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Glacial buzzcutting limits the height of tropical mountains

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ABSTRACT

The widespread correlation between snowline elevation and mountain height is evidence that glacial buzzcutting puts a cap on mountain growth. The match is strongest for mid-latitude ranges, where glacial erosion has persisted over Pleistocene climate cycles and tends to truncate mountain range elevation near the upper limit of the late-Pleistocene snowline. Signs of a glacial buzzsaw are weakest in tropical ranges, where glacial erosion features are sparse and generally restricted to cold periods such as the Last Glacial Maximum (LGM). Here we show that glacial erosion does indeed truncate tropical mountains, often close to the cold-phase snowline. It does so on a cyclic basis, with glacial landscapes expanding during cold periods, and contracting during largely ice-free warm periods as fluvially-driven escarpments encroach on all sides. We find evidence of this cyclicity in the perched terrain of the Chirripó massif in Costa Rica, where surface-exposure age dating and topographic analysis show that LGM denudation occurred across a glacial landscape that has shrunk during post-LGM scarp encroachment. We find a similar story in the Central Range of Taiwan, where scarp encroachment is even more severe. We deduce that, during the Pleistocene, cold-phase glacial erosion has imposed a ceiling on tropical mountain growth, and that even the archetypally steady-state landscape of Taiwan has been subject to strongly cyclic changes in erosion rate.

1 Introduction

It has long been thought that ice-driven erosion can limit the height of mountain ranges, and there is a common (albeit controversial) attitude that glacial erosion has imposed a near-global topographic ceiling during the Pleistocene. The idea that glacial erosion processes have cut down most rock mass above the glacial equilibrium line altitude (ELA) is so pervasive that it has come to be colloquially referred to as the "glacial buzzsaw" (Brozović et al., 1996). The glacial buzzsaw hypothesis is an important contribution to the broader realization that landscapes evolve under the interacting influence of tectonically-driven crustal deformation and climatically-modulated erosion (e.g., Willett et al., 2006), and debate over its prevalence cuts to the core challenge of disentangling climatic and tectonic imprints on landscapes. The strongest proponents of a global glacial buzzsaw have implied that erosion usually overpowers even the strongest tectonic forces once uplifted rock mass reaches the glacial equilibrium line altitude (ELA), a climatically-determined elevation. Detractors claim that the power of the glacial buzzsaw has been, at best, overstated, and that in most cases, glaciations have ornamented topography whose dominant features are more accurately ascribed to patterns of tectonic forcing (Hall and Kleman, 2014). Here, we add a new spin to the story of the glacial buzzsaw by looking at its spatial imprint in an extreme environment: the hot tropics. We find that glacial erosion has limited the height of two tropical mountain ranges, one in Central America and the other in Taiwan, to the cold-phase ELA. This surprising result breathes new life into the glacial buzzsaw hypothesis by showing that it works with remarkable efficiency even in the warmest in places, and where it is thought to have little or no influence (e.g. Egholm et al., 2009).

1.1 Brief history of thought on the glacial buzzsaw

The first articulation of the idea that ice-driven erosion could limit mountain range height came over a century ago, when Dawson (1895) asserted that frost-cracking had kept the ridgeline of the Kamloops, British Columbia, at a relatively constant elevation. His claim was speculative and called attention to a striking similarity in the elevation of mountain summits throughout the range, and their apparent correlation with the bottom of the frost-cracking window. Penck (1905) made a more formal argument that the onset of glacial erosion had reshaped the Western Alps and lowered them to the cold-phase ELA. In this remarkable paper, Penck alludes to the idea that the topographic structure of the modern Western Alps is geologically young, and (in the absence of any plate tectonics framework) hypothesized that glacial erosion had outpaced the rate of rock uplift there. By the late 20th century, Porter (1977) and Broecker and Denton (1989) proposed that glacial erosion had limited mountain height on a near-global scale. Their claim was based on a growing body of data on the position and fluctuation of the Pleistocene ELA, and a qualitative match between peak mountain height and the bounds of Pleistocene ELA fluctuation along the length of the American Cordillera.

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The modern conceptualization of the glacial buzzsaw began with the work of Brozović et al. (1996, 1997), who found that the hypsometric maximum (modal average elevation) of large (~1000 km^2) regions in the Karakoram are just below the modern snowline (used as a proxy for the modern ELA), despite major differences in erosion rate, geologic structure, and local climate. They attributed this topographic feature to headward erosion focused at the ELA by cirque glaciers, a process that leaves low-sloping topography near the ELA. This line of reasoning has now been echoed for two decades, and the process of cirque headward propagation flattening valleys near the ELA has been replicated in numerous numerical modelling efforts (Anderson, 2006; Egholm et al., 2009; Macgregor et al., 2009;), and a hypsometric maximum that lies within the bounds of late-Pleistocene ELA fluctuation has been widely heralded as the signature of the "glacial buzzsaw" (e.g., Oskin and Burbank, 2005; Mitchell and Montgomery, 2006; Egholm, et al., 2009; Mitchell and Humphries, 2015).

Hypsometry has thus been demonstrated to be a compelling tool for assessing the significance of glacial erosion. Among the most important of its applications is the global analysis of Egholm et al. (2009), who used large swaths of topography (1°x1° tiles) to show that on the scale of a mountain range, hypsometric maxima almost never lie above the upper limit of late-Pleistocene ELA fluctuation. Furthermore, most mountain ranges affected by Pleistocene glaciation have a hypsometric maximum between the upper and lower bounds of ELA fluctuation, much in the way that Brozović et al. (1997) first described the glacial buzzsaw's signature in the Karakoram. It should also be noted that thermochronometric data support the coincidence of a rapid shift in exhumation rate and the onset of Pleistocene glaciation in many places that bear the glacial buzzsaw signature (Thomson et al., 2010; Shuster, 2011; Herman et al., 2013; Fox et al., 2015), thus adding confidence that, at least in some places, the topographic signal of buzzcutting is matched by an expected increase in exhumation rate coincident with Pleistocene glaciation. In summary, the glacial buzzsaw hypothesis is presently supported by a topographic signature that can be found on a global scale, and by thermochronometric data that broadly support a pattern in enhanced exhumation that coincides with the onset of wide-spread glaciation.

