thrust faulting (Cembrano et al., 2000; Thomson, 2002; Rosenau et al., 2006).

Surface-uplift rates in the Main Cordillera are poorly constrained and only few studies provide information (Figure 1) (Farias et al., 2005; Charrier et al., 2007; Schildgen et al., 2007). Therefore, we additionally consider Neogene uplift rates derived in the forearc along the coast. They vary between 0.2 and 2 mm/a and are relatively similar along the margin, except a patch of faster uplift in the region offshore the Arauco peninsula at ~37°S. This is a locally restricted forearc phenomenon and controlled by local high-angle reverse faults (Kaizuka et al., 1973; Radtke, 1989; Atwater et al., 1992; Nelson and Manley, 1992; Ortlieb et al., 1996a; Ortlieb et al., 1996b; Marquardt et al., 2004; Le Roux et al., 2005; Melnick et al., 2005; Pino and Navarro, 2005; Bookhagen et al., 2006; Le Roux et al., 2006; Melnick et al., 2009; Rehak et al., in revision) (Figure 1).

2.2 Climatic setting

From 15° to 42°S the western flank of the Andes spans various climatic zones and rainfall regimes from the hyperarid subtropical Atacama Desert with year-round aridity in the north and semiarid environments with few months of winter rain in the central part to year-round humid regions in the south (Figure 1). Mean annual precipitation increases from zero in the core of the Atacama desert at 22°S to about 3500 mm/a on the windward eastern side of the Andes at 41°S (e.g. Kummerow et al., 2000; New et al., 2002; TRMM, 2007; Bookhagen and Strecker, 2008). In the north the prevailing humid trade winds and the low-level Andean Jet from the east and northeast are blocked by the Andean topography. Together with the upwelling cold water of the Humboldt Current this promotes

hyperaridity on the western flanks in the north (Hartley and Chong, 2002). In contrast, the Westerlies directly deliver moisture to the western flanks south of $\sim 30^{\circ}\text{S}$. The equatorial position of the Pacific anticyclone enables these westerly frontal systems to penetrate northwards and carry moisture into the arid regions as far north as 30°S , occasionally during mid-winter even up to 27°S (e.g. Schwerdtfeger, 1976). Between 27° and 30°S this frontal rain occurs between May and August accounting for 90 % of the total annual rainfall with each month's precipitation normally originating from only one frontal passage (e.g. Schwerdtfeger, 1976). The largest precipitation gradient occurs between 30° and 35°S , where precipitation rates increase by one order of magnitude from 300 to 3000 mm/a.

Superposed on annual periodic rainfall variation, the region around 28°S is subjected to episodic variations from arid to semiarid conditions due to increased moisture transport from the south controlled by a northward shift of the Southern Westerlies during glacial periods (e.g. Scholl et al., 1970; Heusser, 1989; Clapperton, 1993; Veit, 1996; Lamy et al., 1998; Lamy et al., 2000; Jenny et al., 2002; Romero et al., 2006; Hebbeln et al., 2007). Less extensive shifts of the Westerlies have occurred on shorter timescales in the recent geologic past, i.e. in Holocene (Lamy et al., 1999). In contrast, precipitation patterns of the plateau regions north of 27°S appear to be relatively stable over glacial-interglacial cycles (Garreaud et al., 2003). Taken together, the present-day and past conditions define a broad climatic and geomorphic transition zone between 28° to 35°S , which is characterized by annual and glacial-interglacial rainfall variations.

The climatic zonation of the study area is controlled by long-lived, hemisphere-scale circulation patterns. The hyperarid conditions in the Atacama Desert were established at least 10 to 15 Ma ago, the arid conditions may have already been established during the Cretaceous, which promotes the extremely low geomorphic process rates in this region (Hartley, 2003; Dunai et al., 2005; Nishiizumi et al., 2005; Rech et al., 2006). The general circulation in the southern Central Andes has persisted during the Pleistocene and may have been launched in the middle Miocene, which implies that the humid conditions in the south have been stable and controlled by the Southern Westerlies over a very long time span (Schwerdtfeger, 1976; Bice et al., 2000; Haselton et al., 2002; Blisniuk et al., 2005). This is supported by the fact that the distribution of precipitation is mimicked by the thickness of sediment fill in the trench off South America, which document a significant increase of sediment thickness south of the Juan-Fernández Ridge and support the long-term stability of the general precipitation pattern along the western Andean margin (Bangs and Cande, 1997; Melnick and Echtler, 2006) (Figure 1).

South of 41°S drainage basins are not integrated and frequently change size due to glacial erosion of ice caps (Mardones, 2006). Consequently, these basins were not included in our analysis. Between about 18°S and 28°S recent glaciers do not exist and the elevation of the modern perennial snowline as well as the lowest Pleistocene glacier extent lie well above 5000 m, that is they exceed the maximum elevation of the drainage basins (Rabassa and Clapperton, 1990;

Ammann et al., 2001). Hence, these catchments have not been influenced by Pleistocene glaciations.

