8.7 Glacial Erosion Processes and Rates

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Glossary

Abrasion The small-scale mechanical breakdown of rock surfaces caused by stress and motion at the contact point between a clast embedded in a glacier and the underlying bedrock. Abrasion produces fine rock particles (glacial flour) that are transported away in meltwater and leaves erosional traces that include striations and polish on a rock surface.

Entrainment Refers to the processes that result in rock particles or sediments being incorporated within, attached to, or moved by the glacier ice.

Erosion Refers to the removal of soil or rock from a location and includes both processes that breakdown or reduce the strength of the material and processes responsible for the movement of the material from its original location.

Paraglacial environments These are transient, nonglacial environments that are not, at present, directly impacted by glaciation, but have process types and rates that are responding to recent deglaciation.

Plucking Also sometimes termed as quarrying, it includes both the fracture of underlying rock as well as the entrainment of rock fragments that result from either fracture due to glacial action or those that have been isolated by preexisting bedrock cracks.

Regelation Is the refreezing of water that was originally ice.

Abstract

Glacial erosion includes processes that occur directly in association with glacial ice, such as abrasion, plucking, physical and chemical erosion by subglacial meltwater, as well as processes that are enhanced or modified by glaciation. Studies of the mechanics of subglacial rock fracture indicate that erosion is controlled strongly by basal sliding velocity and bedrock lithology. Measuring or reconstructing erosion rates related to glaciation is challenging, and results from a variety of methods show that glaciers and ice sheets have erosion rates that range from essentially zero under cold-based ice up to 10^1 mm yr^{-1} under temperate ice.

8.7.1 Introduction

In a landscape that undergoes a cycle of active glaciation, a wide range of processes break down rock and then pick up and move this material to other locations; the breakdown, entrainment, and initial transport of material is what is commonly meant when a geomorphologist uses the term 'erosion.' Spatial patterns in erosion drive changes in the shapes of land surfaces, measured across scales from millimeters to thousands of kilometers, and it is these changes in shape that produce the distinctive landforms and landscapes that are the focus of Chapters 8.8 and 8.9. In fact, because areas that have been glaciated have distinctive landforms and landscapes, it has long been argued that the rates of glacial erosion must be very high (and in this chapter, the authors have considered the evidence for rates of glacial erosion). The transport and subsequent deposition of eroded material also produces distinctive landforms and deposits that are discussed in Chapters 8.10 and 8.11.

Some erosion processes in glacial landscapes are a direct effect of glacial ice, and those that occur at the base of a glacier are discussed in more detail in this chapter. However, other processes in a glaciated landscape are impacted directly or indirectly by glaciation, for example, enhanced rockfall from

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the upper part of a slope that has been steepened by the removal of rock from the lower part of the slope by direct glacial action, or enhanced fluvial erosion in a valley as a result of pulses of meltwater produced by melting of glacier ice further up the valley (*see* Chapter 8.12; **Figure 1**). Thus, it is not uncommon for measures of the rates of glacial erosion and descriptions of the impacts of glaciation to include a range of processes, including those strictly associated with the direct impact of glacial ice as well as those that are modified in rate or pattern by the presence of glaciers in the landscape. A glaciated landscape that we study today is typically conditioned by glaciation but represents the product of a complex range of glacial, paraglacial (Ballantyne, 2002), and periglacial processes.

Figure 1 Modern alpine glacier landscape, Glacier No. 1, Tianshan Mountains, China. Many processes in a glaciated landscape are impacted directly or indirectly by glaciation, for example, rockfall on the glacier from slopes that have been steepened by glacial action and modified fluvial processes downstream as a result of pulses of meltwater and sediment from the melting of glacier ice.

