5.13 Glacially Influenced Tectonic Geomorphology: The Impact of the Glacial Buzzsaw on Topography and Orogenic Systems

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Abstract

Glacial erosion is approximated as a simple linear function of basal sliding, but is actually dependent on a myriad of spatially and temporally varying boundary conditions. Glacial erosion is more proficient than fluvial erosion, but both are limited by the local rate of tectonic rock uplift. The unique characteristics of glacial topography can be linked to mechanics of the erosion process and local conditions (e.g., selective linear erosion under polythermal ice conditions), as demonstrated in surface process models. Some kinematic aspects of glacial erosion are exclusive, including a focusing of erosion at equilibrium-line altitude, aggressive headwall retreat, and relatively rapid topographic adjustment to climate change. As a result of these characteristics, orogenic systems behave differently when dominated by glaciers. Under the right conditions, rapid rock uplift and efficient glacial erosion limit topography, induce steady-state orogenic flux, and focus crustal strain, thus imparting geodynamic significance to glacial climate. This hypothesis of a 'glacial buzzsaw' has been verified both topographically and exhumationally in a variety of settings worldwide, as well as in numerical models of orogenic evolution. Unraveling exceptions to this, including 'teflon peaks' and glacial thresholds, is key to improving understanding of the tectonic geomorphology of glaciated terrain.

5.13.1 Introduction

Glacial erosion is known to be very proficient and to have an extreme influence on topography and tectonic systems (Hallet et al., 1996; Brocklehurst and Whipple, 2002; Berger et al., 2008a). As such, a comprehensive review of tectonic geomorphology would be incomplete without a dedicated discourse on the processes of glacial erosion and their effects. This chapter reviews the basics of glacial erosion processes as a foundation, given the important distinctions that these processes have relative to more familiar fluvial erosion. The chapter implicitly considers glacial erosion to mean erosion by temperate, alpine glaciers, and exclude erosion by polar glaciers and ice sheets unless otherwise stated. This review of glacial erosion culminates with the physical basis for erosion rules that are commonly used in surface process models and highlights both the complexity and elegant simplicity of glacial erosion. Next, the chapter reviews how glacial erosion influences topography, including examination of how glacial erosion modifies pre-existing fluvial topography and the timescales over which this occurs, as well as discusses whether glaciers increase topographic relief. This leads to a review of our understanding of specific elements of glacial topography, such as longitudinal profiles, based on modeling. Finally, the influence of glacial erosion on orogenic systems, based on combined empirical and modeling-based approaches is explored. This includes the importance of glacial equilibrium-line altitude (ELA), the glacial buzzsaw hypothesis, thresholds in glacial erosion, and the partitioning of deformation and reshaping of orogenic wedges.

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5.13.2 Basics of Glacial Erosion

5.13.2.1 Erosional Processes

Bedrock erosion by glaciers is accomplished as a combination of abrasion and quarrying. Abrasion, or scouring, occurs as glaciers drag debris over the ice–bedrock interface. This is a smoothing process, which removes rough edges and results in streamlined bedrock forms, such as striations, grooves, and the ice-flow-facing sides of roche moutonnées (cf. MacGregor et al., 2009). Abrasion produces fine sediment by direct bed erosion and the comminution of clasts that have become entrained in the ice. The rate and efficiency of abrasion are controlled by the basal sliding rate and the presence of tools, or hard entrained debris (Hallet, 1979; Lliboutry, 1993). Abrasion is efficient on slopes that face opposite the direction of ice flow, but inefficient where ice-bed separation occurs downstream of subglacial topographic features. Quarrying is the glacial equivalent of fluvial plucking, and occurs as glaciers fracture and remove large blocks from the bedrock floor. This produces coarse sediment and increases bed roughness (Hallet, 1996). The efficiency of quarrying depends on the presence of rough edges or steps on the bed, basal water pressure (Cohen et al., 2006), and the material properties of the rock, particularly the spacing of discontinuities (Duhnforth et al., 2010). Quarrying is more efficient downstream of subglacial topographic features, and is in turn important for generating new irregularities in the bed topography.

The relative importance of abrasion and quarrying depends on local conditions, including the bedrock properties, subglacial hydrology, topography, and temperature. Abrasion can have nonlinear dependence on special conditions, such as the influx of large debris to the glacier bed via the bergschrund or crevasses or the production of sand grains via chemical and physical weathering (Lliboutry, 1993). The efficacy of abrasion and quarrying can vary over short distances beneath individual glaciers or over time at a given location, with timescales of variation that can be decadal, seasonal, or even as short as days (Anderson et al., 2006). Lliboutry (1993) suggested that quarrying should dominate early during a glacial cycle, where there are numerous rough edges and rock is fractured, and that abrasion should gradually become more important as more intact bedrock is exhumed. Although either process can dominate locally, they are to some degree coupled, such as via the influence of quarrying on abrasion by controlling the production of erosive tools (Anderson et al., 2006). Despite local variation, however, there are indications that quarrying is, on average, responsible for more bed lowering than abrasion (Loso et al., 2004; Amundson and Iverson, 2006). For example, Riihimaki et al. (2005) estimated that quarrying is responsible for at least 2/3 of erosion beneath a small glacier in southern Alaska, based on the relative yields of bedload and suspended load in glacial outwash.

A particularly important process that sets dynamic boundary conditions for glacial erosion is the subglacial hydrologic system, which broadly includes variables of subglacial water pressure, water storage, water discharge, flux of meltwater from the surface, and subglacial channelization (cf. Fountain and Walder, 1998; Clarke, 2005; Flowers, 2008; Kavanaugh, 2009). Basal hydraulics are important for glacier sliding, which in turn influences abrasion and quarrying. For example, supercooling

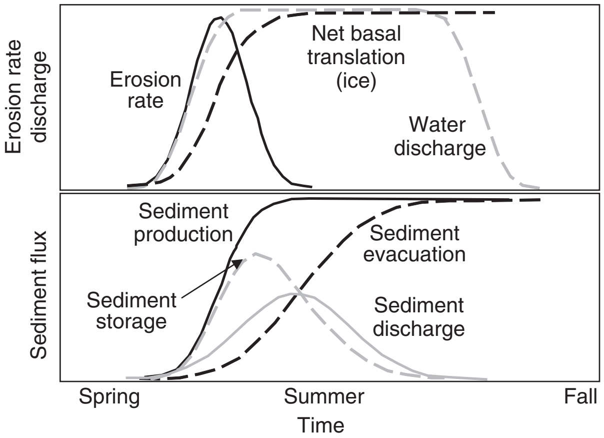

Figure 1 Conceptual relationships between glacier flow (net distance of basal translation, or horizontal ice motion), erosion rate, water discharge (at toe of glacier), and sediment budget (production, discharge at toe of glacier, net evacuation and storage), based on observations in the ablation region of the Bench Glacier in Alaska (redrafted based on interpretations of Riihimaki et al., 2005). The net ice motion is based on seasonal GPS observations of one mid-glacial station (#2) in 2002, and was not included in the original conceptual diagram. Notice the correlation between glacier sliding, erosion, and water discharge, all of which peak in early summer. The water discharge remains high throughout the melt season, although glacier sliding and erosion both decrease. Sediment production is linked to erosion, whereas sediment discharge lags behind sediment production because it also depends on water discharge. Sediment evacuation represents the net amount of sediment discharged. The rate of sediment discharge eventually decreases at the peak of the melting season, despite continued high meltwater discharge, because the sediment production due to erosion has ceased and sediment storage has been fully depleted.

and ice growth in overdeepenings can freeze a glacier to its bed, thereby minimizing further erosion and 'grading' the glacier (Alley et al., 2003). Basal water pressure also controls the pore pressure in bedrock cracks, which influences the ease of quarrying (Cohen et al., 2006), as evident in the prevalence of quarrying in cavities in the lee of subglacial topography where fluids accumulate (Hallet, 1996). Subglacial hydrologic discharge and connectivity of outwash conduits are also critical for removal of sediment produced by glacial erosion, which could otherwise accumulate into a till layer that would effectively shield bedrock from erosion (Hooke, 1991; Tulaczyk et al., 1998; Alley et al., 2003; Riihimaki et al., 2005). Correlations have been observed between subglacial water discharge and both sediment yield and glacier sliding (Humphrey and Raymond, 1994; Bogen, 1996; Riihimaki et al., 2005) (**Figure 1**). The importance of the subglacial hydrologic system, which is largely fed by surface meltwater reaching the bed via crevasses and moulins, is also apparent in seasonal variations in glacier velocity, which tend to spike when subglacial fluids are most abundant just before seasonal integration of subglacial plumbing and major outwash events (Humphrey and Raymond, 1994). The lack of available meltwater in the accumulation area of a glacier may also contribute to the limitation in glacial erosion above the ELA (Thomson et al., 2010). All these factors make glacial erosion processes highly sensitive to subglacial hydrology, local climate, and other ambient conditions.

5.13.2.2 Glacial Erosion and Glacier Sliding

Given the complexity of glacial erosion processes, it may be surprising that net glacial erosion can be effectively modeled as a simple function of basal ice velocity. Erosion and sliding velocity have been empirically correlated on individual glaciers in Alaska (**Figure 2**), despite expected complexities including sediment storage and a lag related to periodic sediment flushing (Humphrey and Raymond, 1994; Riihimaki et al., 2005). The erosion–sliding correlation may be partly due to the fact that both are influenced by the subglacial hydrologic system; meltwater discharge affects both sediment yield and the basal pore pressure that affects glacier flow (Riihimaki et al., 2005). Nonetheless, observations imply that erosion and sliding are linearly related, with ~0.1–1 mm of erosion per meter of glacier sliding. This general correlation is supported by other field evidence. Brocklehurst and Whipple (2007) showed that ice flux, by proxy of glacier area, influences how valleys respond to changes in rock uplift rate in the Southern Alps and Northwestern Himalaya. Amundson and Iverson (2006) found that differential sliding velocity, by proxy of ice balance velocity, explains the height of most hanging valleys in mountainous western Canada. The hanging valleys appear to result from the variation in erosion rate of glaciers of different size and discharge. Thermochronometry also supports the link between erosion and glacier sliding. Berger and Spotila (2008) observed that exhumation is focused in a zone where the ELAs of numerous glaciers intersect the windward flank of the St. Elias Range, Alaska, consistent with observations in parts of the Patagonian Andes (Thomson et al., 2010). Shuster et al. (2011) also showed that thermochronometry and topography of Fiordland, New Zealand, can be explained if glaciers fully reset the landscape with erosion that scales specifically to glacier sliding velocity (as opposed to ice discharge).