3. Methodology

Along the western Andean flank from 15.5°S near Arequipa (Peru) to 41.5°S to Chiloé Island (Chile) we extracted a total of 120 drainage basins, comprising 64 main catchments draining into the Pacific and 56 equally-sized sub-catchments with a mean drainage area of 1000 km². According to their location, the main catchments were grouped into forearc, Andean mountain front, and arc catchments (Figure 2). This enables us to detect scale effects and to distinguish between basins shaped by different geomorphic processes. Forearc catchments drain only the Coastal Cordillera and have not been affected by Quaternary glaciations. In contrast, mountain-front catchments also include the Central Depression and the Andean foothills, nevertheless they do not extend into the high Andes and have mostly not been glaciated during the Quaternary. Arc catchments extend from the Pacific to the principal Andean watershed. Sub-catchments are parts of the arc catchments and constitute the uppermost headwater basins along the Andean main drainage divide. All sub-catchments were extracted based on a similar contributing area of ~ 500 to 2000 km² in order to identify and exclude orographic and scale effects. For each basin and its trunk stream we calculated 19 variables describing geometry, relief, and climate. Our analysis is based on the new medium-resolution (25 x 25 km) rainfall dataset 3B42 TRMM averaged over 8 years from 1998 to 2006 (Kummerow et al., 2000; TRMM, 2007) and topographic SRTMV3-90-m data (USGS, 2007).

The variables are defined and summarized in Table 1. All variables were correlated in order to identify relationships between basin geometry, range relief, climate, river-profile concavity and hypsometric integral. Total basin relief is calculated as the difference between the absolute maximum and the absolute minimum elevation of a catchment (Figure 2). In contrast, local relief describes the relief range in circles with a certain radius averaged for each basin. We apply a radius of 4.5 km, which assures the incorporation of the characteristic bandwidth of local relief in our region (e.g. Champagnac et al., 2007; Purves and Korup, 2007) (Figure 2). Modern annual snowline and lowest Pleistocene glacier extent are compiled from Schwerdtfeger (1976) and Rabassa and Clapperton (1990).

River profiles are often interpreted with respect to climatic conditions, fluvial processes, rock-uplift rates, and evolutionary stage (e.g. Strahler, 1952; Summerfield, 1991; Ohmori, 1993; Whipple and Tucker, 1999; Snyder et al., 2000; Kirby and Whipple, 2001). The concavity of a river-profile can be expressed as a profile-concavity index (PCI) (e.g. Demoulin, (1998). This reduces the river profile to a single number and enables correlation with other variables. The PCI is an indicator for the general channel topography and calculated as the normalized area under the normalized distance-elevation curve of the river (Figure 2). A diagonal drawn from the source to the mouth would

result in a value of 0.5; thus, profiles with a PCI > 0.5 reflect convex shapes. The PCI is similar to stream-profile concavity used by Zaprowski et al. (2005), who found that this index covaries with θ , the regression slope in log slope – log area plots. These authors hence concluded that both values are effective measures of profile concavity. We do not identify rainfall variations along single longitudinal-river profiles in this study.

The hypsometric integral (HI) describes the distribution of elevations in a drainage basin and thus basin topography. The HI is calculated as the normalized area under a normalized area-elevation curve, following Strahler (1952) (Figure 2). It is interpreted as a reflection of the geomorphic evolutionary stage and the degree of tectonic activity affecting landscapes (Strahler, 1952; Summerfield, 1991; Ohmori, 1993). Mature landscapes without significant tectonic activity are characterized by low relief contrasts and hence low integrals, whereas young, actively uplifting and deeply incised areas display high integrals (e.g. Strahler, 1952). However, basins showing the same HI are probably much older in arid environments than in humid climates due to differences in rainfall and denudation rates. In contrast, two basins, which started to uplift at the same time and are situated in different climatic zones should exhibit significantly distinct HI-values. Hence, HI-values might not only reflect the age of landscapes, but also erosional processes.

The variables were correlated separately for all basin groups. In our interpretation, we will mainly focus on the key parameters relief, elevation, rainfall, and snowline that we identified to elucidate the interactions between climate and topography. The significance level for all correlations was set to the 95 % confidence interval of a two tailed t-test. As a goodness-of-fit measure for the correlations and regressions we used the constant of regression (R^2).

Lithologies were determined by a geological map (Servicio Nacional de Geología y Minería, 2003). They vary non-systematically along the active continental margin between 15° and 40°S and do not correlate with climatic zonation. Data on surface-uplift rates in the Main Cordillera is very sparse, thus, we have to rely on proxy indicators such as surface-uplift data from the forearc, the intensity of active deformation, and trench-normal shortening.

4. Results and interpretation

The analysis of 120 drainage basins along the Central and Southern Andes shows evidence for a prominent imprint of climate on topography. Landsat TM images of typical headwater basins reveal the distinct character of local catchment relief in the different climatic zones (Figure 3). The northern basins in the hyperarid part show a very low local relief. This landscape is virtually undissected and characterized by extremely low erosion rates < 0.1 m/Ma (Nishiizumi et al., 2005) resulting in the preservation of mid-Miocene relict surfaces (Rech et al., 2006) (Figure 3A). Only those regions traversed by rivers originating in the high volcanic arc have deeply incised canyons (Zeilinger et al., 2005; Hoke et al., 2007; Schildgen et al., 2007). In contrast, headwaters in the

climatic transition zone between 28° to 35°S are highly dissected. These sub-catchments have steep hillslopes, sharp ridges, and record a glacial overprint (Figure 3B). The catchments in the humid south between 35° and 40°S display a smooth local relief with moderate hillslopes and subrounded ridges (Figure 3C). In contrast, south of 40°S the landscape has steeper hillslopes and a higher degree of dissection (Figure 3D). These observations are clearly reflected in our morphometric data suggesting a strong influence of rainfall and snowline elevation on topography along the western flank of the Andean margin (Figures 4 and 5, Tables DR1, DR2 and DR3 in Data Repository).