The idea that the glacial buzzsaw limits mountain height is a provocative one, and significant questions about its prevalence have persisted since its conception. Among the most pressing involve how to disentangle the topographic signatures of non-glacial processes from those of glaciations (e.g., Hall and Kleman, 2014). Ironically, one of the most problematic cases comes from the glacial buzzsaw's birthplace, the northwest Himalaya, where Van der Beek et al. (2009) used a combination of thermochronometry and topographic analysis to claim that Pleistocene glaciation there took advantage of an uplifted eroded surface—and that the

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correlation between region-scale hypsometric maxima and the Pleistocene ELA is coincidental. It has been proposed that this kind of topographic inheritance may make up the majority of the global buzzsaw signature (Hall and Kleman, 2014).

To make matters worse for proponents of the buzzsaw, in some places Pleistocene exhumation rates inferred from thermochronometric data show that glacial valley incision greatly outpaced cirque backcutting, calling into question the premise that glacial erosion should limit mountain height through horizontal erosion processes (Valla et al., 2011). More recently, it has been proposed that the fast, recent exhumation rates inferred from thermochronometric data suffer from observational bias (Willenbring and Jerolmack, 2016). Our paper does not directly address these issues, and instead focuses on the unique geomorphology of little-known, apparent remnants of the glacial buzzsaw in the tropics, which we use to make the case that glacial erosion has limited the height of mountain ranges even in places where it restricted to small areas.

1.2 Glaciation in the tropics

Virtually every assessment of the glacial buzzsaw has passed over the tropics, with the notable exceptions of the extensively glaciated Cordillera Blanca of Peru (Margirier et al., 2016) and the Rwenzori Mountains of Uganda (Ring, 2008). The highlands of every continent straddling the tropics were glaciated during the global Last Glacial Maximum (gLGM) (Hastenrath, 2009), and likely during other intermediate cold phases throughout the Pleistocene (Porter et al., 1989), but glacial erosion has been discounted as a factor in limiting the height of the majority of these mountain ranges (e.g., Egholm et al., 2009; Morell et al., 2012; Herman et al., 2013; Hsu et al., 2016). An obvious reason for excluding the tropics from the list of glacially buzzcut ranges is that their hypsometry, on the gross scale, shows very little area near the lower-limit of the Pleistocene ELA, and significant hypsometric maxima several km below it (Egholm et al., 2009). A second reason is that tropical mountain ranges need to be well above 3000 m for cold-phase ice erosion at all, and thus glaciations often occupy quickly eroding highlands whose height is thought to be controlled by the rate of fluvial incision.

We present evidence that glacial erosion has limited the height of two steep, narrow, and quickly eroding tropical mountain ranges in Taiwan and Costa Rica. We also find that the preservation potential of their glacial landscapes is compromised by the headward encroachment of surrounding fluvial landscapes, a process that has likely obscured the importance of glacial buzzcutting in both places throughout the Pleistocene. We first describe results from the Talamanca Range of Costa Rica (Figs. 1-3), where we develop topographic metrics to quantify the imprint of glacial erosion and present ^{10}Be surface exposure age dates

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that tie glacial erosion to the LGM. We then apply these metrics to the Central Range of Taiwan (Fig. 2-3), where previous work has indicated isolated LGM glaciation, and uncover signs that cold-phase glacial erosion was likely much more widespread there than has been assumed.

2 Tectonic and geomorphic setting

2.1 Cerro Chirripó, Talamanca Range, Costa Rica

The Talamanca Range is a high section of the Central American Volcanic Arc that stretches for ~175 km from central Costa Rica to western Panama (Fig. 1A). The range largely comprises Miocene volcanics and intermediate plutonics that intruded volcanic rocks around 8 Ma (Drummond et al., 1995) and cooled to <65°C by around 5 Ma (Morell et al., 2012). The central Talamanca coincides with the subduction of the aseismic Cocos Ridge, one of the most striking features of the Central American convergent margin. Subduction of the Cocos Ridge is thought to have contributed to the onset of rapid rock uplift, the development of a bivergent wedge, and the cessation of arc volcanism in the Talamanca (Morell et al., 2012). Recently, several authors have converged on the conclusion that Cocos Ridge subduction initiated sometime after 3 Ma, and that the extinct arc has been uplifted by a minimum of ~2 km in this time (Morell et al., 2012; Zeumann and Hampel, 2017).

Several studies have also attempted to link the erosional history of the Talamanca to the onset of Cocos Ridge subduction, including, for example, river longitudinal profile analysis coupled with thermochronometric data (Morell et al., 2012) and the macro-scale morphology of river networks (Zeumann and Hampel, 2017). Significant disequilibrium observed in Talamanca river networks is thought to record a switch to a higher rate of rock uplift during the last 3 Myr and that this fast rock uplift rate persists today. Important elements of this general narrative include the presence of areas of anomalously low-relief topography found at high elevations (e.g., between 2000-3000 m), which are thought to represent an eroded surface that has been advected to its present-day elevation by rapid rock uplift associated with the subduction of the Cocos Ridge. Our work does not attempt to directly address this interpretation, and neither do our results contradict the claim that a major shift in the rate of rock uplift has occurred recently in the Talamanca. Rather, we make a case that there was sufficient rock mass above the ELA for glaciation to occur during LGM, and probably during cold stages prior to the LGM.

The highest landscape of the Talamanca Range is the Chirripó massif, a low-relief terrain spanning ~75 km^2, perched above ~3000 m above sea-level, and surrounded by rugged high relief ridge-and-valley topography.

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Glacial landforms – such as lateral moraines, glacially striated bedrock, roches moutonnées, over-deepened lakes, cirques and U-shaped valleys – were first reported on Chirripó in the 1950's and have been studied episodically since then (Barquero and Ellenberg, 1986; Bergoeing, 1978; Hastenrath, 1973; Lachniet and Seltzer, 2002; Orvis and Horn, 2000; Shimizu, 1992; Weyl, 1955). The most prominent cirques cut into the Cerro Chirripó peak at 3819 m, but smaller cirques are also scattered around the massif. Lateral moraines have been mapped at elevations as low as ~3150 m and as high as 3450 m in Valle de las Morrenas and Valle Talari, and hummocky recessional type moraines can be found on cirque floors as high as 3500 m.