There are also theoretical reasons why erosion should scale with glacier sliding. For example, ice velocity should govern the flux of abrasive particles and the forces between particles on a glacier bed, thereby relating ice velocity to abrasion rate (Hallet et al., 1979). Other factors that play a role may be of less importance to counter the effect of ice flux on many temperate glaciers. For example, ice thickness does not seem to be critical. Abrasion rates are not dependent on normal stress exerted by a glacier (Hallet et al., 1979). Likewise, given that ice flows in the direction of steepest ice-surface gradient and that steeper slopes translate to faster ice velocity, ice thickness tends to be relatively thin and of limited variation along temperate glaciers, such that the basal sliding itself remains more important for erosion (i.e., the thin-ice approximation) (Paterson, 1994; Anderson et al., 2006). Note that the relationship of sliding velocity and erosion also holds only for basal ice flux, not the net motion of a glacier. For temperate glaciers, the fraction of basal motion to surface or total motion can vary from nearly zero (e.g., frozen-based glaciers) to 100% (Paterson, 1994).

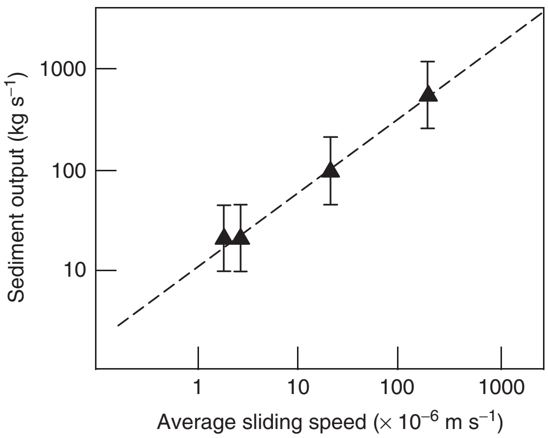

Figure 2 Empirical correlation between sediment yield and average basal sliding speed of the surging Variegated Glacier in Alaska. This correlation is the widely cited basis for the commonly assumed relationship between ice flux and erosion. Based on data of Humphrey, N.F., Raymond, C.F., 1994. Hydrology, erosion, and sediment production in a surging glacier: Variegated Glacier, Alaska, 1982–83. Journal of Glaciology 40, 539–552.

The simple approximation that glacial erosion is linearly related to basal velocity or unit ice flux has been validated in numerous numerical and analytical models that have sought to explain glacial topography with simulations of glacier flow and dynamic ice mass balance (Harbor, 1992; Braun et al., 1999; MacGregor et al., 2000; Tomkin and Braun, 2002; Tomkin and Roe, 2007; Jamieson et al., 2008; Herman and Braun, 2008; MacGregor et al., 2009). The constants in the erosion rules used by these models vary, and some models incorporate additional process-specific terms, such as the clast concentration at the bed, clast comminution, and accumulation of till. Some numerical models also include terms for internal ice deformation, dependent on variables including temperature and pore pressure, which filter out the glacier motion that does not contribute to erosion (Tomkin, 2007, 2009; Tomkin and Roe, 2007). Aside from these variations, each of these models uses a linear relationship between erosion and sliding and is able to accurately predict aspects of glacial topography, such as cross-valley or longitudinal profiles. Many of the basic characteristics of glacial topography can be explained by attaching this simple erosion rule to physics-based models of ice flow (Anderson et al., 2006; Tomkin, 2009).

An important implication for the relationship of glacial erosion and glacial sliding is that erosion should be greatest near the ELA, where ice discharge is maximum (Paterson, 1994; Meigs and Sauber, 2000). This has been shown in simple analytical models of individual glacier profiles (e.g., Anderson et al., 2006) and numerical models of entire glaciated orogens (Tomkin and Roe, 2007) (**Figure 3**). In both cases, the concentration in ice discharge and erosion at ELA are not sharp peaks, but rather broad curves that attain a maximum at the ELA. Note that there are exceptions to this result. Herman and Braun (2008) showed that realistic topography and glacier networks can result in concentrations of ice discharge that do not occur at the ELA, and thereby predict concentrations of erosion elsewhere along the glacier network. Nonetheless, this general phenomenon of erosion focusing at glacial ELA is the basis for the "glacial buzzsaw hypothesis", as discussed in Section 5.13.4.1.

5.13.2.3 Rates of Glacial Erosion

Long-standing questions about glacial erosion include how fast it can proceed, how it compares to fluvial erosion, and

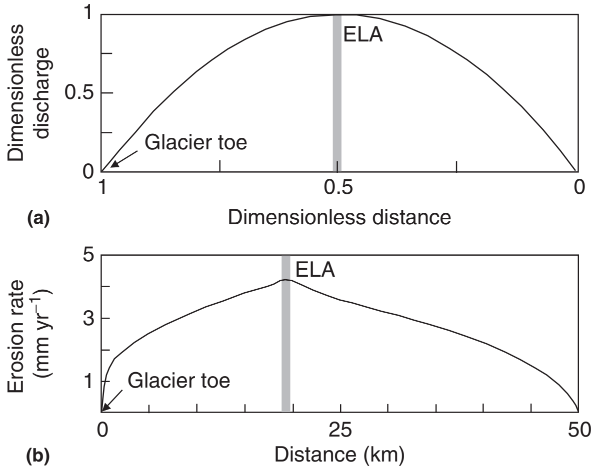

Figure 3 Modeled concentration of ice flux and erosion near the ELA. (a) Results of the analytical model of Anderson et al. (2006), showing how dimensionless ice discharge peaks at the ELA. The accumulation-area ratio of 0.5 differs from typical values ($\sim$0.65), because this simple model did not take into account down-glacier variations in glacier width. Notice how the distribution of ice discharge is a broad bell curve, rather than a sharp peak at ELA. (b) Results of the numerical model of Tomkin and Roe (2007) for evolution of a glaciated orogenic wedge (assuming an 8° taper). Erosion rate peaks at the ELA, which in this case occurs closer to the glacier toe (AAR = 0.6).

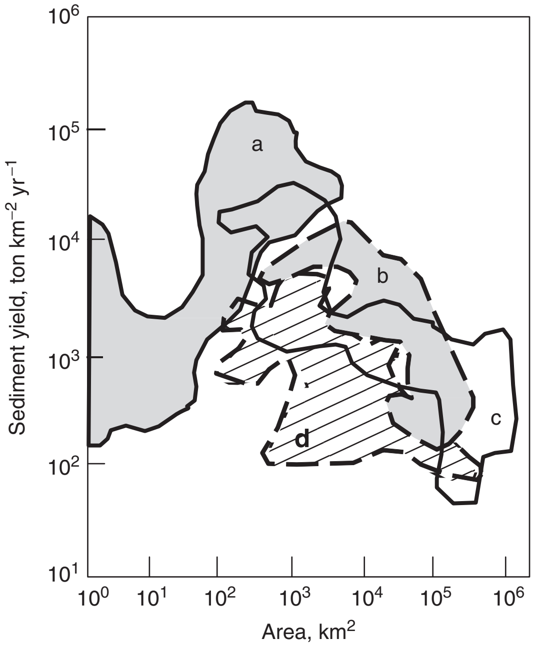

Figure 4 Summary of the results of Hallet et al. (1996) that compare short-term ($\sim$100 yr) unit sediment yield for glaciated and non-glaciated basins worldwide. Field (a) represents the area of high sediment yield for temperate glaciated basins, including both tectonically active and inactive mountain ranges. Other fields represent basins that are not heavily glaciated: (b) partially glaciated Himalayan basins, (c) fluvial basins with relief greater than 3 km from southern Asia, (d) fluvial basins with 1–3 km relief. Notice how the glaciated basins have the highest sediment yield of any basin compared. This is the oft-cited comparison that glaciers are more proficient at erosion than rivers.

how it varies over time. Comparisons of topography have led to a rough consensus that glacial erosion is more effective than fluvial erosion at sculpting landscapes (see Section 5.13.3.1). But how do rates of these processes compare?

A seminal paper promoting glaciers as more efficient eroders than rivers was Hallet et al. (1996). This study summarized data from around the world on glacial and fluvial erosion rates, based on short-term (decades to centuries) sediment yields. Glacial erosion rates were found to span four orders of magnitude (0.01–100 mm yr^{-1}, spanning polar to wet temperate glaciers) and scale with the extent of glacier coverage, but the most rapid rates of erosion observed globally occurred in areas of temperate modern glacial coverage (**Figure 4**). As a result, this paper has emerged to be a favorite citation for the quasiparadigm that glacial erosion is more proficient than fluvial erosion (e.g., this paper had been cited nearly 300 times in peer reviewed journals as of June, 2010, according to the CSA Web of Science Citation Index). A similar range and pattern of short-term erosion rates was also observed in a more recent global review (Delmas et al., 2009). However, there are limitations to using short-term sediment yields to infer erosional proficiency. Sediment yield from glaciers not only include primary bedrock erosion but also expulsion of stored sediment, making short-term sediment yield a dangerous proxy that is potentially time-dependent (Harbor and Warburton, 1993). Koppes and Hallet (2006) revised estimates of erosion rate to be lower for some of the same basins used by Hallet et al. (1996), on the basis of expulsion of stored sediment at a time of glacial retreat.

Another compilation that assessed erosion rates over different timescales suggests that more overlap exists between fluvial and glacial settings. Koppes and Montgomery (2009) study found that high rates of short-term erosion (1–10 mm yr^{-1}) occur in both fluvial and glacial basins, noting in particular the very high sediment yield of rivers in active tectonic regions including Taiwan, the Himalaya, and the Pacific Northwest. They observed that glacial erosion rates decrease as the interval of measurement lengthens (see also Spotila et al., 2004). Other studies have shown that glacial erosion rates peak during periods of glacier retreat, melting, and accelerated ice flow (Koppes et al., 2009; Delmas et al., 2009; Riihimaki et al., 2005). Glacial erosion may be unsteady through time, and glacial systems may even undergo periods of relatively limited erosion, such as during advances or glacial maxima. In contrast, fluvial erosion appears steadier through time, suggesting that long-term, mean erosion by glaciers may not exceed that by rivers. Indeed, the fastest long-term rates of exhumation observed globally, $\sim$0.5–1 cm yr^{-1}, occur in both glacial and fluvial settings (e.g., Burbank et al., 1996; Spotila et al., 2001; Ehlers et al., 2006; Berger and Spotila, 2008). However, the Koppes and Montgomery (2009) study also has limitations, including a reliance on relatively limited geographic areas for the highest fluvial erosion (e.g., Taiwan and disequilibrium volcanic areas) and the unknown effect of landuse on short-term sediment yield in fluvial areas that are likely partially cultivated. The claim that fluvial rates are time-invariant is also based on a comparison of spatially averaged short-term rates (i.e., sediment yields) to point-measurements of long-term rates (i.e., maximum exhumation rates based on thermochonometry), thus weakening the comparison.