In the following sections we (1) introduce the influence of rainfall on the topography of drainage basins, (2) address the impact of glaciations on basin topography, and (3) describe the effect of climate on basin morphometry.

4.1 The influence of rainfall on basin topography

South of 35°S local relief, total catchment relief, and basin mean elevations significantly decrease by 26, 27 and 70 % (Figure 5). This decrease correlates with a considerable increase of rainfall in this region. Generally, a decrease of elevations towards the south could also be related to the descending snowline and a glacial “buzz-saw effect” (Brozovic et al., 1997). However, most parts of the mountain belt in this region are presently not glaciated and the decrease of local relief starts in all catchments when rainfall amounts exceed ~ 800 mm/a (Figure 5). Moreover, the relationships between rainfall and relief parameters can also be observed in forearc and mountain-front basins that have not been influenced by glacial erosion during the Quaternary. Hence, the decrease in elevation appears in fact to be related to increasing rainfall. Essentially, local relief shows very low values where rainfall amounts are highest. These observations indicate that 800 mm/a might be a critical threshold-rainfall amount for these lithologies above which the fluvial system is coupled to the hillslopes and efficiently erodes these environments. We therefore suggest that the low mountain-range relief in the zone between 35° to 40°S is caused by efficient, protracted fluvial erosion characterized by prolonged channel incision and the coupling of channels and hillslopes.

4.2 The glacial imprint on basin topography

Our data shows negative correlations between snowline as well as Pleistocene glacier extent and basin mean elevation as well as total relief (Figures 4G and 4H, Tables DR1 and DR2 in Data Repository). Between ~18° to 28°S the modern snowline and the lowest extent of Pleistocene glaciation lie well above the maximum elevations of the ranges (Figure 5). In contrast, basins south of ~ 27°S (arc catchments), ~ 28°S (sub-catchments), and ~ 32°S (mountain-front catchments) have been at least partially glaciated during the Pleistocene (Figure 5). Most interestingly, in the region between 28° and 35°S, where mean

annual rainfall amounts are low, local relief displays a prominent maximum (Figure 5). In fact, once the maximum elevation is undercut by the snowline, basin sectors above this line display a progressive increase in local relief, hence the topography above the snowline gets progressively steeper (Figure 4D). We therefore suggest that the high local relief between 28° and 35°S is the result of repeated glacial erosion during Quaternary times. The steep topography and high local relief can also be observed in catchment parts that are recently not glaciated (Figures 3 and 5). Interestingly, this implies that between 28° and 35°S the present relief shows a clear imprint of past climatic episodes. The preservation might have been promoted by the present-day low rainfall amounts and inefficient erosion.

Both, modern and Pleistocene ELA start to descend southward when rainfall amounts increase along the western Andean flank (Figures 4E and 4F), which is dominated by westerly sourced moisture south of 28°S (Clapperton, 1994; Haselton et al., 2002). Thus, northward-migrating (South) Westerlies may have reached these latitudes, intensifying precipitation during glacial and delivering sufficient moisture for glaciation in these threshold environments (e.g. Heusser, 1989; Lamy et al., 1999). This supports the notion that glaciation in large parts of the central Andes is moisture-controlled, which is in good agreement with results from Haselton et al. (2002) who concluded that the Pleistocene snowline in the southern Altiplano-Puna Plateau is more sensitive to moisture increase than temperature.

4.3 Relations between climate and basin morphometry

Variables describing the geometric shape of drainage basins do not show correlations with climatic parameters (Table DR 1 in Data Repository). In contrast, the vertical shape of basins appears to be strongly affected by climate. PCI and HI are reduced with enhanced rainfall (Table DR3 in Data Repository). The inverse relationships between rainfall and PCI as well as rainfall and HI probably reflect the more efficient erosion at higher rainfall rates, where rivers and hillslopes adjust faster to changes in boundary conditions and approach equilibrium. Our data thereby supports the conclusions of studies, which suggested that sparse precipitation correlates with convex profiles (high PCI), whereas concave profiles (low PCI) develop in regions with intense rainfall (Roe et al., 2002; Zaprowski et al., 2005). HI scales with mean elevation and rainfall (Table DR3 in Data Repository). This corroborates the idea that the HI actually reflects the efficiency of erosion processes, not necessarily the age of landscapes. Hence, in synoptic studies of mountain belts in similar climatic settings HI-values should rather be used relative to each other as proxies for tectonic activity as indicated by results of Montgomery et al. (2001).