Orvis and Horn (2000) provided the first rigorous ELA estimate for maximum ice extent at Chirripó. They suggested an ELA of ~3500 m, based on standard ice surface reconstruction and a combination of balance ratio (BR) and accumulation area ratio (AAR) methods. This estimate corroborates earlier estimates by Weyl (1955) and Hastenrath (1973), who also suggested an ELA of 3500 m based on the elevation of cirque floors. Lachniet & Seltzer (2002) independently estimated a similar ELA using both AAR methods and the maximum elevation of lateral moraines. Although multiple authors have converged on a consistent ELA estimate using a variety of methods, the only age constraints for glacial timing available are minimum-limiting bulk ^{14}C dating of organic material in postglacial lakes (Orvis and Horn, 2000). Today, Costa Rica contains no glaciers, and snow has never been reported even at the highest elevations. Prior studies also inferred a lack of glacial activity during the Holocene.

2.2 Nanhudashan, Central Range, Taiwan

The Central Range of Taiwan is the product of the oblique collision of the Luzon Arc and Eurasia, and it is comprised of metamorphosed marine sediments and pre-Cenozoic basement (Suppe, 1981). Due to the oblique nature of collision, it is thought that deformation has propagated from north to south for 5-7 Ma (Byrne and Liu, 2002). The relatively constant peak elevation and width of the Central Range and its thermal history are thought to be an indication of topographic and flux steady state for the northern 150 km of the range (Willett and Brandon, 2002; Stolar et al., 2007). Recent work has countered this idea by putting forth the case that a rapid increase in exhumation rate occurred along the entire strike of the range starting between 1-2 Ma (Hsu et al., 2016). Low-relief surfaces found at high elevations (above 3000 m) in Taiwan are thought to be remnants of an eroded surface formed sometime prior to 1-2 Ma (Ouimet et al, 2015).

The highest peaks of Taiwan's Central Range bear signs of LGM glacial erosion (Hebenstreit et al., 2006; Siame et al., 2007; Hebenstreit et al., 2011) and lie within ~500 m of the estimated LGM ELA of ~3400 m

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(Hebenstreit et al., 2011). The best preserved of these remnants are found at Nanhudashan, in the northeast Central Range. Siame et al. (2007) carried out ^{10}Be analysis of scoured bedrock and boulders perched on moraines at Nanhudashan and found relatively young (15-9 ka) glacier retreat ages, but concluded that the end of glaciation in Taiwan is broadly consistent with end of the global LGM. Hebenstreit et al. (2011) found similar ^{10}Be ages of glacial landforms, and estimated an LGM ELA of ~3400 m.

3 Methods

Here we attempt to assess glacial erosion features at Cerro Chirripó and Nanhudashan in the context of the glacial buzzsaw. The glacial geomorphology of both places has already been described in detail, but no prior work has made a direct attempt to look for and describe the buzzsaw signature. Furthermore, prior work has implied that glacial erosion in both places has been ineffective at limiting mountain height (e.g., Egholm et al., 2009). We therefore develop topographic analysis techniques specifically for these tropical massifs, using Cerro Chirripó as a training ground, and then Nanhudashan as a separate application.

3.1 Topographic assessment of glacial erosion

The question of whether the glacial buzzsaw has acted in Costa Rica and Taiwan hinges on measurement scale. The signature of the glacial buzzsaw is a hypsometric maximum within the bounds of Pleistocene ELA fluctuation, and on the scale of the entire Talamanca and Central Range, there is very little area near the lower limit of ELA fluctuation. However, a hypsometric maximum near the ELA is thought to arise from headward cirque propagation (Brocklehurst & Whipple, 2004; Oskin & Burbank, 2005; Shuster et al., 2011; Gran-Mitchell & Humphries, 2015), and since this process acts on the scale of a first-order glacial catchment, a hypsometric maximum near the ELA for a large area (~1000 km^2) presumably records the aggregate of many individual valleys with hypsometric maxima near the ELA. Our first goal was to assess whether the glacial buzzsaw signature was recorded in individual glaciated valleys, and to do this we implemented focused hypsometric analysis.

Focused hypsometric analysis is simply our attempt to objectively isolate glaciated catchments in our targeted massifs. Glaciated catchments in these places typically have an area of 5 km^2 and total relief of 800 m, so we delineated catchments that drained to ~800 m below the main divide of the massif (Fig. S9). Our method of high catchment delineation ensures that we capture all ~5 km^2 catchments that, if they lie at high elevations, would be glaciated, and that these catchments cover substantial elevation (e.g. >500 m) below the estimated

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ELA of 3500 m. This collection of high catchments allows us to focus hypsometric analysis on the highest landscapes of the mountain range, and to rigorously determine the dominant elevation of the main divide.

Next, we sought a way to quantify how the hypsometry of glaciated valleys compares to the fluvial systems they occupy. In other words, we sought a straightforward way to assess the significance of glacial-type hypsometry on the scale of larger basins dominated by fluvial erosion. To do this, we calculated the hypsometry of the largest possible catchments whose headwaters were entirely composed of glaciated valleys. Glacial features had been previously mapped by multiple groups and confirmed by our own work, and so extracting these large catchments was a relatively straightforward procedure: we delineated each large catchment using an outlet elevation just above the confluence with a trunk stream of a completely non-glaciated catchment. There are more sophisticated ways of comparing the hypsometry of glacial catchments to the larger fluvial basins they occupy, and to this end we present a preliminary analysis in which we track how the hypsometric maximum of a catchment varies with catchment outlet elevation (Fig. S8), but we suggest that for our purposes, the simple method of comparing the hypsometry of small catchments and large catchments is sufficient.

3.2 Surface exposure age dating at Chirripó

Six samples were collected for ^{10}Be exposure dating from boulders embedded in both lateral and frontal recessional moraines at 3400–3500 m elevation in Valle de las Morrenas and Valle Talari, two samples from scoured bedrock within ~15 m of the Chirripó summit, and one sample from a landslide boulder sourced from a cirque headwall (Fig. S2). Moraines in both valleys extend >1 km farther below the lowest sample sites (Fig. S1). Processing at Lamont-Doherty Earth Observatory and measurement at Lawrence Livermore National Laboratory followed standard procedures (e.g., Schaefer et al., 2009), and ^{10}Be ages were calculated with the CRONUS-Earth online calculator (Balco et al., 2008) v. 2.2, using a low latitude, high elevation production rate obtained in Peru by Kelly et al. (2013) and the scaling scheme of Lal (1991) and Stone (2009) (Table S1, S2). Our conclusions are not affected by the choice of production rate or scaling.