Thus, although there is a loose consensus that glaciers are generally more effective at erosion than rivers, there are exceptions to this and it appears that, given the right conditions, either system is capable of extreme erosion rates (up to 1 cm yr^{-1}) that keep pace with rock uplift at rates determined by tectonics. The variation in glacial erosion rates over time, and the fact that no exclusive, controlled comparison of fluvial and glacial rates can be made, also make it difficult to assess which agent of erosion is more powerful. However, the fact that glaciers exist means that there are places in the world where fluvial erosion rates were not adequate to keep pace with rock uplift on their own. Likewise, the persistence of glacial topography long after deglaciation and the rapid onset of glacial topography in a glaciated fluvial landscape both suggest that glaciers are on average more proficient (Anderson et al., 2006).

5.13.2.4 Complexities and Exceptions

As alluded to above, there are numerous complications to glacial erosion, and the rule that glacial erosion is a linear function of sliding velocity should be viewed with caution and evaluated on a case-by-case basis. Unlike fluvial erosive power, which increases with slope and downstream with increasing discharge in a relatively predictable fashion, glacial erosion is more subject to local conditions that can cause spatial and temporal variation. As discussed, subglacial hydrology, including the water pressure, volume, temperature, and the integration and connectivity of subglacial plumbing, is one complicating factor that can play a major role in both glacier sliding, individual erosional processes, and the transport of sediment produced by glacial erosion that can otherwise accumulate at the bed and protect bedrock from erosion (Humphrey and Raymond, 1994; Bogen, 1996; Alley et al., 2003; Riihimaki et al., 2005). Glacial erosion is also particularly dependent on bedrock properties, including the importance of hard grains and clasts needed for abrasion and the influence of fracture spacing on quarrying (Lliboutry, 1993; Augustinus, 1992; Glasser and Ghiglione, 2009; Duhnforth et al., 2010). Patterns of ice flow, including the confluence of glacial valleys that can result in irregular variations in ice discharge along glacial profiles is another potential complication to glacial erosion that has been illustrated in models that take into account two- or three-dimensional glacial networks (Herman and Braun, 2008; Jamieson et al., 2008; MacGregor et al., 2009).

Another major factor is the ratio of total glacial motion that is accommodated by basal sliding versus ice deformation (e.g., Paterson, 1994). Glacial erosion nearly stops when the basal ice of a glacier becomes frozen to the bed, such that subglacial meltwater becomes absent and bedrock no longer 'feels' the motion of the glacier. Extremely low rates of glacial erosion have been well documented for polar glaciers and continental ice sheets, which are largely frozen-based. For example, much of Sweden exhibits preserved relict preglacial landforms that have experienced minimal erosion over multiple glacial cycles based on cosmogenic dating (Fabel et al., 2002; Hattestrand and Stroeven, 2002; Stroeven et al., 2006). Similar observations of minimal bedrock erosion have been made for the edge of the Laurentide ice sheet in North America (Colgan et al., 2002). Thomson et al. (2010) showed that frozen-based glaciers in the Patagonian Andes have likely protected an active orogen from erosion, leading to mean elevations that extend above ELA. These observations do not mean that frozen-based glaciers cannot erode at all, however. Studies of cold-based glaciers in the Allan Hills, Antarctica, show that some glacial erosion is possible, such as where the bed has significant topography and prominent outcroppings may be plucked, or where cavities on lee sides of features result in ice-bed separation and local abrasion (Atkins et al., 2002; Davies et al., 2009).

The influence of ice-bed temperature on glacial erosion can also lead to complex polythermal erosive conditions, in which local basal melting may result in localized erosion. This can lead to relief production, where wet-based glaciers promote erosion of valleys, whereas frozen-based glaciers at higher, colder altitudes inhibit summit erosion (i.e., 'selective linear erosion;' Sugden, 1978; Tomkin and Braun, 2002; Jamieson et al., 2008). This has been observed and modeled in the Torngat Mountains of northeastern Canada based on cosmogenic dating (Staiger et al., 2005). In this study, interfluves covered with felsenmeer (frost-shattered bedrock and boulders) experienced some (<2 m Ma^{-1}) glacial erosion proportional to glacier sliding, but the coefficient in the erosion rule was several orders of magnitude lower than for wet-based conditions (Staiger et al., 2005). Briner et al. (2006) similarly showed the deep fjord valleys on Baffin Island had been produced by polythermal erosive conditions. Glaciers covering interfjord plateaus were thin and frozen-based, leading to minimal erosion based on cosmogenic dating. These conditions can lead to spontaneous creation of fjords at the edge of continental ice sheets, where topographic steering of ice into rapid outlet ice streams results in thicker, warmer-based ice, and incision of deep canyons into a continental margin (Briner et al., 2006; Kessler et al., 2008). The gentler gradient of valleys relative to interfluves contributes to thicker, warmer-based ice at lower elevations, an effect that can also occur along individual glaciers, such as at overdeepenings (Jamieson et al., 2008).

The impact of polythermal ice conditions on glacial erosion illustrates the potentially nonlinear nature of glacial erosion and the importance of variables that control the thermal regime of the ice–substrate interface, including geothermal heat flux, surface temperature, ice thickness, ice velocity, topographic slope, bed morphology, strain heat release from ice deformation, hydrology, preexisting permafrost conditions, and glacier networks (Hattestrand and Stroeven, 2002; Alley et al., 2003; Staiger et al., 2005; Herman and Braun, 2008; Jamieson et al., 2008). The thermal state of the bed may also change with time, possibly explaining observations of cyclic periods of active erosion with the advance and retreat of glaciers (Delmas et al., 2009; Koppes and Montgomery, 2009). As a result, the relationship between glacier sliding and erosion should not be viewed as an absolute rule.

5.13.3 Glacial Erosion and Topography

5.13.3.1 Alpine Glacial Topography

The features of typical alpine glacial topography have been widely described in previous studies (e.g., Sugden and John,

Figure 5 Photograph of a typical alpine basin that has been shaped by glacial erosion in an tectonically active mountain range (Garnet Canyon, Teton Range, Wyoming; photo by J. Spotila). The deglaciated valley shows the characteristic U-shape. The longitudinal profile is approximately linear in the lower reaches but steepens near the headwall, where numerous irregular steps occur. Notice the steep valley walls and oversteepened peaks situated in the interfluves. The peak on the right is the Grand Teton, which is a classic example of a teflon peak. From Foster, D., Brocklehurst, S.H., Gawthorpe, R.L., 2010. Glacial-topographic interactions in the Teton Range, Wyoming. Journal of Geophysical Research 115(F01007), 20.

1976; Brocklehurst and Whipple, 2002). The classical form of a temperate, glaciated mountain range includes cirques, U-shaped cross sections, hanging valleys, low slope longitudinal profiles that are punctuated by steep steps and overdeepenings, high headwalls that rise to arêtes and horns and are flanked by truncated spurs, and hillslopes that are oversteepened beyond the angle of repose (**Figure 5**). Anderson et al. (2006) pointed out that glacial valley networks tend to have bulbous ends, and are not dendritic like fluvial networks, and that glacial topography tends to be knobby and bumpy above the ELA, but becomes smoother further down glacial valleys; a transition that might relate to basal ice conditions. Glaciated topography is clearly distinguishable from the typical V-shaped valleys and convex hillslopes of a fluvial landscape (Kirkbride and Matthews, 1997). Recent studies have quantified the magnitude of difference between glacial and fluvial topographies, using parameters such as geophysical relief, basin hypsometry, and the degree of curvature of U-shaped valleys (Kirkbride and Matthews, 1997; Brocklehurst and Whipple, 2002, 2004; Brook et al., 2006) (**Figure 6**). For example, heavily glaciated topography tends to have extensive area at low elevation (relative to the local maximum), and hence low hypsometric integral, due to focusing of erosion near ELA (Brocklehurst and Whipple, 2004). Basin hypsometry is thus a particularly useful metric for deducing the scale of former glacier coverage and its impact on topography. Studies like this illustrate that there is fundamental information about glacial erosion processes that can be learned from just topography itself.

Insight into how glacial erosion affects a landscape, and into whether glacial or fluvial erosion is more proficient in a given mountain setting, can be gained by investigating how glacial erosion modifies a formerly fluvial topography. This type of study is well suited for space–time substitution, by comparing modern drainages dominated by either process in a given location. Based on studies in the Olympic Mountains, Washington (Montgomery, 2002), the Sierra Nevada, California (Brocklehurst and Whipple, 2002), the Bitterroot Range, Montana (Naylor and Gabet, 2007), Idaho (Amerson et al., 2007), the French Alps (van der Beek and Bourbon, 2008), and the Southern Alps (Brook et al., 2006), several generalities can be drawn. Glacial erosion generally appears to remove a greater volume of rock from topography than fluvial erosion, as documented by increasing valley width, ridge-valley relief, valley cross-sectional area and the degree of curvature of the valley floor, and basin geophysical relief with increasing degree of glaciation (**Figure 7**). This suggests that glaciers can be more proficient than fluvial erosion and that glaciation can generate relief in mountain landscapes. Glaciers also respond differently to increases in rock uplift. Although fluvial profiles steepen to accommodate faster erosion as rock uplift increases, glacial profiles steepen far less, suggesting that other mechanisms must be at work to enable more rapid erosion that keeps pace with increasing rock uplift (Brocklehurst and Whipple, 2007).

The timescale at which glaciers will modify a pre-existing fluvial topography is another interesting aspect of the topographic effect of glacial erosion. Space for time substitution and comparison of fluvial and glacial topography coupled with estimates of glacial occupancy time based on oxygen isotope records have permitted estimates of the time needed for resetting fluvial topography in the Southern Alps. Results imply that recognizable glacial topography can develop in as little as 200 ka after the onset of glaciation (Kirkbride and Matthews, 1997), and that a fully glacial topography, as indicated by the curvature of cross-valley profiles, can develop in 400–600 ka (Brook et al., 2006, 2008). Numerical models of glacial topography have produced comparable estimates for the timescale of topographic resetting. Harbor (1992) showed that V-shaped valleys can be converted to U-shaped by glacial erosion in as little as 100 ka, but do not attain topographic steady-state until $\sim$500 ka. Other models similarly suggest that only one or two glacial cycles are needed to sculpt longitudinal profiles or topography that are clearly distinguishable as glacial (MacGregor et al., 2000; Jamieson et al., 2008). In contrast, the time required for fluvial erosion to recover from glaciated topography can be longer (Hobley et al., 2010). The fact that there are few transitional landscapes and that areas that were once glaciated still exhibit glacial topography, whereas fluvial landscapes do not appear to be preserved once glaciated, implies that the timescale of topographic resetting by glacial erosion is shorter than for fluvial erosion, thus reinforcing the idea that glaciers are more proficient agents of erosion (Brocklehurst and Whipple, 2002, 2004).