5. Discussion

Our results are in agreement with other empirical and modeling studies, which propose that high precipitation rates lower elevation and relief (Montgomery et al., 2001; Roe et al., 2003; Gabet et al., 2004). Roe et al. (2003) suggest that in rainfall-dominated environments a mountain range can be lowered by more than half due to enhanced precipitation. In the tectonically active Himalaya, Gabet et al. (2004) document that a doubling of annual rainfall from 2000 to 4000 mm results in a 33 %-decrease in total relief. Despite significantly lower rainfall amounts in the study area, our results document a decrease in mean elevation and total relief on the order of 44 % and 28 %, respectively, for a doubling of rainfall amounts (Table 2). Such a rainfall-induced decrease in elevation and thus enhanced erosivity might be controlled by an increase in drainage density (e.g. Tucker and Bras, 1998; Whipple et al., 1999). In fact, enhanced rainfall is associated with a progressive southward increase of drainage density (Figure 6). High drainage density causes more effective denudation of higher catchment parts that are not integrated in the fluvial network when drainage density is low (Figure 6). As a result, in fluvially dominated areas total and local relief decrease with increasing drainage density and increasing rainfall.

Consequently, we distinguish two geomorphic sectors on the western flank of the Central to Southern Andes. Between 35° and 40°S fluvial erosion, enhanced by high protracted rainfall amounts, appears to be prolonged and very efficient. As a result, local relief is diminished (Figure 5). In contrast, between 28° to 35°S rainfall amounts are below a critical threshold, rendering fluvial erosion ineffective. Based on our data we conclude that the pronounced high local relief has been generated by glacial erosion during the Quaternary. This implies that the topography we observe today is a relict from past climatic episodes. Hence, the region between 28° and 35°S appears to be a transient landscape, which does not reflect the presently prevailing geomorphic processes.

Clearly, erosion and topography in mountain belts are also determined by tectonic processes. Due to the lack of uplift data in the Main Cordillera, we applied the shortening rate as a proxy for potential surface deformation. This complicates the comparison of tectonic processes. However, changes in shortening rate imply basic changes in deformation processes and can thus be regarded as tectonic boundary conditions. In principle, the Andean shortening rate decreases southward, absolute shortening is reduced by more than 30 % from high shortening (> 300 km) to low shortening (< 100 km) at around 30°S (e.g. Allmendinger et al., 1990; Vietor and Echtler, 2006) (Figure 1). We observe the local relief maximum at 28° to 35°S, and thereby already in a sector where shortening is low. This is the opposite of what might be expected if shortening would be the controlling factor for enhanced local relief. In addition, the flat-slab segment does not coincide with the low-relief topography between 28° and 35°S. Consequently, the sectors defined by high local relief and dominant geomorphic process are not consistent with tectonic segments. Finally, the high-relief sector does not correlate with a distinct rock type characterized by low erodibility, but mainly comprises Cretaceous to Miocene volcano-sedimentary and volcanic rocks, which are abundant in large parts of the Andes (Servicio Nacional de Geología y Minería, 2003). A lithologic control on general morphologic and

topographic trends can thus be excluded as well. In summary, neither changes in shortening rate, subduction properties, and lithology nor the spatial distribution of deformation can solely account for the high local relief sector. This suggests that local relief along the Andean margin is mainly the result of surface processes determined by present and past climatic conditions. Especially the zone between 28° and 35°S appears to show a prominent imprint of transient climate episodes.

6. Conclusion

We analyzed climatic and topographic parameters of 120 drainage basins along the western Andean flank between 15.5° and 41.5°S based on SRTMV3-90m topographic data and the new medium spatial resolution satellite-derived Tropical Rainfall Measurement Mission (TRMM) dataset. In order to diagnose the imprint of different climatic zones on the relief of mountain ranges, we correlated climatic with relief parameters and investigated the impact of different geomorphic regimes on relief evolution.

Our results suggest that (1) increased rainfall diminishes mean and maximum elevations of mountain ranges and provide a compelling support for the idea that drainage density is a key parameter in landscape evolution. In fact, the observed decrease of elevations appears to be caused by the high drainage density in humid regions, which enables the integration of the entire catchment in the fluvial network and thereby the efficient coupling of channels and hillslopes. Additionally, our data indicates that (2) glacial erosion increases basin relief and limits total range topography. This might be caused by focusing erosion on summits and ridges above the Equilibrium Line Altitude (ELA). Eventually, our analyses show that (3) the profile-concavity Index (PCI) and the hypsometric integral (HI) strongly depend on rainfall, while the geometric basin shape appears not to be affected by climate.

Consequently, our study suggests that the generation of local relief is a force balance between protracted fluvial erosion and glacial erosion. In this respect, the two processes act as opposite agents. Whereas glacial erosion appears to create local relief, persistent moderate rainfall above a critical threshold appears to smooth it.

Overall, we conclude that the catchment-scale relief of the Andean mountain chain distinctly reflects the dominant geomorphic process, which is determined by past and present regional climatic regimes. Notwithstanding, we do not challenge that the large-scale architecture of the Andes is maintained by tectonic processes and the reactivation of inherited structures. The geomorphic signature of the Western Andes between 28° and 35°S, however, expresses significant transient components, which may reflect erosion processes under different climatic conditions during the geologic past. In general, basin topography could thus be used to identify the efficiency of past paleoclimatic conditions and geomorphic processes.