3.3 Mapping of post-glacial scarp

The glacial valleys at Chirripó generally have a much lower topographic slope than the fluvial valleys flanking them, and the very presence of well-preserved glacial landforms, such as sharp-crested moraines, indicates that post-glacial erosion is slow in glaciated zones. In contrast, signs of relatively fast erosion, such as frequent landsliding, are apparent in the surrounding fluvial valleys in satellite imagery and air

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photographs. There are no direct measures of erosion rate for these fluvial catchments, but it is thought that a fluvial erosion rate of ~1 $mmyr^{-1}$ has been sustained in parts of Costa Rica for >2 Myr (Morrell, et al., 2012). The boundary between the fluvial and glacial domain in these landscapes usually gives rise to an erosion front, that is, a pronounced topographic break between the slowly eroding, relatively low-sloping glacial valleys and steep, quickly eroding fluvial valleys. We interpret these erosion fronts as transient features that mark an ongoing, headward push by the fluvial network into disconnected glaciated valleys.

We used a combination of high-resolution (<1 m) satellite images, aerial photographs, a slope map produced with 1 arc-second (nominal 30 m resolution) Shuttle Radar Topography Mission (SRTM) digital elevation, and field observations to map erosion fronts in both places (Fig. 1C). Erosion front mapping was guided by both sharp changes in slope as well as the abrupt disappearance of glacial deposits and the transition to non-glaciated bedrock cliff faces (Fig. S3-5). It should be noted that our erosion front mapping was a qualitative endeavor. To our knowledge, there are no standard metrics designed to discern the kind of competition between perched glaciated valleys and headward propagating fluvial valleys that we observe, and thus we find a qualitative analysis to be appropriate. Traditional, fluvial-based quantitative metrics, including normalized channel steepness, are not particularly useful for our purposes, since the mere presence of significant glaciation has altered the landscape beyond any meaningful interpretation of these metrics.

To guide our qualitative assessment, we developed a set of rules. First, we used a binary slope map (threshold of 35°) to identify places where low-sloping glacial valley floors made a hard transition to a fluvial-linked escarpment (Fig. 1). In places where these escarpments where linked to amphitheater heads, we mapped the entire amphitheater head, and thus the initial smoothness of erosion front boundaries are set by the 30 m resolution DEM, and not by the sub-meter resolution imagery. Next, we used the imagery to check that all mapped erosion fronts coincided with the disappearance of glacial features or otherwise clear signs of ongoing erosion, such multiple landslide scars. Finally, we excluded mapped zones that appeared to be related to isolated events, such as single landslide scars that were not unambiguously linked to the ongoing propagation of the fluvial network into glaciated terrain.

4 Results

4.1 Topographic signature of buzzcutting

Most high-elevation catchments at Chirripó have a hypsometric maximum within several meters of the estimated LGM ELA (Fig. 2-3). Large catchments, whose upper divide is composed entirely of glaciated

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valleys, almost all have a multimodal hypsometry (Fig. 3) but in each catchment the higher modal elevation also sits at the estimated ELA (Fig. 3). The strength of the lower modal elevation varies substantially from valley to valley, and primarily according to catchment geometry. Thus, the observation that the hypsometric maximum matches the ELA in high elevation (glacial) catchments is not purely a function of our choice of scale. That is, the glacial buzzsaw signature is apparent even when the catchment's outlet elevation is multiple km below the ELA.

Two high catchments have a hypsometric maximum somewhat higher than the cold-phase ELA, which we cannot reasonably be attribute to lithology (Fig. S7), aspect, or local variation in the height of the ELA. Of particular relevance, these catchments appear to be heavily modified by scarp encroachment, which leads us to infer that fluvial scarp encroachment has erased the lower portion of their glacially eroded topography and has biased their hypsometric maximum to higher elevations.

To quantitatively describe the pattern of fluvial scarp encroachment into glaciated terrain, we define two new metrics. The first, which we term the ELA-Relative Modal Elevation (ERME), measures the difference between high catchment hypsometric maxima and the estimated LGM ELA of 3500 m (Orvis and Horn, 2000). The second, which we call the Scarp Encroachment Ratio (SER), is an approximation of the headward distance traveled by the fluvial escarpment into each glaciated valley (Fig. 4). To estimate this distance, we extracted glacial catchments at an elevation threshold of 3000 m, which is a good approximation for the lower limit of glacial ice at the LGM in both Costa Rica and Taiwan. We then approximated the area below the escarpment A_c by finding the area of the catchment below the lowest elevation along the escarpment. We calculated a corresponding length scale for both scarp-affected area (L_c) and the above-escarpment area (L_g). SER is the relative length scale

$$SER = \frac{L_c}{L_c + L_g} \tag{1}$$

These metrics show that high catchments with a hypsometric maximum above the 3500 m ELA are also those most heavily encroached (Fig. 5), a pattern that most likely results from the loss of glacially-eroded terrain to the fluvial network below as scarps propagate headward into glacial valleys. The selection of a 3000 m benchmark used to estimate scarp encroachment is based on a rough estimate, and the most literal interpretation of SER is that it measures of the total elevation gain of post-glacial scarp retreat. SER is a first estimate of catchment area lost to scarp encroachment—a value otherwise difficult, and perhaps impossible, to obtain, and it is better interpreted as a measure of the proportion of glacially eroded terrain and scarp-affected terrain in high catchments.

4.2 Timing of glacial erosion at Chirripó

Our ^{10}Be ages tie the glacial Chirripó landscape to the LGM and provide the first constraints on its termination in Costa Rica. Lateral and recessional moraine boulders yielded ages between 18.3 ± 0.5 ka and 16.9 ± 0.5 ka, and in Valle de las Morrenas ages tend to young toward the headwall. Near the Chirripó summit, a bedrock surface gave an age of 22.0 ± 0.7 ka. This age may reflect thinning of the ice prior to ~18 ka, if there is no inherited ^{10}Be in the sample. A landslide boulder sourced from a cirque headwall and deposited above moraines gave an age of 15.2 ± 0.5 ka, which also places a constraint on ice-free conditions at Chirripó. The other bedrock sample yielded an age of 8.9 ± 0.4 ka, which we consider unrealistically young, likely due to bedrock burial by soil and/or other sediment (Fig. S2)

4.3 Comparison with Nanhudashan, Taiwan

ERME and SER reveal patterns of buzzcutting and scarp encroachment at Nanhudashan that are similar to those found in Costa Rica, only more advanced (Figs. 2-3, 5). Of three glacial catchments at Nanhudashan, two show greater scarp encroachment than any glacial catchment at Chirripó, and this is reflected in their ERME and SER values. The escarpments propagating into glacially eroded terrain at Nanhudashan, some of which were mapped in previous studies (e.g., Hebenstreit et al., 2006; Willett et al., 2014), are particularly spectacular. When combined, our observations at Chirripó and Nanhudashan capture a continuum of scarp encroachment into glacially eroded landscapes and the alteration of glacial-type hypsometry.