The observation that transition to a glacial topography corresponds to an increase in relief suggests that relief increased during the late Cenozoic due to glaciation, consistent with the hypothesis that climate change produced isostatic peak uplift in mountain ranges worldwide (Molnar and England, 1990). However, the scale of relief production due to glacial erosion may not be adequate to generate significant

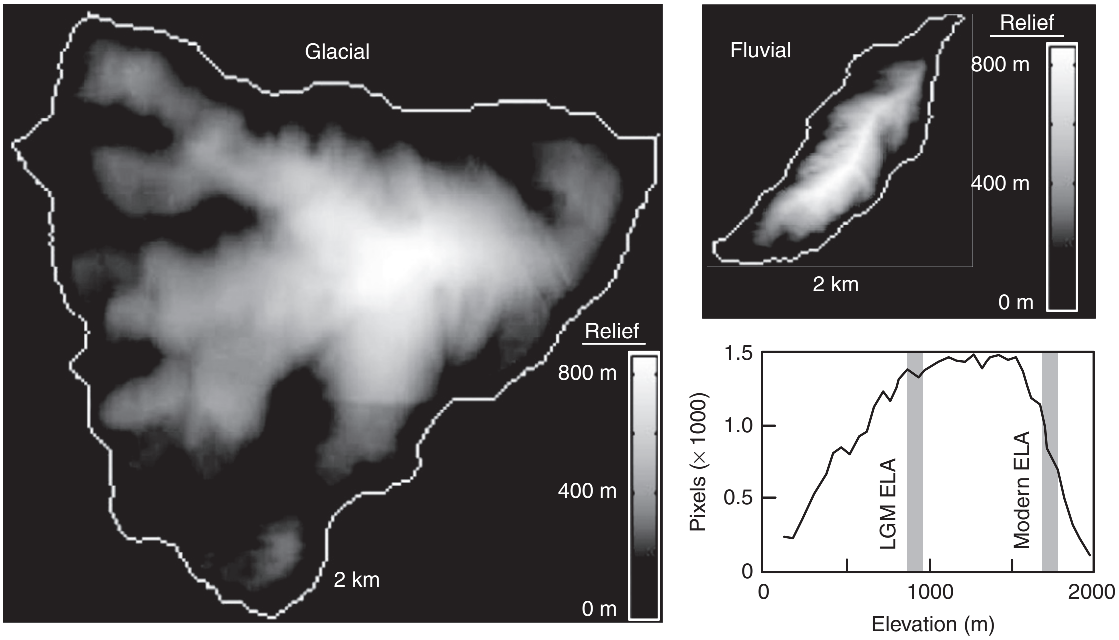

Figure 6 Typical hypsometry of glacial basins. The dark shaded images represent the distribution of sub-ridgeline relief for typical glacial and fluvial basins of the Sierra Nevada. Notice how the glacial basin has a bulls-eye pattern of relief that is focused closer to the basin outlet, and the basin itself is more equant in dimension. The diagram in the lower right shows the distribution of elevation for a glaciated basin in an active tectonic region (Southern Alps; Brocklehurst and Whipple, 2007). Note how the majority of the land area lies between modern and paleo-ELA, whereas nearly all the land area lies below the modern ELA, consistent with elements of the glacial buzzsaw hypothesis. This differs from the hyspometry of fluvial basins, which commonly display tails, in which a very small fraction of the landscape continues to high elevations that far exceed the mean. Adapted from Brocklehurst, S.H., Whipple, K.X., 2002. Glacial erosion and relief production in the Eastern Sierra Nevada, California. Geomorphology 42, 1–24; and Brocklehurst, S.H., Whipple, K.X., 2007. Response of glacial landscapes to spatial variations in rock uplift rate. Journal of Geophysical Research 112(F02035), 18 p.

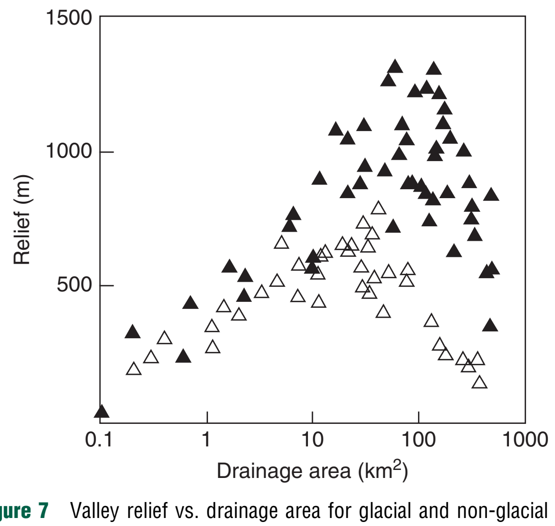

Figure 7 Valley relief vs. drainage area for glacial and non-glacial basins of the Olympic Mountains. Adapted from Montgomery, D.R., 2002. Valley formation by fluvial and glacial erosion. Geology 30, 1047–1050. Black triangles are glacial basins and open triangles are fluvial basins. The consistently higher relief of glaciated basins suggests that, at least in this case, glaciers are more proficient at removing mass and sculpting topography than rivers.

(i.e., hundreds of meters) peak uplift. Several studies have noted that the increase in relief due to glacial erosion is size-dependent, and is only significant for the largest alpine glaciers (Brocklehurst and Whipple, 2002; Montgomery, 2002; Amerson et al., 2007; Brocklehurst et al., 2008). More common smaller glaciers may thus not be as efficient at producing relief and causing isostatic peak uplift. Whipple et al. (1999) also suggested that relief production by glacial erosion is limited by the scale of ice buttressing of rock slopes and the height of hanging valleys, both of which should scale with the thickness of mountain glaciers, which is typically only a few hundred meters. Brocklehurst et al. (2008) confirmed this by finding a correlation between ice thickness at the last glacial maximum and the magnitude of relief production (for different measures of relief) for ranges in the western US (**Figure 8**). The magnitude of relief production under a glacial climate also depends on the preservation of interfluves, which is dependent on both glacial and bedrock conditions as well as the proficiency of weathering (see discussion of 'teflon peaks' below). If erosion and weathering of peaks and ridges are enhanced by climate cooling, the magnitude of relief production should be limited (Whipple et al., 1999; Brocklehurst and Whipple, 2002).

5.13.3.2 Specific Models of Evolving Glacial Topography

An effective validation of understanding of glacial erosion processes is the successful modeling of evolving glacial topography using simple to comprehensive, physically based numerical and analytical models. Considerable recent progress in this area has significantly improved and validated

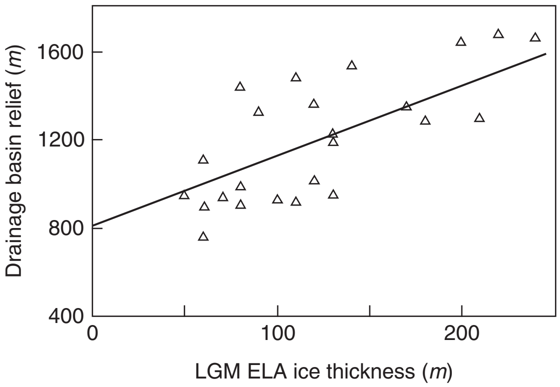

Figure 8 Empirical correlation of net drainage basin relief and glacier thickness at the last glacial maximum (LGM), based on data from the western US. Adapted from Brocklehurst, S.H., Whipple, K.X., Foster, D., 2008. Ice thickness and topographic relief in glaciated landscapes of the western US. Geomorphology 97, 35–51. This correlation implies that any relief production associated with glaciation of former fluvial valleys would also scale with ice thickness, consistent with the prediction of Whipple et al. (1999) that glaciation should produce only minimal relief production and isostatic uplift of peaks.

understanding of numerous aspects of glacial erosion and the resulting characteristic topography.

Numerical models have been successful at explaining features of glacial longitudinal profiles and why they fundamentally differ from their fluvial counterparts. Glacial longitudinal profiles exhibit lower concavity than fluvial profiles and contain extensive flat reaches that are stepped, contain sediment and water-filled overdeepenings, and tend to display junctions with tributaries that are mismatched in elevation (i.e., hanging valleys) (Oerlemans, 1984; Whipple et al., 1999; Brocklehurst and Whipple, 2002; Anderson et al., 2006). The comprehensive one-dimensional numerical models of MacGregor et al. (2000) simulated the evolution of long profiles, starting with a pre-existing fluvial profile and modifying it with glacial erosion. An ice mass-balance flow model coupled with an erosion rule that scales with ice discharge produced realistic-looking glacial long profiles, which exhibited concavity in upper reaches and were flatter in lower elevations. These models also showed that overdeepenings are produced by excess ice flux at tributary junctions and that glacial erosion is generally more effective than fluvial erosion at modifying long profiles. Anderson et al. (2006) conducted similar modeling, which showed that the peak in erosion near ELA, associated with the peak in unit ice discharge, produces the flat profiles that steepen only above the ELA. Likewise, steps and overdeepenings resulted from excess discharge, whereas hanging valleys resulted from the lower discharge of tributaries. However, their models could not reproduce realistic accumulation-area ratios (AAR $\sim$0.65), unless they implicitly assumed a typical along-profile variation in glacial valley width. This suggests that features of glacial long-profiles are dependent on not only unit ice discharge and basal ice velocity, but also the processes that control the cross-profile shape of glacial valleys.

Glacial cross-profiles have been successfully modeled as well. Harbor (1992) showed that the classical U-shape to glacial valleys can be simulated by ice flow and an erosion rule dependent on basal velocity. The key to producing this shape are lateral variations in sliding velocity, which result from the profile shape itself. Valleys initially respond to erosion by vertical incision, but this gives way to more rapid erosion at valley margins due to the high erosion rate on steep slope angles of valley walls. Harbor (1992) showed that U-shaped valleys tend to adjust to the maximum ice discharge (i.e., modern glaciers are undersized to the valleys) and develop rather quickly on pre-existing fluvial profiles, as discussed above and corroborated by other studies (Brook et al., 2006). This, along with the models of long profiles, show that there are feedbacks between deepening and widening processes that combine to result in the characteristic glacial valley form.