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Tables

Variable	Unit	Definition	Source
Vertical shape			
profile concavity		normalized area under the normalized distance-elevation profile (diagonal from source to mouth would be 0.5) $PCI > 0.5$ reflect a strongly convex profile	after Demoulin, 1998
hypsometric integral		normalized area under a normalized area-elevation curve after Strahler	after Strahler, 1952
Geometry			
basin area	km ²	area of the drainage basin	
basin length	km	longest distance from outlet	
basin width	km	longest distance perpendicular to basin length	
basin form		basin length/basin width	
basin elongation ratio		diameter of a circle with basin area/basin length	after Schumm, 1956
Relief			
basin mean elevation	m	average elevation calculated within a drainage basin	
total basin relief	m	basin maximum elevation - basin minimum elevation	
basin relief ratio		$\text{relief/length} * 100$	
basin local relief	m	local relief averaged over a circle with a diameter of 10 cells	after Summerfield, 1991
channel drop	m/m	(channel maximum elevation - channel minimum elevation)/channel length	
Climate			
mean annual rainfall	mm/a		TRMM 30km dataset
mean annual temperature	°C		cru-dataset, 10' grid (http://www.cru.uea.ac.uk/cru/data)
mean annual ground frost	d/a		cru-dataset, 10' grid (http://www.cru.uea.ac.uk/cru/data)
modern perennial snowline	m		after Schwerdtfeger, 1976; Rabassa and Clapperton, 1990
lowest Pleistocene glacier extent	m		after Schwerdtfeger, 1976; Rabassa and Clapperton, 1990
Erosion			
erosion index		$e = Q^{0.5} * S$, with Q = mean annual rainfall, S = channel drop	after Finlayson et al., 2002
erosion intensity	km ³ /a	$IE = P * A * S$, with P = mean annual rainfall, A = drainage area	after Montgomery et al., 2001

Table 1: Definition and sources of the calculated variables. We mainly concentrate on basin mean elevation, basin total and local relief, mean annual rainfall, and snowline as well as glacier extent.

Mean annual rainfall [mm/a]	Δ Mean elevation [%]	Δ Total relief [%]	Position
50 - 100	118	8	Forearc
	18	-20	Mountain Front
	0	3	Arc
	44	-37	Subbasins
	45	-12	Average
100 - 200	-24	-38	Forearc
	-13	28	Mountain Front
	28	-9	Arc
	0	7	Subbasins
	-2	-3	Average
200 - 400	55	29	Forearc
	-33	-17	Mountain Front
	-50	4	Arc
	-34	91	Subbasins
	-16	27	Average
400 - 800	-76	-31	Forearc
	-85	-40	Mountain Front
	-34	-30	Arc
	-26	-27	Subbasins
	-55	-32	Average
800 - 1600	<-41	<-48	Forearc
	---	---	Mountain Front
	<-30	<-14	Arc
	-29	-10	Subbasins
	-33	-24	Average
>400	-44	-28	Average

Table 2: Relief reduction due to a doubling of rainfall. See the amount of rainfall and the respective change in mean elevation and total relief for each basin group. Note that mean elevation and total relief are only significantly reduced by doubling rainfall amounts above rainfall rates of 400 mm/a.