5 Discussion

5.1 The glacial buzzsaw in the tropics

Our most important finding is that every catchment flanking the main divide of both the Talamanca and Central Range has a hypsometric maximum that is set by glacial erosion. The presence of glacial erosion features, the LGM age of these features, and the multimodal hypsometry of large glaciated catchments all support the claim that glacial erosion is uniquely responsible for the match between high hypsometric maxima and the ELA. Our contributions have been to identify and describe this glacial-type hypsometry for the first time in both ranges and to tie it directly to evidence of glacial erosion.

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There are many ways to explain the hypsometry of either mountain range without invoking height-limiting glacial erosion, but any proposed process history for these mountain ranges must accommodate the non-trivial condition that high-catchment hypsometric maxima coincide with the cold-phase ELA in both places. In other words, any plausible evolution of these two tropical mountain ranges must either include the glacial buzzsaw or the coincidence that high hypsometric maxima of both ranges occur at a similar elevation and the simultaneous coincidence that this elevation is the LGM ELA. For example, it could be that the peak elevation of both ranges is controlled by non-glacial processes, and those processes are broadly in equilibrium with tectonically-driven rock uplift. If this is the case, it must also be true that both ranges have coincidently reached an equilibrium elevation (~3500-3800 m) that includes just enough rock mass above the cold-phase ELA for periodic glaciation. Various permutations of this line of reasoning lead to the same conclusion: one range could be in topographic steady state, and the other in state of transience, or both ranges could be in a state of transience with respect to peak elevation. In any of these cases, glaciations must have occupied an inherited topographic structure that is unrelated to glacial erosion. Our interpretation, that the peak elevations of both the Talamanca and Central Range is set by glacial buzzcutting, simply favors the most parsimonious explanation.

That the determination of the ELA is based on the geometry of present topography, since it is derived from geomorphic features such as lateral and terminal moraines, does not weaken our parsimonious explanation. For example, fast uplift rates in Taiwan will certainly push the glacial features presently observable to some elevation above the apparent cold-phase ELA in the future, and had the ELA estimate in either place been established 5-10 kyr later than today, we would make the claim that glacial erosion had capped the elevation of the Central Range at some modestly higher elevation. To be clear, our argument is that glacial buzzcutting in these places is intermittent but sufficient to limit mountain height, and that if the ELA has repeatedly come back to similar elevations during glacial periods, then the entire range will be limited to that elevation on the long term. Put another way, the highest catchments of both ranges are at an elevation that was prone to ice sufficient to drive glacial erosion, regardless of when or where that glacial erosion took place. Again, we argue the most parsimonious explanation for this phenomenon is that in both the Talamanca and Central Range, glacial erosion has repeatedly cut peaks down to the ELA.

5.2 Hypsometry at different scales

Glaciated tropical mountain belts have traditionally been considered to be limited in height by fluvially-driven erosion and not glacial erosion in part because glacial erosion features make up a small portion of the landscape. An important question is how to assess the significance of the buzzsaw signature in the valleys

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where we observe it, and whether the analysis of small catchments has produced a spurious correlation between catchment hypsometric maxima and the ELA. To address this question, we turn to the interesting result that large catchments with glaciated upper valleys have a multimodal hypsometry.

A hypsometric maximum near sea-level is usually observed for very large areas dominated by fluvial erosion because large-scale deposition favors low elevations, but in erosive landscapes the modal elevation of a catchment is a strong function of the branching structure of its drainage network (Willgoose and Hancock, 1998). Thus, the sub-ELA modal elevations in our large catchments should be interpreted as a function of the geometry of the catchment drainage network. Large catchment hypsometry shows that regardless of the geometry of the fringing fluvial network, each catchment has a discernible glacial influence (Fig. 3). The multimodal hypsometry of flanking fluvial catchments is (perhaps counterintuitively) a sign of the strength of glacial erosion in these places: no matter the geometry and style of erosion in the flanking fluvial valleys, the highest modal elevation in nearly all large catchments is the ELA. The simplest explanation of this pattern is that glacial erosion has overprinted the pre-existing fluvial hypsometry in every catchment.

The only large catchment whose hypsometry does not have a clear modal elevation near the LGM ELA is SW Nanhudashan, which is also the most heavily encroached of the glacial valleys we analyzed. The hypsometry we observe could be brought about by glacial ornamentation of a plateau in the final stages scarp consumption, but, again, we argue that the most parsimonious explanation for the hypsometry we observe at SW Nanhudashan (given the patterns of glacial erosion and scarp encroachment elsewhere at Nanhudashan and at Chirripó) is a competition between glacial erosion and scarp encroachment, such that a fluvially-driven escarpment has propagated headward into a glacial valley, and nearly wiped it from the landscape entirely.

5.3 Scarp encroachment into glaciated landscapes

LGM glaciation in both the Talamanca Range and Central Range was isolated to relatively small massifs, and it is self-evident that these landscapes make up a small proportion of the total mountain range relief (~800 m out of ~3800 m in both Costa Rica and Taiwan). Why suggest that glacial erosion has ultimately capped the rise of either mountain range if glaciation appears to have had such a small spatial imprint? The answer lies in the apparent duel between glacial erosion and fluvial scarp encroachment.