Another interesting aspect of glacial valley development is the lengthening of profiles by active erosion at the head (i.e., headwall and cirque migration). Empirical studies have shown that glacial headwall migration can outpace vertical glacial erosion by several times, increases with the scale of glaciation, and can outpace that of opposing fluvial valleys (Brocklehurst and Whipple, 2002; Oskin and Burbank, 2005; Naylor and Gabet, 2007; Shuster et al., 2011). These studies suggest that headwall retreat is a fundamental aspect of glacial modification of valley topography (**Figure 9**). Some studies suggest that this is the primary form of relief production in valleys due to onset of glaciation (Whipple et al., 1999; Brocklehurst and Whipple, 2002). Given that headwalls are situated above the ELA, it is not immediately clear how glacial erosion can accomplish rapid headwall retreat. MacGregor et al. (2009) performed comprehensive numerical simulations of headwall backwearing to explore how this occurs. Their model incorporated terms for blowing and avalanching snow, freeze-thaw weathering of the wall, and excessive influx of abrasion tools at glacier heads. Even without taking into account rotational ice flow in cirques or complex hydrology that might result in polythermal ice conditions in cirques (e.g., Hooke, 1991), their models successfully showed that headwall-specific processes can produce backwearing (**Figure 10**). Cirques did form at the heads of long profiles and caused rapid retreat of headwalls, primarily due to the large influx of snow from adjacent slopes, focusing of precipitation by the evolving topography, effective quarrying of fractured bedrock, rapid ice sliding that will occur on the steep cirque bed, abundant physical weathering of exposed bedrock and the resulting availability of erosion tools (MacGregor et al., 2009). These results show that cirque formation and headwall retreat are fundamental aspects of glacial erosion and can be simulated with physically based models of ice flow, precipitation, and glacial erosion. Headwall retreat will not be effective under all conditions, however, and Brocklehurst and Whipple (2007) showed that it can be outpaced by vertical glacial incision where rock uplift rates are rapid.

Comprehensive surface process models, which take into account multiple erosional processes, tectonic uplift, and climate, have also been proven effective at simulating the evolution of glacial topography. Braun et al. (1999) simulated fluvial, hillslope, and glacial erosion under various conditions, and showed that glacial erosion does excavate a larger volume of rock than fluvial conditions and that the volume of ice that can be supported by the landscape is dependent on both elevation and the glacial topography itself. Their results also showed that glacial erosion peaks near the end of glacial maxima and that erosion in general is most rapid under conditions of disequilibrium, such as when climate is changing. Their model also showed that glacial erosion can reach a steady state, where the rates do not increase despite increase in ice volume. Tomkin and Braun (2002) used a similar comprehensive modeling approach to test specifically for the magnitude of relief generated by glacial conditions. Their results showed that relief production depends on climate; relief increases when glaciers are cold-based at high elevations, but decreases under other conditions. Other models have successfully shown that realistic two- and three-dimensional glacial topography can be generated from simple glacial erosion rules and basic ice dynamics (Tomkin, 2009). When these models begin with real topography, they show that glacial erosion can actually be quite complex, spatially variable, and dynamic; unlikely to develop a steady-state (Herman and Braun, 2008). The power of these cutting-edge surface process models is that they take into account so many processes and conditions, as well as the coupling between them. For example, the MacGregor et al. (2009) model takes into account clast production, comminution, sediment transport, flexure and uplift, physical weathering, hydrology, climate variations, orographic precipitation based on the evolving topography, and, like all models, ice dynamics and glacial erosion.

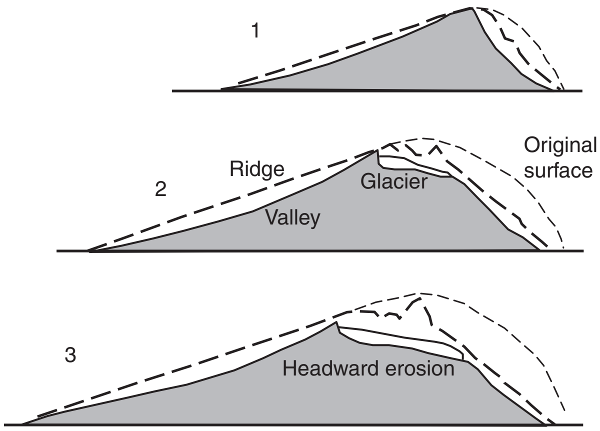

Figure 9 Conceptual model of glacial landscape evolution in an actively uplifting mountain range that is dominated by headwall retreat (redrafted from Oskin and Burbank, 2005). The three-step model shows sequential uplift of a range. The original surface is denoted (dashed line), as inferred for the Kyrgyz Range based on a preserved unconformity. Notice how the onset of glaciation (i.e., when the topography reaches the ELA) results in limit to the height of the range. As more topography pushes above the ELA, glacial cirques retreat by headwall erosion, pushing the divide far away from the range crest (shown by the short dashed lines).

5.13.4 Influence of Glaciers on Tectonics

5.13.4.1 The Glacial Buzzsaw Hypothesis

The observation that glacial erosion is highly proficient and characteristically more effective than fluvial erosion has led to the development of the idea that glaciers are capable of eroding as fast as topography can be rebuilt via rock uplift. In the simplest terms, the 'glacial buzzsaw' hypothesis posits that glacial erosion is able to limit topography under appropriate conditions (where topography generically means elevated land area, or mountain growth, and not necessarily a specific topographic metric, such as mean elevation or relief). Obviously, this would have significant implications for landscape evolution and orogenic systems if valid, and as a result this idea has been the focus of intense study over the past few years.

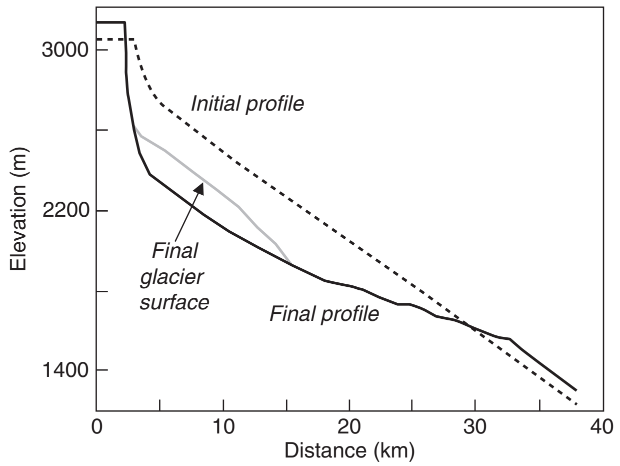

Figure 10 Results of a numerical simulation of longitudinal valley evolution of a glaciated basin. Adapted from MacGregor, K.R., Anderson, R.A., Waddington, E.D., 2009. Numerical modeling of glacial erosion and headwall processes in alpine valleys. Geomorphology 103, 189–204. Note how glacial erosion results in lowering of an initial profile (dotted line), resulting in a somewhat linear glacial profile at lower reaches and an oversteepened profile at the glacier headwall. This oversteepening is what enables cirque retreat. Also note how the final glacier surface itself is small; this is because glacial erosion has outpaced rock uplift and has thus been self-limiting. Model based on variable-temperature simulation and assumed erosion to be linear with glacier sliding.

The buzzsaw hypothesis was developed by Brozovic et al. (1997), on the basis of topographic trends in the Northwest Himalaya and Karakoram. (Note that the label of 'buzzsaw' does not actually occur in their paper, but was linked to this study via an abstract by the same authors.) Hypsometry for these actively uplifting areas showed peaks in elevation distribution and minima in slope distributions at snowlines, or the mean Quaternary ELA (**Figure 11**). The mean topography paralleled a rise in ELA across the ranges, but did not correlate to significant spatial gradients in tectonic rock uplift. Brozovic et al. (1997) interpreted this to be the result of glacial and periglacial processes placing an upper limit on altitude and relief, and suggested that this resulted from the proficiency of glacial erosion and its tendency to peak at ELA. Their idea was that glaciers concentrate erosion at ELA and keep pace with tectonic uplift, preventing the majority of topography from rising above mean snowlines. If too much topography rises above ELA, more glaciers form, and thus erosion accelerates. Similarly, a negative feedback loop exists, by which too much glacial erosion undercuts a glacier, resulting in less-erosive fluvial conditions that would in turn allow topography to rise to the ELA. Evidence that glaciers undercut themselves in such a self-defeating way includes the outcomes of numerical simulations (MacGregor et al., 2000; Egholm et al., 2009) and empirical observations of reduction in glacier size despite climate change fostering glaciation in the Patagonian Andes (Kaplan et al., 2009). The buzzsaw model also holds that glacial topography itself reinforces the buzzsaw, in that isolated peaks (i.e., a few percent of the landscape), or topographic ''lightening rods'', would rise high above ELA and focus precipitation and avalanching snow back onto glacial valleys, keeping the vast majority of the landscape near the mean ELA. Glacial erosion may also inhibit fluvial erosion downstream, by overloading rivers with sediment (Whipple et al., 1999; Korup and Montgomery, 2008).

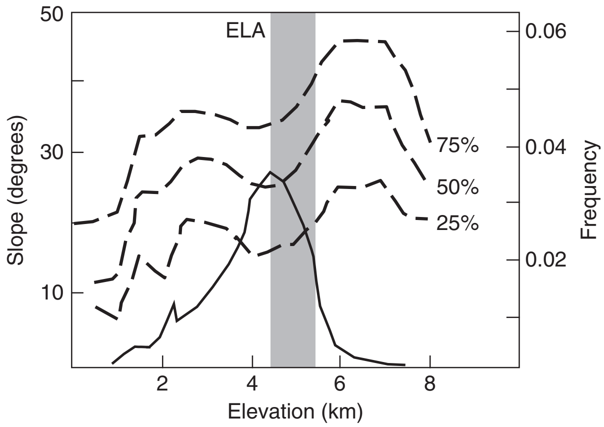

Figure 11 Topography of the northwest Himalaya that formed the empirical basis for the glacial buzzsaw hypothesis. Adapted from Brozovic, N., Burbank, D.W., Meigs, A.J., 1997. Climatic limits on landscape development in the northwestern Himalaya. Science 276, 571–574. The frequency of elevation for their entire study area is shown by the solid dark line. Distributions of slope, for the 25th, 50th, and 75th percentiles, are shown in the dashed lines. Notice how elevation peaks near the glacial ELA, which also corresponds to minima in all slope percentiles. Very little topography extends above the ELA, all of which has very high slope.