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Forearc	basin_area	basin_length	basin_width	basin_form	basin_elongation_ratio	basin_mean_elevation	total_basin_relief	basin_relief_ratio	basin_local_relief	channel_drop	mean_annual_rainfall	mean_annual_temperature	ground_frost	modern_perennial_snowline	lowest_Pleistocene_glaciers	erosion_index
basin_area	1.000															
basin_length	0.888	1.000														
basin_width	0.780	0.446	1.000													
basin_form	0.029	0.446	-0.550	1.000												
basin_elongation_ratio	0.902	0.043	0.010	-0.889	1.000											
basin_mean_elevation	0.728	0.085	0.004	0.000	-0.199	1.000										
total_basin_relief	0.271	0.439	0.115	0.282	0.216	0.395	1.000									
basin_relief_ratio	0.894	0.784	0.382	0.382	0.166	0.817	0.888	1.000								
basin_local_relief	0.029	0.151	0.000	0.000	0.000	0.000	0.000	0.000	1.000							
channel_drop	-0.223	-0.192	-0.074	-0.104	0.166	0.888	0.417	0.000	0.000	1.000						
mean_annual_rainfall	-0.122	-0.064	-0.327	0.300	-0.204	-0.190	-0.039	-0.048	1.000							
mean_annual_temperature	0.598	0.784	0.146	0.186	0.376	0.410	0.702	0.835	-0.075	1.000						
ground_frost	-0.133	-0.027	-0.142	0.106	-0.044	0.784	0.607	0.888	0.000	0.000	1.000					
modern_perennial_snowline	0.564	0.909	0.540	0.648	0.649	0.000	0.019	0.000	0.000	0.000	0.000	1.000				
lowest_Pleistocene_glaciers	-0.609	-0.635	-0.242	-0.405	0.198	-0.738	-0.776	-0.363	0.219	-0.438	0.000	-0.721	1.000			
erosion_index	0.019	0.002	0.290	0.068	0.391	0.000	0.000	0.108	0.341	0.047	0.000	0.000	0.464	0.000	-0.133	1.000
erosion_intensity	0.833	0.884	0.342	0.232	-0.231	0.847	0.889	0.148	-0.455	0.227	-0.721	1.000				
	0.016	0.015	0.128	0.426	0.665	0.282	0.696	0.002	0.574	0.000	0.275	0.145	0.058	0.403	0.567	0.372
	-0.281	-0.242	-0.138	-0.175	0.064	0.121	0.005	0.301	0.168	0.212	0.360	-0.375	1.000	-0.039	-0.052	0.372
	0.217	0.292	0.557	0.449	0.176	0.600	0.982	0.185	0.468	0.351	0.073	0.094	0.000	0.866	0.823	0.097
	0.616	0.681	0.273	0.353	-0.240	0.843	0.828	0.363	-0.352	0.468	-0.860	0.912	-0.219	1.000		
	0.017	0.001	0.231	0.116	0.294	0.000	0.000	0.105	0.118	0.033	0.000	0.000	0.340	0.993	1.000	
	0.443	0.646	0.192	0.388	-0.262	0.865	0.840	0.419	-0.311	0.519	-0.846	0.978	-0.169	-0.193	-0.133	1.000
	0.044	0.002	0.404	0.092	0.216	0.000	0.000	0.058	0.170	0.016	0.000	0.000	0.464	0.000	0.567	0.372
	-0.621	-0.623	-0.343	-0.183	0.100	0.246	-0.091	0.635	0.130	0.727	0.250	-0.329	0.421	-0.193	-0.133	1.000
	0.016	0.015	0.128	0.426	0.665	0.282	0.696	0.002	0.574	0.000	0.275	0.145	0.058	0.403	0.567	0.372
	0.252	0.086	0.476	-0.423	0.320	0.089	0.141	0.085	-0.058	0.151	0.375	-0.020	0.358	-0.039	-0.052	0.372
	0.271	0.678	0.029	0.056	0.157	0.702	0.541	0.714	0.602	0.514	0.175	0.931	0.112	0.866	0.823	0.097
bold: $r > 0.5$ shaded: $r^2 > 0.5$																
Mountain Front	basin_area	basin_length	basin_width	basin_form	basin_elongation_ratio	basin_mean_elevation	total_basin_relief	basin_relief_ratio	basin_local_relief	channel_drop	mean_annual_rainfall	mean_annual_temperature	ground_frost	modern_perennial_snowline	lowest_Pleistocene_glaciers	erosion_index
basin_area	1.000															
basin_length	0.616	1.000														
basin_width	0.884	0.622	1.000													
basin_form	-0.609	-0.635	-0.784	1.000												
basin_elongation_ratio	0.888	0.358	0.889	-0.812	1.000											
basin_mean_elevation	-0.074	-0.075	-0.135	0.342	-0.179	1.000										
total_basin_relief	0.185	0.187	0.082	0.282	0.074	0.824	1.000									
basin_relief_ratio	-0.300	-0.706	-0.364	0.431	-0.270	0.872	0.836	1.000								
basin_local_relief	0.187	0.000	0.104	0.051	0.236	0.001	0.000	0.000	1.000							
channel_drop	0.653	0.601	0.045	0.185	0.096	0.871	0.890	0.002	0.000	1.000						
mean_annual_rainfall	-0.651	-0.635	-0.695	0.640	-0.679	0.689	0.478	0.896	0.808	1.000						
mean_annual_temperature	0.117	0.168	0.189	-0.328	0.164	-0.847	-0.811	-0.710	-0.285	-0.743	1.000					
ground_frost	0.018	-0.222	-0.039	0.300	-0.062	0.384	0.371	0.436	0.319	0.692	-0.456	1.000				
modern_perennial_snowline	0.668	0.549	0.589	0.283	-0.150	0.883	0.851	0.256	0.147	0.286	-0.421	-0.304	1.000			
lowest_Pleistocene_glaciers	0.059	0.045	0.002	0.220	-0.035	0.871	0.890	0.002	0.000	0.209	0.057	0.180	0.465	1.000		
erosion_index	0.131	0.120	0.098	0.113	0.338	0.843	0.826	0.609	0.174	0.662	-0.804	0.214	0.443	0.889	1.000	
erosion_intensity	-0.437	-0.617	-0.457	0.443	-0.458	0.718	0.213	0.674	0.808	0.789	-0.412	0.520	0.241	0.430	0.355	1.000
	0.048	0.003	0.037	0.044	0.037	0.028	0.154	-0.225	-0.272	-0.395	0.064	0.181	-0.051	0.052	0.114	-0.163
	0.887	0.436	0.909	-0.614	0.849	0.025	0.184	-0.225	-0.272	-0.395	0.064	0.181	-0.051	0.052	0.114	-0.163
	0.000	0.048	0.000	0.003	0.000	0.914	0.388	0.326	0.232	0.076	0.685	0.434	0.825	0.432	0.299	0.479
bold: $r > 0.5$ shaded: $r^2 > 0.5$																

Table DR1: Correlations between variables: forearc catchments and mountain front catchments.

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Table DR 2: Correlations between variables: arc catchments and subcatchments.