The ongoing destruction we observe of glacial landscapes by scarp encroachment points to their low preservation potential, and indeed, several glacial valleys appear to have lost substantial area to scarp encroachment since they were last affected by ice at the LGM, ~17 ka (Fig 5, Fig. S3-5). The rate at which scarps propagate into glacial landscapes is difficult to estimate, but it is clear that substantial shrinkage of these landscapes can happen within the long (~100 kyr) period of late-Pleistocene glacial-interglacial cycles, if not in Costa Rica then certainly in Taiwan. We thus deduce that the simplest explanation for the correlation between the peak elevation of both ranges and the simultaneous correlation between peak elevations and the ELA is cyclic phases of buzzcutting separated by phases scarp encroachment. In this context it is not surprising that glaciated massifs are sparse in these ranges and that their tile-scale hypsometric maxima sit below the cold-phase ELA, even though their height is ultimately tied to this elevation. We further deduce that climatic fluctuation induces a duel between downward-pushing, outward-spreading, glacier-driven erosion fronts during glacial maxima, and horizontally-pushing, fluvially-driven erosion fronts during warm periods (Fig. 6). Importantly, even if the erosion front duel is biased towards scarp encroachment and entire glacial landscapes are erased during warm periods such as the present interglacial, it is still cold-phase glacial buzzcutting that imposes the ultimate top-down ceiling on mountain growth.

5.4 Steady state landscapes revisited

For decades, the spatially constant width and maximum elevation of the highly dynamic Central Range of Taiwan have been considered signs of steady state, with vertical erosion balanced by tectonic uplift (Suppe, 1981; Willett & Brandon, 2002). This simple model of steady state has been placed in question by demonstrations that horizontal divide migration arises from autogenic dynamics (Stark, 2010; Willett et al., 2014). Our deduction that the maximum elevation of the Central Range of Taiwan is tied to pulses of glacial erosion casts further doubt on the concept of steady-state *sensu stricto* in the Central Range. Instead, our evidence in both Taiwan and Costa Rica points to a cyclicity that comprises constant imbalance between erosion rate and rock uplift, with climatically induced, alternating phases of glacial erosion and horizontal scarp erosion. This oscillatory pattern of erosion may forge the signature of steady state by lowering the Central Range to a spatially constant maximum elevation.

6 Conclusion

Glacial erosion has limited the elevation of tropical mountain ranges throughout the Pleistocene by repeatedly buzzcutting peaks that rise above the cold-phase ELA. Glaciation accelerates erosion during cold phases through buzzcutting that removes rock mass above the ELA and primes the landscape for rapid horizontal scarp encroachment during warm periods. We argue that a competition between glacial buzzcutting and horizontal scarp encroachment promotes a cyclic imbalance between uplift rate and erosion rate in tropical mountain ranges, preventing the establishment of steady-state in the classical sense.

COMPETING INTERESTS

The authors declare that they have no conflict of interest.

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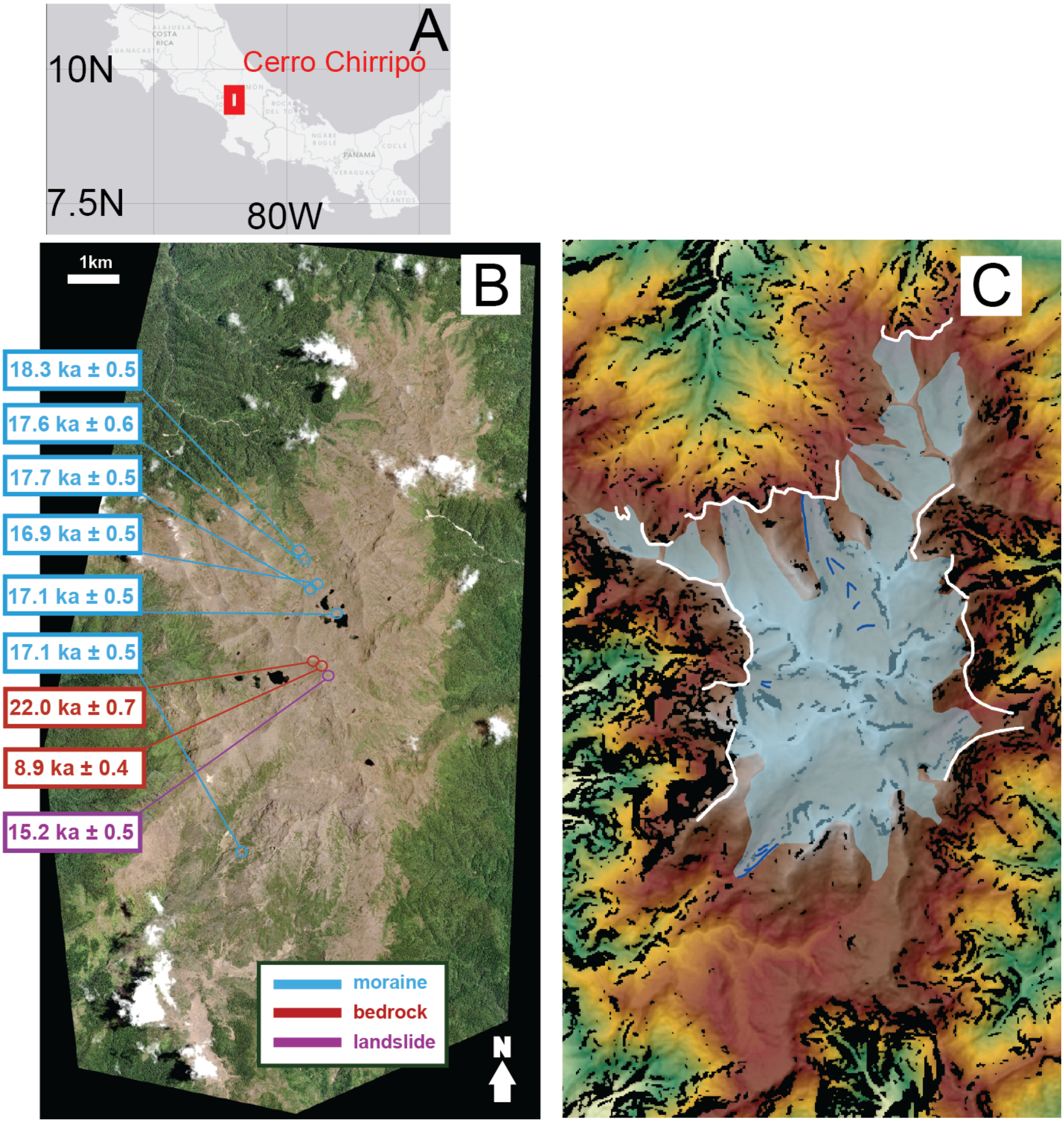

Figure 1: Glacial features at Cerro Chirripó, Costa Rica. A: Location map. B: ^{10}Be exposure ages ($\pm 1\sigma$) on WV2 image. C: Reconstructed LGM ice extent (blue) with moraines (dark blue), post-glacial scarp (white), elevation (green through brown), and >35° slopes (black). 1" SRTM DEM.