The topographic trends that inspired this concept had been observed earlier, such as by Porter (1981). Since the buzzsaw hypothesis was posed, other studies have recognized similar topographic trends in other locations. These include the Cascade Range, Washington (Mitchell and Montgomery, 2006), where mean swath topography clearly parallels the rise in modern glaciers and paleo-ELA across the range (**Figure 12**), but does not reflect a ten-fold variation in rock uplift rate. Other locations where this has been observed include the Andes (Montgomery et al., 2001; Thomson, 2002), the northern Basin and Range, where the glaciers are actually quite small (Foster et al., 2008), northwest Tibet (Seong et al., 2009), and the European Alps (Anders et al., 2010). Anders et al. (2010) added the idea that cirques limit the elevation of adjacent peaks, such that even peaks that rise above the ELA will still be limited and show concordance to the ELA. Egholm et al. (2009) capped this all with a global topographic

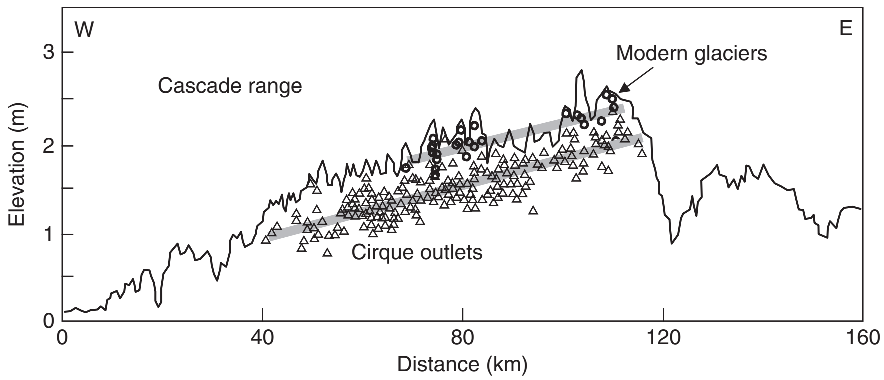

Figure 12 Correlation of mean swath topography and glacial ELA in the Cascade Range, Washington, consistent with the predictions of the glacial buzzsaw hypothesis. Adapted from Mitchell, S.G., Montgomery, D.R., 2006. Influence of a glacial buzzsaw on the height and morphology of the Cascade Range in central Washington State, USA. Quaternary Research 65, 96–107. The swath topography (their swath #2) runs from west to east across the range. The elevation of modern glaciers (circles) and deglaciated cirque outlets (triangles) are shown, with gray bars representing the regressions of both trends. Notice how the topography parallels both of these trends, although is situated closer to the elevation of modern glaciers. This implies glaciers limit topography, and that the mean topography only extends a fixed height above the local LGM ELA.

analysis, which showed that there's almost no surface area on the continents that rises above modern snowlines, regardless of tectonic environment. They showed that hypsometric maxima parallel ELA globally, although this is slightly different from as defined in the original buzzsaw hypothesis. Rather than strictly limiting topography to ELA, the Egholm et al. (2009) analysis identified clustering of topography at ELA. The global analysis also revealed that peaks are generally less than 1.5 km above the modern snowlines, consistent with Anders et al. (2010) finding in the Alps.

In addition to meeting the general topographic predictions, there is additional supporting evidence for the buzzsaw hypothesis. Several studies have shown that sharp gradients in tectonic rock uplift appear to have little effect on regional topography in glacially dominated areas, where topographic steady-states have been invoked, such as in the Patagonian Andes (Thomson, 2002; Thomson et al., 2010) and in the St. Elias orogen (Meigs and Sauber, 2000; Spotila et al., 2004; Berger and Spotila, 2008). Note that similar observations have also been made in nonglaciated areas, however (Spotila et al., 2001). Other studies have also found that the influence of ELA's position in a landscape is evident even in locations which do not meet the strict requirement that ELA limits all topography, such as in the Sierra Nevada, California (Brocklehurst and Whipple, 2004). Models have also confirmed that glaciers may be able to keep pace with rock uplift under certain conditions (e.g., Egholm et al., 2009). For example, the numerical surface process models of Tomkin (2007, 2009) showed that there is climate-dependent elevation lowering, similar to that predicted by the buzzsaw hypothesis.

There are exceptions and limitations to the buzzsaw hypothesis, however. First, it is important to point out that not all active orogens reach glaciated elevations, indicating that fluvial erosion can be adequate to erode topography before the invocation of the glacial buzzsaw. Another potential problem is circularity, given that the ELA itself is naturally influenced by local topography and precipitation, so a correlation between ELA and topography could be the result of a different causality (Østrem, 1972; Porter, 1977; Meigs and Sauber, 2000). There are also some mountain ranges that do not fit with the predictions of the buzzsaw hypothesis, such as the Southern Alps, where elevation, relief, uplift rate all correlate with distance from the Alpine fault, but not the ELA (Kirkbride and Mathews, 1997; Brocklehurst and Whipple, 2007) (**Figure 13**). Others have suggested that vigorous glacial erosion can actually increase relief in active orogens, rather than limit it (Brook et al., 2008). Other possible flaws with the buzzsaw hypothesis include potential limits to glacial erosion, as have been observed in comprehensive surface process models (Braun et al., 1999) and proposed based on the mechanics of specific processes (Lliboutry, 1993; Alley et al., 2003; Staiger et al., 2005; Jamieson et al., 2008).

Another exception to the idea that glaciers limit topography, which was actually permitted in the original formulation of the hypothesis (Brozovic et al., 1997), is that isolated peaks may rise high above ELA. For example, Meigs and Sauber (2000) showed that despite a mean elevation of 1225 m above sea level (asl) in the St. Elias range, Alaska, about half of the landscape rises above ELA, and a tiny fraction of the landscape attained heights as great as 6050 m asl; much higher than ELA. This has led to the development of the idea of 'teflon peaks,' which are conceptually similar to the 'topographic lightening rods' in the original Brozovic et al. (1997) model. Teflon peaks, as proposed by Anderson (2005) and developed further by Foster et al. (2008), rise above the snowline because they are shielded from glacial erosion by frozen basal ice conditions (akin to selective linear erosion; Sugden, 1978; Thomson et al., 2010), because of the inability of snow and ice to accumulate on the steep bedrock slopes (hence like teflon; precipitation does not stick but continues to fall to the valley below), and because they replace valleys and space-out cirques and thus reduce glacier drainage density (Brocklehurst and Whipple, 2007). Teflon peaks will also accelerate erosion in valleys below, by focusing precipitation and avalanching, providing tools for erosion into the bergschrund, and insulating the valley glacier from ablation by a blanket of rock fall debris (Brozovic et al., 1997; Foster et al., 2008). The height of teflon peaks should depend on bedrock material properties and the original spacing of large valleys (Foster et al., 2010). A prevalence of such peaks may indicate that the glacial buzzsaw is locally defeated, such as where glaciers are small and rock uplift rates are rapid (e.g., Teton Range; Foster et al., 2008) (**Figure 5**). Teflon peaks are not ubiquitous features in glacial topography, however. Anders et al. (2010) and Egholm et al. (2009) showed that peak elevations are commonly limited by the

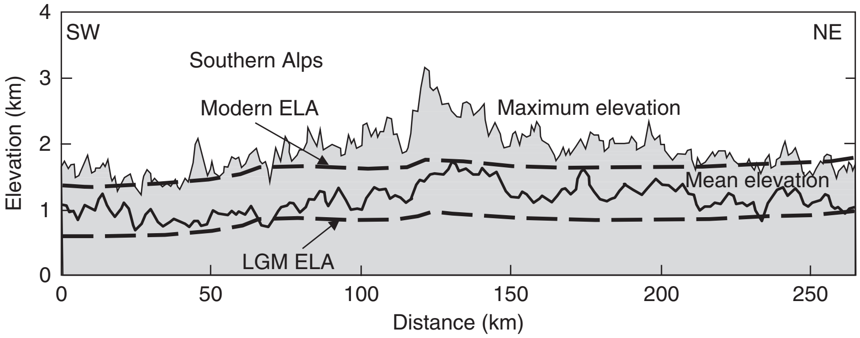

Figure 13 An example of an actively uplifting mountain range that violates the glacial buzzsaw hypothesis: the Southern Alps, New Zealand. Adapted from Brocklehurst, S.H., Whipple, K.X., 2007. Response of glacial landscapes to spatial variations in rock uplift rate. Journal of Geophysical Research 112(F02035), 18 p. Maximum and mean elevation from a swath along the trend of the central Southern Alps, parallel to the Alpine fault, do not correlate with ELA (modern or LGM). The rise in elevation that bears no relationship to ELA is instead linked to a local increase in rock uplift rate.

altitude of cirques, akin to a glacial base level, which can effectively prevent the unlimited growth of teflon peaks. Extremely high teflon peaks may require high rates of tectonic rock uplift, where erosion of glacial headwalls may not be able to keep pace, thereby resulting in the growth of extremely high headwalls and overall steepening of basins (Brocklehurst and Whipple, 2007).

Based on these exceptions, it is clear that there is no simple, absolute glacial buzzsaw that will limit topographic growth under all conditions. In some cases, glaciers appear to be quite capable at erosion and do seem to limit topography. Glaciers also naturally concentrate erosion at ELA. However, it is clear that glaciers can be relatively ineffective at erosion under some conditions, and in certain cases will not be capable of keeping pace with rock uplift. This leads to an interesting idea that the glacial buzzsaw may be scale dependent and only activates after a certain threshold of glaciation has been reached. Small glaciers may not have the discharge to produce adequate sliding that can erode as rapidly as needed (e.g., Kirkbride and Matthews, 1997; Amundson and Iverson, 2006). Topography that includes numerous large teflon peaks and inherently small valleys may thus be poorly predisposed for development of a glacial buzzsaw (Foster et al., 2008). Foster et al. (2010) suggested that only basins $>20\ \mathrm{km}^2$ in the Teton Range were capable of hosting glaciers large enough that could keep pace with rock uplift. Likewise, increasing rates of tectonic rock uplift may require larger glaciers for the glacial buzzsaw to be in effect, suggesting that glacial erosion and the evolution of glacial topography also depend on rates of tectonic uplift (Brocklehurst and Whipple, 2007; Foster et al., 2008).