8.7.2 Processes of Glacial Erosion

Glacial erosion includes the breakdown and loosening of rock or sediment by processes directly related to glacier ice, as well as the entrainment and initial transport of this material by glacier ice or meltwater (Fu and Harbor, 2011). Traditionally, the primary processes of glacial erosion have been described as: (1) abrasion, (2) plucking (quarrying), and (3) physical and chemical erosion by subglacial meltwater (e.g., Sugden and John, 1976). This basic scheme was based partly on the interpretation of processes involved in the formation of characteristic glacial landforms and sediments; although additional research has advanced understanding of the components of these processes, this basic organizing scheme is still useful. Physical and chemical erosion by subglacial meltwater are discussed in Chapter 8.6, and the movement of subglacial sediments (sometimes termed soft-bed erosion) is discussed in Chapter 8.10, so these components of glacial erosion will not be discussed further in this chapter.

Rock fracture is fundamental to both abrasion and plucking, and all rocks undergoing glacial erosion have preexisting cracks across a range of scales, from microscopic flaws and structural planes in individual minerals to large joints and bedding planes on scales up to tens of meters (**Figure 2**). Unless preexisting cracks have completely isolated a mass of

(a) (b)

Figure 2 Plucked and scoured rock surfaces in Sweden. (a) Ice flow direction was toward the camera, picture shows lee side of a large rock hump. Note bedding planes play a major role in controlling fracturing on the plucked surface. (b) Ice flow was from right to left. Note complex spatial patterns of fracture in this area of intensive scouring with lee-side plucked surfaces.

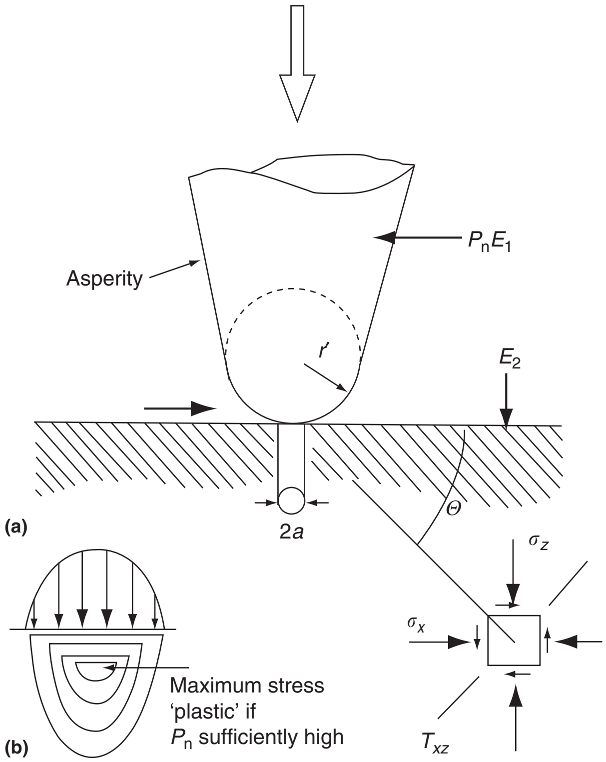

Figure 3 Contact of a spherical asperity with bedrock under a load (P_n). (a) The asperity tip has a radius r'. The asperity and bedrock possess Young's moduli of E_1 and E_2, respectively. The pattern of loading through a cross section of the contact area ($2a$) is shown. Shear (T) and normal (σ) stresses in bedrock resulting from surface loading are also indicated. The load is separated into normal and horizontal components. (b) Shows the contours (isolines) of constant maximum shear stress beneath the point (for a normal load only). Reproduced from Drewry, D.J., 1986. Glacial Geologic Processes. Edward Arnold, London, UK.