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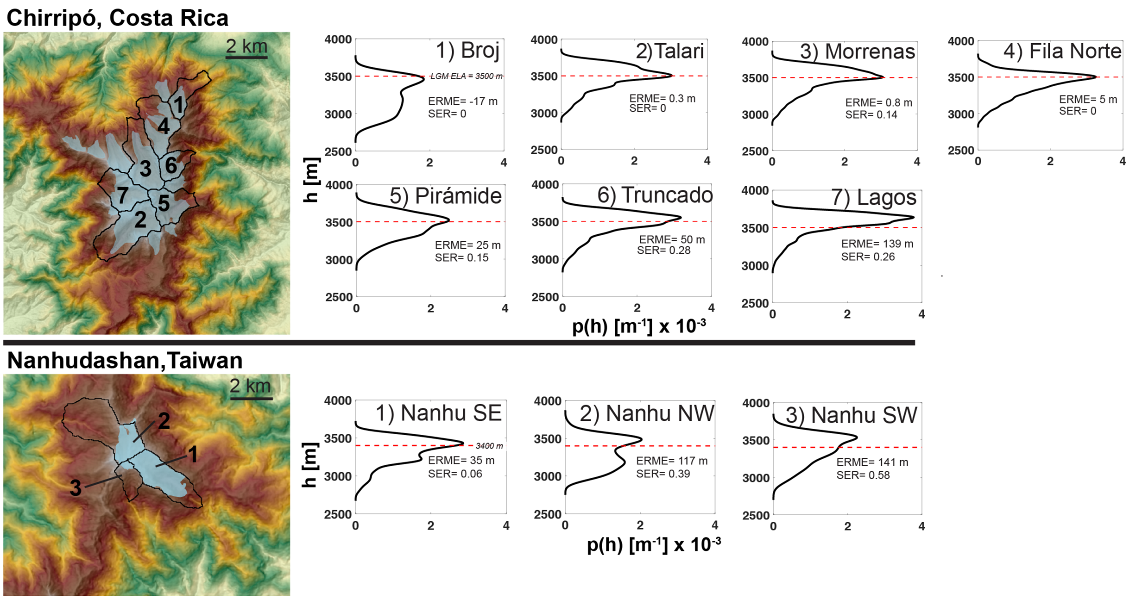

Figure 2: Chirripó and Nanhudashan hypsometry. Focused hypsometry for Cerro Chirrpó (top) and Nanhudashan (bottom). Target catchments are delineated in black, with corresponding elevation probability distribution on right (each PDF is constructed using kernel density estimation). LGM ice extent is in blue shade. Red dashed line in elevation PDF plots is LGM ELA (3500 m at Chirripó, 3400 m at Nanhudashan).

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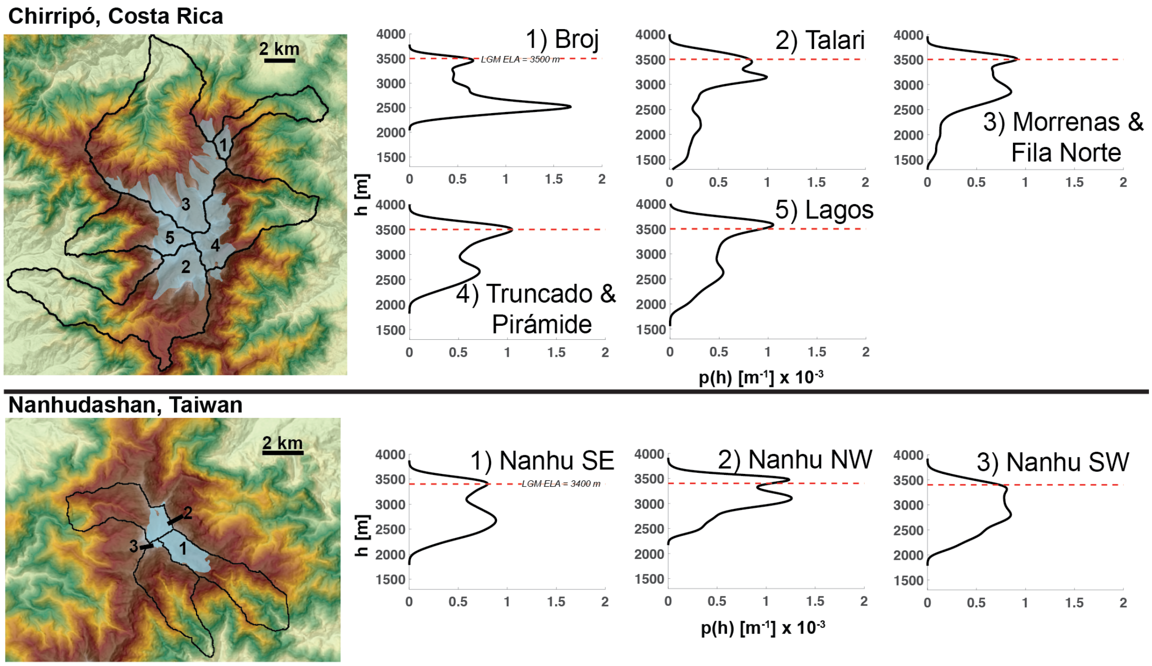

Figure 3: Large catchment hypsometry. Colors same as Fig. 2. Catchments analyzed are the largest possible catchment whose headwaters are completely glaciated.