Atc	basin_area	basin_length	basin_width	basin_form	basin_elongation_ratio	basin_mean_elevation	total_basin_relief	basin_relief_ratio	basin_local_relief	channel_drop	mean_annual_rainfall	mean_annual_temperature	ground_frost	modern_perennial_snowline	lowest_Pleistocene_glaciers	erosion_index
basin_area	1.000															
basin_length	0.664	1.000														
basin_width	0.814	0.432	1.000													
basin_form	-0.437	0.126	-0.740	1.000												
basin_elongation_ratio	0.542	-0.237	0.591	-0.762	1.000											
basin_mean_elevation	-0.190	-0.071	-0.111	0.127	-0.153	1.000										
total_basin_relief	0.009	-0.168	0.111	-0.129	0.496	0.816	1.000									
basin_relief_ratio	0.859	0.498	0.523	0.449	0.819	0.000	0.734	1.000								
basin_local_relief	0.037	-0.142	0.142	0.120	0.412	0.006	0.734	1.000								
channel_drop	-0.136	-0.335	0.061	-0.369	0.265	0.006	0.526	0.687	1.000							
mean_annual_rainfall	-0.412	-0.560	-0.328	0.125	-0.016	0.706	0.776	0.829	0.448	1.000						
mean_annual_temperature	0.252	0.316	0.107	-0.044	-0.016	-0.860	-0.855	-0.716	-0.542	-0.850	1.000					
ground_frost	-0.002	-0.052	-0.127	0.265	-0.114	-0.599	-0.458	-0.308	-0.440	-0.171	0.386	1.000				
modern_perennial_snowline	0.961	0.818	0.573	0.233	0.613	0.003	0.032	0.163	0.040	0.445	0.076	0.804	1.000			
lowest_Pleistocene_glacier_extert	0.087	0.023	-0.062	0.038	-0.095	0.821	0.742	0.451	0.357	0.546	-0.716	-0.804	0.765	1.000		
erosion_index	-0.216	-0.163	-0.094	0.093	-0.068	0.829	0.869	0.648	0.423	0.795	-0.861	-0.442	0.766	1.000		
erosion_intensity	0.333	0.468	0.676	0.681	0.704	0.000	0.001	0.001	0.051	0.000	0.039	0.000	0.868	1.000		
erosion_index	-0.208	-0.131	-0.058	0.078	-0.108	0.813	0.829	0.596	0.367	0.737	-0.837	-0.438	0.765	0.868	1.000	
erosion_index	0.352	0.562	0.799	0.730	0.632	0.000	0.003	0.003	0.075	0.636	-0.224	0.079	0.105	0.103	-0.012	1.000
erosion_index	-0.080	-0.433	-0.100	-0.121	0.187	0.098	0.301	0.442	0.373	0.536	0.010	0.316	0.727	0.643	0.957	
erosion_index	0.724	0.044	0.659	0.591	0.406	0.064	0.173	0.039	0.087	0.010	0.316	0.727	0.643	0.957	0.372	1.000
erosion_index	0.804	0.336	0.611	-0.363	0.511	-0.440	-0.156	-0.307	-0.139	-0.323	0.262	-0.351	-0.471	-0.471	-0.513	0.372
erosion_index	0.000	0.126	0.002	0.079	0.015	0.040	-0.489	0.165	0.538	0.142	0.046	0.239	0.110	0.027	0.015	0.688
bold: $r > 0.5$, shaded: $r > 0.5$																
Subbasins	basin_area	basin_length	basin_width	basin_form	basin_elongation_ratio	basin_mean_elevation	total_basin_relief	basin_relief_ratio	basin_local_relief	channel_drop	mean_annual_rainfall	mean_annual_temperature	ground_frost	modern_perennial_snowline	lowest_Pleistocene_glaciers	erosion_index
basin_area	1.000															
basin_length	0.387	1.000														
basin_width	0.714	-0.096	1.000													
basin_form	-0.408	0.446	-0.818	1.000												
basin_elongation_ratio	0.453	-0.592	0.642	-0.795	1.000											
basin_mean_elevation	-0.456	-0.393	-0.149	-0.011	-0.042	1.000										
total_basin_relief	0.016	0.112	0.040	0.099	-0.079	-0.044	1.000									
basin_relief_ratio	0.908	0.409	0.768	0.468	0.565	0.750	0.651	1.000								
basin_local_relief	-0.208	-0.890	0.055	-0.278	0.408	0.147	0.651	1.000								
channel_drop	0.080	0.000	0.685	0.038	0.002	0.280	0.000	0.485	1.000							
mean_annual_rainfall	0.659	0.956	0.889	0.920	0.451	0.003	0.000	0.000								
mean_annual_temperature	-0.592	0.006	-0.683	0.532	-0.551	0.372	0.413	0.225	0.005	1.000						
ground_frost	0.000	0.956	0.000	0.000	0.000	0.005	0.002	0.095	0.973	-0.551	1.000					
modern_perennial_snowline	0.519	0.144	0.253	-0.181	0.273	-0.829	-0.111	-0.090	0.411	-0.551	0.000					
lowest_Pleistocene_glacier_extert	0.000	0.290	0.060	0.182	0.042	0.000	0.415	0.000	0.000	0.000	1.000					
erosion_index	0.015	0.008	0.073	0.046	-0.074	-0.825	-0.029	-0.223	-0.197	-0.135	0.392					
erosion_index	-0.344	-0.405	-0.058	-0.075	0.115	0.748	0.016	0.223	0.006	0.165	-0.516	-0.834	1.000			
erosion_index	0.009	0.002	0.672	0.583	0.400	0.910	0.098	0.967	0.967	0.224	0.000					
modern_perennial_snowline	-0.460	-0.215	-0.180	0.114	-0.183	0.803	0.343	0.066	-0.533	0.465	-0.815	-0.327	0.514	1.000		
lowest_Pleistocene_glacier_extert	-0.438	-0.202	-0.184	0.105	-0.189	0.879	0.343	0.066	-0.533	0.465	-0.815	-0.327	0.514	0.868	1.000	
erosion_index	0.001	0.135	0.175	0.436	0.163	0.879	0.343	0.066	-0.533	0.465	-0.815	-0.327	0.514	0.868	1.000	
erosion_index	-0.284	0.028	-0.360	0.334	-0.290	-0.182	0.303	0.238	0.460	0.575	0.196	-0.127	0.061	-0.269	-0.328	1.000
erosion_index	0.034	0.838	0.003	0.012	0.030	0.180	0.023	0.078	0.000	0.147	0.352	0.655	0.031	0.014	0.014	0.423
erosion_index	0.579	0.274	0.246	-0.094	0.209	-0.714	0.059	-0.068	0.368	-0.250	0.819	0.320	-0.420	-0.778	-0.781	0.423
erosion_index	0.000	0.041	0.068	0.462	0.122	0.000	0.665	0.618	0.002	0.030	0.000	0.016	0.001	0.000	0.000	0.001
bold: $r > 0.5$, shaded: $r > 0.5$																