5.13.4.2 Glacial Erosion and Climate–Tectonic Coupling

An effective glacial buzzsaw, by definition, should lead to steady-state topography and steady-state flux in orogenic systems. This means that the rate of erosional efflux from a glaciated orogen could readily equate to the rate of tectonic influx, thereby precluding tectonic accretion. Given the importance of a steady-state flux for the behavior of orogenic wedges (Whipple and Meade, 2004), this implies that glacial erosion, and the climatic conditions that enable it, have significant geodynamic implications for mountain building throughout the geologic record. Atmospheric temperature, not just precipitation, may accordingly be a controlling factor in collisional tectonics (Tomkin and Roe, 2007).

The effect of erosion on orogenic systems has been explored through recent coupled process models. Numerous analytical and numerical models have shown the importance of erosion and the degree of coupling between surface processes and tectonics in orogenic wedges (Beaumont et al., 1992; Koons, 1995; Willett, 1999). In an orogenic wedge, an increase in erosional efficiency will cause narrowing and lowering of the wedge relief and an increase in rock uplift rate (Whipple and Meade, 2004; Roe et al., 2006). Orogen size and dynamics can thus be controlled by climate. Spatially nonuniform erosion in these systems, such as should occur at glacial ELA, can also significantly alter deformation pattern and pressure–temperature–time paths during exhumation

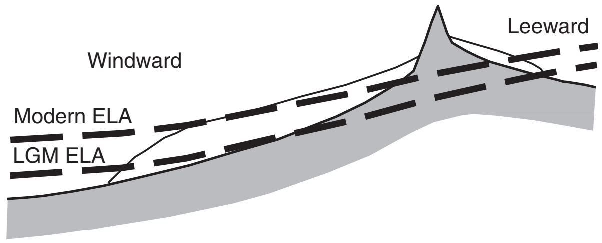

Figure 14 Conceptual diagram of the effect of orographic precipitation on the size of glaciers and ELA of a coastal mountain belt. Adapted from Meigs, A., Sauber, J., 2000. Southern Alaska as an example of the long-term consequences of mountain building under the influence of glaciers. Quaternary Science Reviews 19, 1543–1562. The windward face of the range traps more precipitation, resulting in larger glaciers. The decrease in precipitation away from the coast results in a rise in both modern and LGM ELA and results in smaller glaciers on the leeward side of the range. If glaciers influence mean topography, orographically-induced asymmetry in glacier distribution will have a major effect on the evolution of mountain ranges.

(Whipple and Meade, 2004). Modeling of a glaciated orogenic wedge suggests that orogen-scale exhumation will become focused at and just below the mean ELA (Tomkin and Roe, 2007; Tomkin, 2007). This differs from fluvially dominated orogens, in which rock uplift is focused near the core of the range. Tomkin and Roe (2007) also showed that the dimensions of glaciated wedges are more sensitive to climatic fluctuations, such as precipitation rate, than fluvial wedges. Other numerical models show that glacial erosion can be more important in eroding orogenic wedges than fluvial erosion. Tomkin (2007) showed that total orogenic erosion is linearly proportional to the degree of glacier coverage, and explored this result for four test cases of real world glaciated orogens. The results were broadly consistent with the glacial buzzsaw effecting rapid erosion, implying that the onset of glaciation in the late Cenozoic increased rates of rock uplift. Other models have shown that the effect of glacial erosion on orogenic evolution can actually be complex and dynamic (Herman and Braun, 2008; Tomkin, 2009).

Case studies of real world glaciated orogens have also shown the impact of glacial erosion on tectonic processes. Work in southern Alaska's St. Elias orogen has shown that the long-term exhumation is focused on the windward side of the critical wedge, where precipitation is greater and glaciers are larger (Meigs and Sauber, 2000; Spotila et al., 2004) (**Figure 14**). Moreover, rapid (i.e., $5\ \mathrm{mm\ yr^{-1}}$) exhumation appears to be focused where the mean Quaternary ELA intersects the windward flank, termed the 'ELA front,' matching the predictions of numerical models (Berger and Spotila, 2008; Tomkin and Roe, 2007) (**Figure 15**). Exhumation rates along this zone of focused denudation do not vary with the size (surface area) of glaciers draining across it, however, suggesting a local violation of the rule that glacier erosion rate correlates to ice discharge. This may indicate that small glaciers are able to keep pace with rock uplift by cutting down to the 'base level' established by larger glaciers, such that tectonics ultimately sets rock uplift rates here, or that erosion by large outlet glaciers is somehow limited, perhaps by overdeepenings (Berger and Spotila, 2008). Although

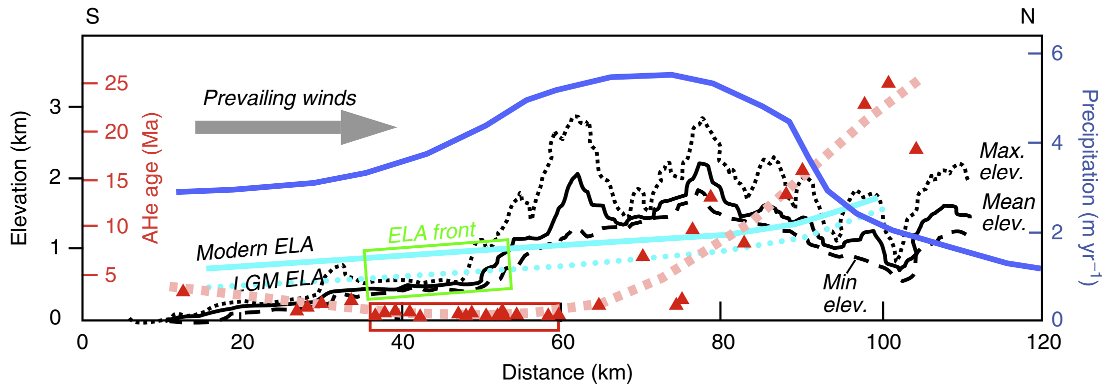

Figure 15 Relationship of the spatial distribution of exhumation and ELA in the St. Elias orogen, Alaska. Adapted from Berger, A.L., Spotila, J.A., 2008. Denudation and deformation of a glaciated orogenic wedge: The St. Elias orogen Alaska. Geology 36, 523–526. Exhumation is represented by the apatite (U–Th)/He ages (or AHe ages, in Ma), where younger ages mean more rapid exhumation. Note how exhumation is most rapid (youngest ages; red box) where the modern and LGM ELA intersect the windward flank of the orogen (the 'ELA front'; blue box). In contrast, exhumation does not correlate with the broad distribution of precipitation, topography (maximum, minimum, and mean swath topography shown), or distribution of convergent structures (not shown). This is consistent with ELA focusing long-term glacial erosion and for the idea that glaciers can act as excavators in dictating patterns of denudation and strain in orogens.

thermochronometry of bedrock suggests that long-term exhumation rates do not peak near massive outlet glaciers that cross the deformation front (e.g., the Bering glacier), detrital dating by Enkelmann et al. (2009) suggests that more rapid exhumation may occur beneath the largest glaciers, where bedrock cannot be sampled (e.g., the Seward glacier). Work in the St. Elias orogen has also suggested that glacial climate forced a change in the dynamics of the entire wedge. Differences in long-term bedrock cooling rates from different thermochronometers show that a sharp acceleration of cooling occurred not at the onset of glaciation but at ~1 Ma, corresponding to a shift in the pattern of glaciation associated with the change in Milkovitch forcing from 41 to 100 ka cycles (Berger et al., 2008a; Ruddiman et al., 1986). Berger et al. (2008a, b) speculated that this change forced the orogen to narrow and reorganize, including onset of backthrusting (i.e., transition to a two-sided wedge) and termination of distal structures via sediment burial. The St. Elias orogen thus provides an exceptional case study of the effects of long-term glacial conditions on the evolution of a collisional mountain belt. The degree to which glaciers act as orogen-scale 'excavators' in controlling the pattern of erosional efflux from this orogen implies a high degree of coupling between erosion and tectonics in glaciated collision belts. Other locations where similar inferences have been made include the Andes, Southern Alps, Fiordland (New Zealand), and Olympic Mountains (Thomson, 2002; Thomson et al., 2010; Tomkin, 2007; Shuster et al., 2011; Tomkin and Roe, 2007).

Like the interaction of glaciers and topography, the interaction of tectonics and climate under glacial conditions is complex and dependent on the nature of coupling between numerous processes and conditions. At the scale of individual glacial valleys, glaciers (and therefore climate, by way of ELA) control erosion and shape topography, but are in turn affected by erosive processes and topography itself, for example, shading, effect of rock debris cover on glacier albedo, input of erosion tools into the bergschrund, topographic steering of orographic precipitation, and blowing and avalanching snow (Foster et al., 2008, 2010; Anders et al., 2008, 2010; MacGregor et al., 2009). At the orogenic scale, glaciers are capable of focusing denudation and influencing deformation and orogen architecture, but in turn are heavily dependent on the rates of rock uplift set tectonically (Berger and Spotila, 2008). Rock uplift rate itself appears to be an important parameter in determining how glacial erosion will affect a landscape and how landscapes will react to climate change (Berger and Spotila, 2008; Tomkin and Roe, 2007; Tomkin, 2009; Koppes and Montgomery, 2009). Similarly, ice-mass balance and its related characteristics (e.g., glacier thickness) appear to be influenced by orogenic structure and evolution (Tomkin, 2003). Thus, the climate–erosion–tectonics system, when glacially dominated, is dynamic, and will evolve in a complex fashion that is dependent on ambient conditions and resistant to the imposition of absolute generalities.

5.13.4.3 Orogen and Landscape Response to Glacial Climate Change

It has long been recognized that changes in glacial cover due to climate change can modify landscapes and result in rock uplift. The viscoelastic, isostatic response of a landscape to changes in ice volume has been well documented in areas of recent deglaciation, such as the postglacial surface uplift in Fennoscandia or the Alps (Gudmundsson, 1999; Bungum et al., 2010; Barletta et al., 2006). Glacial retreat in the past century in the fjords of southern Alaska, in the vicinity of Glacier Bay, has likewise resulted in spectacular (~3 cm yr^{-1}) rates of surface uplift. Deglaciation also has the important effect of destabilizing landscapes. Following ice retreat, hillslope erosion increases due to removal of ice buttresses, whereas fluvial networks must adjust and modify the glacial topography and convey large accumulations of liberated glacial sediment (e.g., Meigs and Sauber, 2000). Major changes in glacial cover thus result in substantial reorganization and adjustment of the entire erosion system.