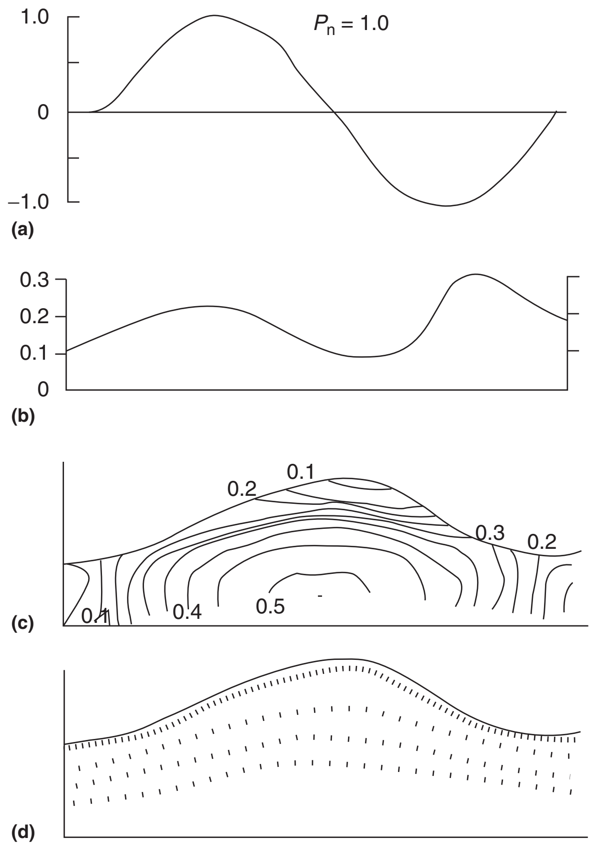

Figure 4 Stress patterns induced in a subglacial bedrock undulation for normal pressure $P_n = 1.0$. Ice flow is from left to right. (a) Distribution of normal stress at the interface; (b) maximum shear stress at the bedrock surface; (c) maximum shear stress contours at depth (MPa); (d) orientation of principle stress axes. Reproduced from Drewry, D.J., 1986. Glacial Geologic Processes. Edward Arnold, London, UK.

rock, stresses induced by glacier ice must extend preexisting cracks by the process of rock fracture as the first step in erosion. A comprehensive review of rock fracture in geological processes is provided by Gudmundsson (2011). At the base of a glacier that is moving relative to its bed, shear stresses related to ice motion relative to the bed, in addition to spatial variations in normal pressure, produce tensile and shear stresses in the underlying rock (**Figures 3** and **4**). Depending on the rock strength and the orientation of preexisting cracks and microfractures relative to the stress field, these stresses may be sufficient to cause rock failure at the tip of the crack. The existence of a crack in the underlying rock actually acts to concentrate the stress at the tip of the crack, magnifying the force that is applied, and this effect increases with crack length. As fracture occurs, stored strain energy is released, and beyond a critical crack length, this additional energy on top of the externally applied stresses is sufficient to continue the crack growth resulting in rapid failure. At the base of a glacier moving over bedrock, stresses are concentrated at the contacts between the bed and clasts (rock fragments or particles), allowing for abrasion, and on bumps and steps where the bed is not smooth, allowing for plucking.

8.7.3 Plucking and Entrainment of Rock Fragments by Ice

Plucking, also sometimes termed as quarrying, includes both the fracture of underlying rock as well as the entrainment of rock fragments that result from either fracture due to glacial action or those that have been isolated by preexisting bedrock cracks. The results of plucking can generally be seen particularly well on the lee sides of bedrock steps or outcrops (**Figure 2**). The fracture pattern is driven in part by preexisting cracks, bedding planes, and weaknesses in the rock, as well as by stress concentrations that occur where there are changes in the slope of the bed (in particular bumps and steps, **Figures 4** and **5**). This fracturing can produce clasts ranging in size from large boulders to very fine rock fragments. Although plucking will exploit preexisting joint patterns and fractures, if erosion is to be maintained, then active fracture of bedrock under ice has to take place after all the preglacial weathered and fractured rock has been removed. As rock is removed differentially by glacial erosion across a landscape, and as ice builds up and decays, new fractures can be produced by changes in the *in situ* rock stress field that result from changing load patterns on the underlying rock (Augustinus, 1995).