	Forearc		Mountain Front		Arc		Subbasins	
	HI	PCI	HI	PCI	HI	PCI	HI	PCI
basin_area	-0.301 0.185	0.042 0.857	-0.186 0.419	0.225 0.326	-0.267 0.230	-0.281 0.205	-0.469 0.000	-0.268 0.046
basin_length	-0.189 0.411	0.157 0.498	-0.272 0.233	0.088 0.706	-0.060 0.790	-0.014 0.952	-0.167 0.218	-0.141 0.299
basin_width	-0.300 0.187	0.005 0.983	-0.216 0.348	0.171 0.459	-0.200 0.372	-0.379 0.082	-0.255 0.058	-0.087 0.524
basin_form	0.155 0.502	0.168 0.467	0.282 0.215	0.031 0.895	0.228 0.307	0.454 0.034	0.151 0.267	0.100 0.463
basin_elongation_ratio	-0.113 0.624	-0.121 0.601	-0.258 0.260	0.126 0.585	-0.273 0.219	-0.318 0.149	-0.236 0.079	-0.124 0.362
basin_mean_elevation	0.731 0.000	0.828 0.000	0.928 0.000	0.573 0.007	0.980 0.000	0.817 0.000	0.496 0.000	0.740 0.000
total_basin_relief	0.245 0.285	0.475 0.030	0.590 0.005	0.399 0.073	0.717 0.000	0.492 0.020	0.175 0.197	-0.199 0.142
basin_relief_ratio	0.719 0.000	0.583 0.006	0.683 0.001	0.214 0.352	0.503 0.017	0.364 0.096	0.165 0.224	-0.142 0.297
basin_local_relief	-0.052 0.823	-0.383 0.086	0.379 0.090	-0.355 0.115	0.323 0.143	0.003 0.989	-0.029 0.835	-0.508 0.000
channel_drop	0.762 0.000	0.587 0.005	0.756 0.000	0.165 0.474	0.694 0.000	0.603 0.003	0.522 0.000	0.231 0.086
mean_annual_rainfall	-0.393 0.078	-0.595 0.004	-0.798 0.000	-0.366 0.103	-0.822 0.000	-0.677 0.001	-0.473 0.000	-0.572 0.000
mean_annual_temperature	0.302 0.184	0.671 0.001	0.426 0.054	-0.103 0.657	-0.534 0.010	-0.382 0.079	-0.284 0.034	-0.411 0.002
ground_frost	0.118 0.612	0.015 0.949	0.462 0.035	0.363 0.105	0.883 0.000	0.740 0.000	0.266 0.048	0.517 0.000
modern_perennial_snowline	0.445 0.043	0.758 0.000	0.804 0.000	0.323 0.153	0.896 0.000	0.748 0.000	0.465 0.000	0.723 0.000
lowest_Pleistocene_glacier_extent	0.481 0.027	0.765 0.000	0.785 0.000	0.324 0.151	0.881 0.000	0.736 0.000	0.446 0.001	0.697 0.000
erosion_index	-0.122 0.598	-0.624 0.003	0.216 0.348	-0.421 0.057	-0.181 0.421	-0.251 0.261	-0.090 0.509	-0.453 0.000
erosion_intensity	-0.562 0.008	-0.680 0.001	-0.571 0.007	-0.759 0.000	-0.567 0.006	-0.537 0.010	-0.494 0.000	-0.565 0.000

bold: $R > 0.5$, shaded: $R^2 > 0.5$

Table DR3: Correlations between Hypsometric Integral (HI), Profile Concavity Index (PCI), and morphometric parameters.