The onset of glaciation in an orogen has also been recognized as having significant geodynamic and geomorphic influence. Several studies have identified accelerations in exhumation rates related to the onset of glaciation, including in the Coast Ranges in British Columbia (Farley et al., 2001; Ehlers et al., 2006; Densmore et al., 2007) and the Svalbard islands in the Arctic (Blythe and Kleinspehn, 1998). Changes in glacial coverage, such as resulting from the change from 41 to 100 ka Milankovitch periodicity (Ruddiman et al., 1986) at $\sim$1 Ma or other subordinate climate changes, have also been linked to major (i.e., more than ten-fold) increases in denudation rates, such as in the Coast Ranges, southern Alaska, the European Alps, Himalaya, and Fennoscandia (Shuster et al., 2005; Haeuselmann et al., 2007; Berger et al., 2008a; Rahaman et al., 2009; Dowdeswell et al., 2010). These increases in denudation should correspond to concomitant increases in sedimentation in adjacent basins. Such changes are also consistent with the hypothesis that late Cenozoic atmospheric cooling enhanced erosion and so influenced mountain building and tectonics on a global scale (Molnar and England, 1990).

At the center of the original Molnar and England (1990) chicken–egg hypothesis for the relationship of climate and tectonic change is the prediction that a transition to glacially dominated erosion can increase topographic incision (i.e., increase topographic relief and decrease mean elevation) and thereby result in isostatic (Airy) uplift of peaks. Although the lithosphere's response to a change in crustal volume associated with incision-related unloading can theoretically be significant (Montgomery, 1994), the magnitude of uplift in real-world cases is limited by the narrow width of mountain belts relative to typical flexural wavelengths and the magnitude of relief produced by onset of glacial conditions (Small and Anderson, 1998; Pelletier, 2004; Cederbom et al., 2004; Champagnac et al., 2009). In most cases, the magnitude of relief production and peak uplift predicted for actual mountain ranges are limited to a few hundred meters (Small and Anderson, 1995; Whipple et al., 1999; Brocklehurst and Whipple, 2002; Brocklehurst et al., 2008; van der Beek and Bourbon, 2008). The contribution of peak height resulting from relief production and glacioisostatic compensation is increased, however, under polythermal ice conditions typical of ice-streams at the margins of polar ice sheets (Kaplan and Miller, 2003; Stern et al., 2005; Medvedev et al., 2008).

An interesting issue related to the response of a landscape or orogen to a change in glacial climate is the timescale at which these systems react. How long did it take to adjust and find a new equilibrium following the transition to the glacial climate of the late Cenozoic? Are response times short enough that landscapes will adjust to climate change over individual ice ages (e.g., 100 ka)? Or, are response times so sluggish that glaciated landscapes and orogens never truly attain steady state? As discussed above, the timescale at which glacial erosion can modify a pre-existing fluvial topography is on the order of several hundred thousand years (Harbor, 1992; Kirkbride and Matthews, 1997; MacGregor et al., 2000; Brook et al., 2006, 2008; Jamieson et al., 2008). This is relatively rapid (i.e., glacial topography can be manifest after only one or two glacial cycles), but implies that landscapes will not attain states of equilibrium at the periodicity of individual ice ages (Herman and Braun, 2008; Tomkin, 2009). Tomkin and Roe (2007) showed that the timescale of response of an orogenic wedge to glaciation is longer. Their numerical simulations suggest that the dimensions and fluxes of orogenic wedges will adjust to onset of glaciation, modeled as a step function change in erosivity, to attain a new steady state only after a few million years. This means that short term climate cycles will be averaged out and not individually ascertainable in the dynamic response of orogenic wedges. That orogens do not respond instantaneously also suggests that interpretations of sudden change in exhumation rate due to specific climate changes observed in case studies should be taken with caution (e.g., Berger et al., 2008a).

Although changes in glacial climate can have major impact on landscape and orogenic systems, not every change need do so. There is some indication that the effect of glacial erosion on topography and orogenic architecture may be dependent on thresholds, such that only climate change of a certain magnitude will cause noticeable topographic or tectonic modification. For example, studies of alpine glaciers have indicated that there is a threshold of glaciation (ice coverage, ice flux, ice sheet integration, size of glacial valleys, etc.) that must be met to significantly affect topography and effect the glacial buzzsaw; climate change resulting in small glaciers may not be sufficient (Brocklehurst and Whipple, 2004; Egholm et al., 2009; Foster et al., 2008, 2010). Brocklehurst and Whipple (2007) indicated that glaciers needed to be 30–100 km^2 in order to be capable of incising rapidly and keeping pace with tectonic uplift. This is consistent with the correlation between glacial erosion and degree of glacier coverage (Hallet, 1996; Montgomery, 2002). Examples of orogens that experienced changes in denudation or architecture attributed to glacial erosion that lagged behind the inception of glaciations also suggest that there is a threshold of glacial erosion that must be met before an erosion system is reorganized (e.g., Shuster et al., 2005; Haeuselmann et al., 2007; Berger et al., 2008a; Dowdeswell et al., 2010).

5.13.5 Discussions and Conclusions

Two contradictory themes for the influence of glacial erosion on topography and mountain building have emerged from this review. In many respects, glacial erosion appears complex, so as to defy the application of simple rules or predictions (Lliboutry, 1993). How glaciers erode depends on many conditions, including the ice mass balance, basal thermal conditions, subglacial hydrology, slope and morphology of the bed, headwall processes, rock uplift rate, and bedrock properties, many of which are coupled to erosion processes and resulting topography (Herman and Braun, 2008; MacGregor et al., 2009). In this sense, glacial erosion appears to be a nonlinear, dynamic system. In contrast, many aspects of glacial erosion and the resulting topography can be relatively simply explained, by assuming that glacial erosion scales linearly with the basal ice velocity or unit ice flux (Anderson et al., 2006). Using relatively simple erosion rules, models have demonstrated how numerous aspects of glacial topography evolve, such as cross-valley profiles (Harbor, 1992). At the same time, complex models that take into account numerous processes and conditions are required to explain other phenomena, such as cirque retreat (MacGregor et al., 2009). This implies that while a very rough approach can explain some first-order aspects of glacial erosion, a comprehensive understanding that is transportable to variable environments requires a much more detailed analysis of specific local conditions.

Glacial erosion is clearly an effective means of denuding landscapes. Glacial erosion rates can be extremely rapid, and there are many indications that glacial erosion is commonly more effective than fluvial erosion. Glacial erosion generally excavates a greater volume of rock from topography than is observed in fluvial landscapes and onset of glaciation can generate moderate relief (Montomery, 2002; Brocklehurst and Whipple, 2002). This is consistent with the idea that late Cenozoic cooling affected erosion, topography, and sedimentation in the many of the world's orogens (Molnar and England, 1990). However, both forms of erosion appear to 'max-out' at similar magnitudes, suggesting they are ultimately trumped by geodynamically limited rock uplift rates (Koppes and Montgomery, 2009). Glacial erosion also appears to be considerably more variable in time than fluvial erosion. In particular, glacial erosion seems particularly effective during periods of melting or ice retreat (Riihimaki et al., 2005), and in general erosion seems to accelerate during periods of climate or landscape transition. In some respects, however, the question of whether glacial erosion is more proficient than fluvial erosion is contrived and moot; both forms of erosion will depend on ambient conditions and, by definition, the two forms of erosion cannot codominate under the same conditions. That the glacial buzzsaw hypothesis has been validated in numerous (but not all) settingsz does imply that glacial erosion has significant geodynamic importance, but beyond this, comparisons between glacial and fluvial erosion seeking a champion can be a distraction from more fundamental process-based questions that need to be addressed to advance our understanding of glacial erosion itself.

Like fluvial erosion, glacial erosion strongly influences tectonics, including dictating patterns of denudation and deformation in orogenic systems (Spotila et al., 2004; Berger et al., 2008b; Berger and Spotila, 2008). There are important differences in how these systems function, however, such as the concentration of erosion midway along glaciers, the greater penetration of orographic precipitation into mountains during glacial climates, and the extreme efficiency of the glacial buzzsaw, where it occurs (Anderson et al., 2006; Anders et al., 2008; Brozovic et al., 1997). As a result, orogenic systems may behave differently when dominated by glacial conditions, and thus climate should be considered fundamental to evolution of orogens and other tectonic processes, such as terrane accretion, throughout the geologic record (Molnar and England, 1990). Exactly how glacial erosion will have influenced a given tectonic setting depends on local conditions, given the potential thresholds for onset of the glacial buzzsaw (Foster et al., 2010). For example, some orogens may respond to the onset of glaciation, whereas others may not respond until a certain degree of glaciation is met, such as forced by a change in the periodicity of Milankovitch cycles (Berger and Spotila, 2008a). Where it is present, however, the glacial excavator should be highly effective at denuding an orogen and lead to a high degree of coupling between erosion and tectonics, including fostering the development of a steady-state flux (Tomkin and Roe, 2007).

As is typical in science, the most interesting lingering questions related to glacial erosion and their effect on tectonics involve the exceptions to rules of behavior. A problem of major importance for the evolution of alpine topography and orogenic systems is how and why the glacier buzzsaw can be violated, the role of thresholds, and what controls the height and character of teflon peaks (Foster et al., 2010). Equally revealing may be the exceptions to the general rule that glacial erosion scales with basal ice flux, such as can result from subglacial hydrology or the thermal regime of the ice-substrate interface (Riihimaki et al., 2005; Jamieson et al., 2008). Understanding the temporal variation and timescale dependence of glacial erosion is also a prime target of future research. Another direction of investigation that will advance our understanding will be a focus on what controls sediment production in glacial landscapes. Studies of headwall retreat and teflon peaks have shown the importance of periglacial weathering, which in turn is important for the erosional processes of the glacier itself (Oskin and Burbank, 2005; Foster et al., 2010; MacGregor et al., 2009). Studies have shown that periglacial erosion may contribute the majority of sediment that is transmitted by a glacier (O'Farrell et al., 2009), and frost shattering may actually be a mechanism of limiting relief in glacial landscapes (Hales and Roering, 2009). Detailed studies focusing on these secondary aspects of the glacial system are needed. Likewise, quantitative field studies of the links between processes and conditions in glacial systems are needed to enhance physics-based erosion laws, such as quantification of contributions of abrasion versus plucking, measurements of input of clasts into the bergschrund, rates of subglacial weathering, and the influence of bedrock properties on topographic and orogenic evolution (Lliboutry, 1993; Riihimaki et al., 2005). Finally, additional case studies of the glaciated orogenic systems that exhibit a high degree of coupling between climate, erosion, and tectonics, are needed (e.g., the St. Elias orogen, the Southern Alps, and the Patagonian Andes). These constraints will enable more complex, comprehensive, three-dimensional modeling of ice flow and resulting erosion, in what amounts to a systems-based approach to understanding glacial erosion and its effect on evolving topography and tectonics.

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Biographical Sketch

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