Rapid subglacial fracture occurs when the pattern of stress induced both by the ice and by any embedded clasts causes a tensile stress at a crack tip that exceeds the fracture strength of the rock. For ice flow over an uneven bed (**Figures 4** and **5**), the force applied by ice at the ice-bed contact is tangential to the bed and this produces a distinctive pattern of differential stress. In case of a simple undulation (**Figure 4**), the surface stress has its maximum values down-glacier (lee side) of the top of the bump (**Figure 4(b)**), and the principle stress axis is parallel to the surface in this area (**Figure 4(d)**), and thus this region is where fracturing is most likely to occur.

For ice flowing over a bedrock step where there is a cavity on the down-glacier side of the step, the maximum stresses and thus the locations of probable fracture initiation and crack propagation occur both on the bedrock ledge and at the point where the ice regains contact with the bed on the down-glacier side of the cavity (**Figure 5**). Cavities occur where ice flows over bumps or steps at a velocity that does not allow vertical deformation to keep basal ice in contact with underlying rock (Hallet, 1996; Iverson, 1991), and act to reduce the contact area between the bed and overlying ice, which in turn increases the average stress on the areas of contact (**Figure 5**). If the cavity is filled with water, then the force applied at the ice-bed contact is controlled by the effective pressure (P_e), which is the difference between the basal water pressure (P_w) and the ice overburden pressure (P_i):

$$P_e = P_i - P_w \qquad [1]$$

Theoretically, high fracture rates should occur where effective pressures are high, which is when the ice is thick and water pressures at the bed are low. However, temporal variations in basal water pressure in cavities can produce very large short-term increases in local stress, and these temporal variations can be largest when basal water pressures fluctuate between very high and very low values (Iverson, 1991). Experiments conducted at the Svartisen Subglacial Laboratory by Cohen et al. (2006) showed that subglacial rock crack propagation occurred during periods of rapidly falling cavity water pressures. In Hallet's (1996) analysis of glacial quarrying, the maximum plucking rates occur with fast sliding velocities (leading to cavity formation and high shear stresses in areas of ice-bedrock contact) and low effective pressures (0.1–1 MPa) resulting from high basal water pressures. Note that the pattern of fracture shown in **Figure 5** will result in headward erosion of the steps, maintaining the stepped topography rather than smoothing out the feature.

Finally, it is interesting to consider whether there could be plucking on a glacier bed that does not have steps or bumps; presumably at rates far lower than in areas with distinct topography (Hallet, 1996). For differential crack growth, this would require either differential patterns of preexisting fractures or rock strength (i.e., a variation in the underlying rock) or a spatial variation in ice flow, which could be produced by complex patterns of basal thermal regimes (*see* Chapters 8.4 and 8.5 and Harbor et al., 2006). If this produced a pattern of differential fracturing, then for plucking to occur there would also need to be a mechanism to entrain blocks of rock that are separated from the surrounding rock by fractures.

For plucking to occur, it requires both that the material is loosened or weakened by fracture and the removal of this material from its original location by entrainment. The primary entrainment mechanism at the base of glaciers involves incorporation of fragments into basal ice as a result of freezing. On small spatial scales, this can result from a heat pump effect (Robin, 1976) associated with bumps or steps at the base of a glacier. When ice at the pressure melting point flows over a bedrock bump or step, higher pressures on the up-glacier side (**Figures 4** and **5**) produce melting. The water that is produced by melting flows to regions of lower pressure on the lee side of the bump or step where a cavity may exist. If the lower pressure regions are below the pressure melting point, then the meltwater will refreeze creating a regelation layer. This process can be maintained because meltwater freezing on the lee side of a bump releases heat, which can be conducted through small bedrock bumps to the up-glacier side, creating a positive feedback loop, which has been termed the heat pump effect. For plucking, what is important is that the refreezing of water to the bed of the glacier can incorporate into the glacier ice clasts at the bed that were loosened by fracturing. These clasts will then be removed from the bed (eroded and entrained) as the regelation layer ice moves further down the glacier.