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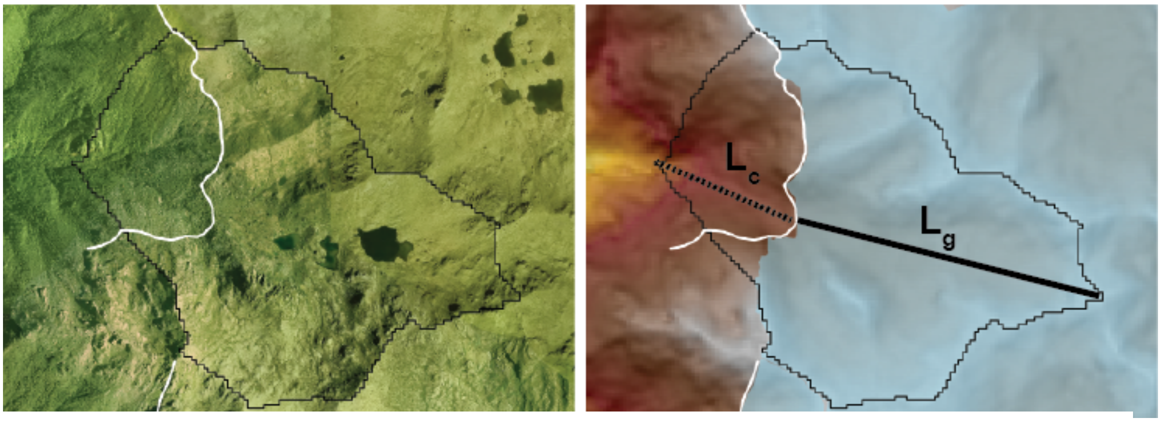

Figure 4: Scarp Encroachment Ratio (SER) calculation. We delineate the catchment that drains to 3000 m for each glacial valley, and then calculate the area above and below mapped escarpments. Length scales L_c and L_g are the square root of each mapped area. Field photos of escarpments are included in the supplement (Fig. S3).

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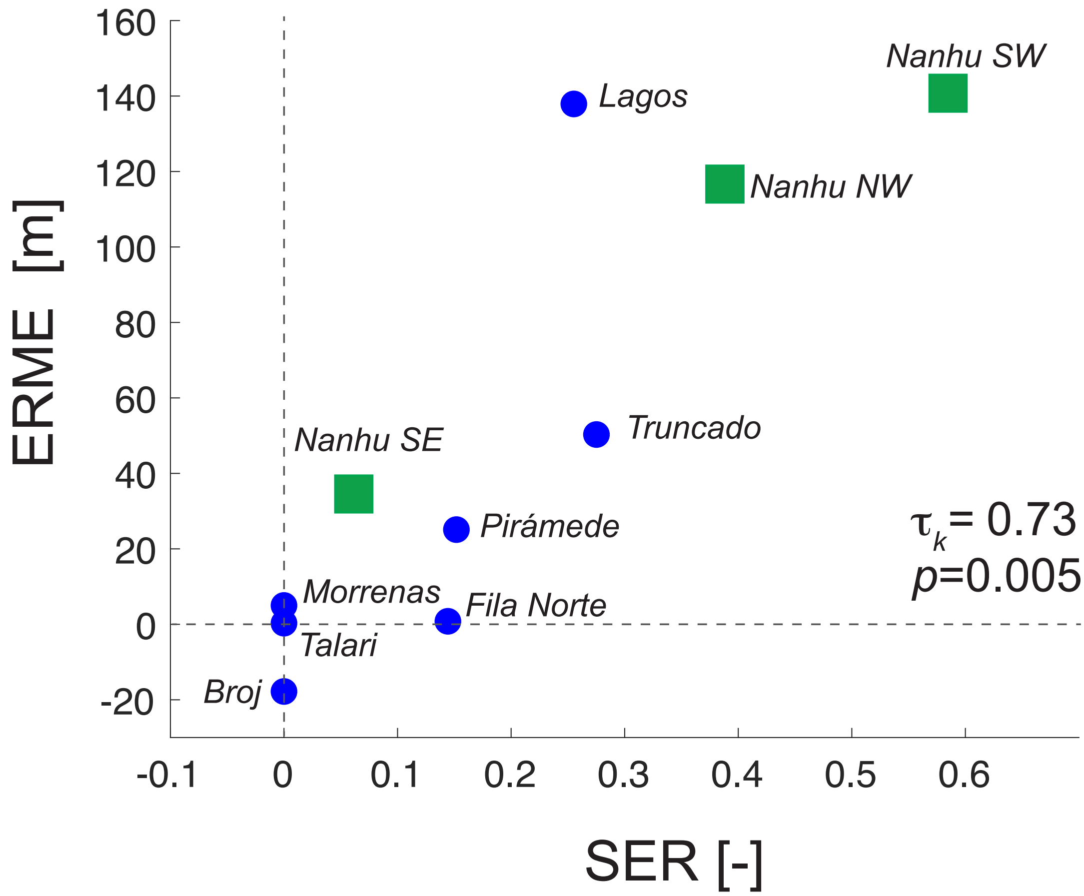

Figure 5: ERME vs. SER. ELA-Relative Modal Elevation (ERME) and Scarp Encroachment Ratio (SER) for each catchment at Chirripó (blue) and Nanhudashan (green) (names in Fig. 2). Kendall's Tau (0.73) reported for entire data set. One-tailed significance test yields a p-value of 0.005.

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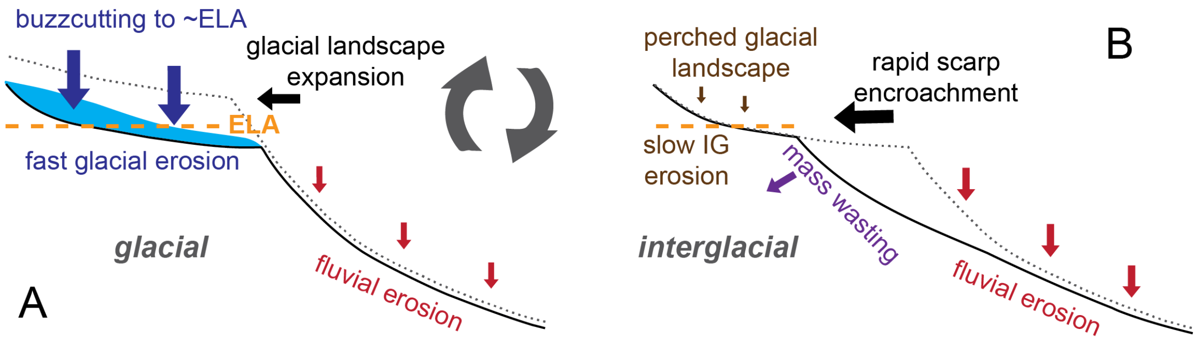

Figure 6. Cartoon of erosion front duel. (A) Glacial buzzcutting removes rock mass above the cold-phase ELA and expands low-relief valleys, after which (B) horizontally moving scarps driven by fluvial incision encroach glacial valleys.