The heat pump effect operates over a small scale (up to meters), but there are also large-scale spatial and temporal changes in basal thermal conditions under glaciers and ice sheets (*see* Chapter 8.5). Where there is a change from net melting to net freezing of subglacial water, clasts may become frozen into the basal ice and can be removed as the ice flows

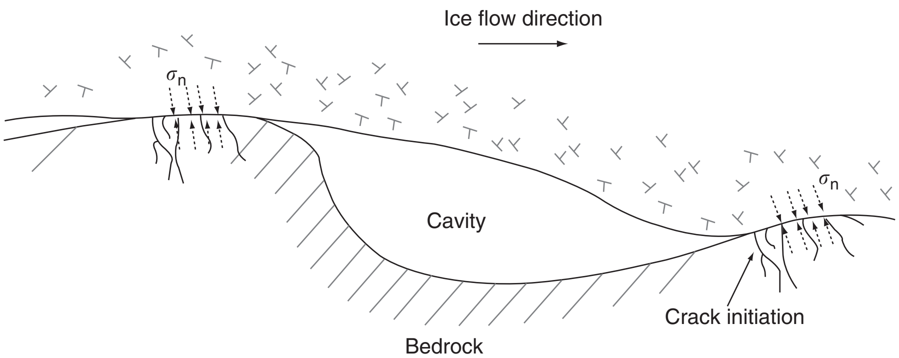

Figure 5 Locations of maximum normal stress and probable fracture initiation associated with ice flowing over a bedrock step with a cavity. Modified from Hallet, B., 1996. Glacial quarrying: a simple theoretical model. Annals of Glaciology, 22: 1–8, with permission from IGS.

down-glacier. This includes boundaries between warm- and cold-based ice patches as well as temporal changes in basal thermal regimes at a point during a period of glacier advance and retreat. Such patterns and changes allowing for entrainment of loosened and fractured rock underlie models that explain large-scale differences in glacial erosion that drive glacial landscape types, as reviewed in Chapter 8.9.

8.7.4 Abrasion

Abrasion occurs when clasts embedded in the basal layers of a glacier scratch, score, polish, and wear down the bedrock over which they pass. Crushing and fracture at the point where the clast comes in contact with the underlying bedrock produces a pattern of rock modification that results in small-scale features such as striations (longitudinal scratches that are millimeters across and have lengths from centimeters to meters, Chapter 8.8) and chattermarks (**Figure 6**), and produces fine material (glacial flour) that is generally washed away in subglacial meltwater. Under a microscope, striations can be seen to include numerous crescent-shaped fractures (Benn and Evans, 1998; Liu et al., 2009) that indicate that they are produced by numerous discrete crushing and fracturing events at the microscale.

An apparent paradox in glacial erosion is how ice, which is mechanically weaker than rock, can produce abrasion. Why is a clast embedded in ice not pushed back into the ice as a result of ice fracture or deformation, rather than held against the underlying rock surface with sufficient force to produce crushing and fracture at the point where the clast contacts underlying bedrock? One of the primary reasons for this is that the ice is in contact with the clast over a large area, whereas the clast is in contact with the underlying bedrock at a small point (an asperity, **Figure 6(a)**). Thus, a small force per unit area over the large contact surface between the ice and the clast is then applied to a very small contact area between the clast and the bedrock, resulting in a very large force per unit area. This is sometimes described using the analogy of a thumbtack (drawing pin) where a small force per unit area is applied to the broad top of the thumbtack by a thumb, resulting in a sufficiently large force per unit area at the point of the pin to drive the sharp end of the thumbtack into a piece of wood.

In a simple abrasion model (**Figure 6(a)**), a clast embedded in basal ice is forced toward the underlying bedrock while also being moved along the bed by down-glacier ice motion.

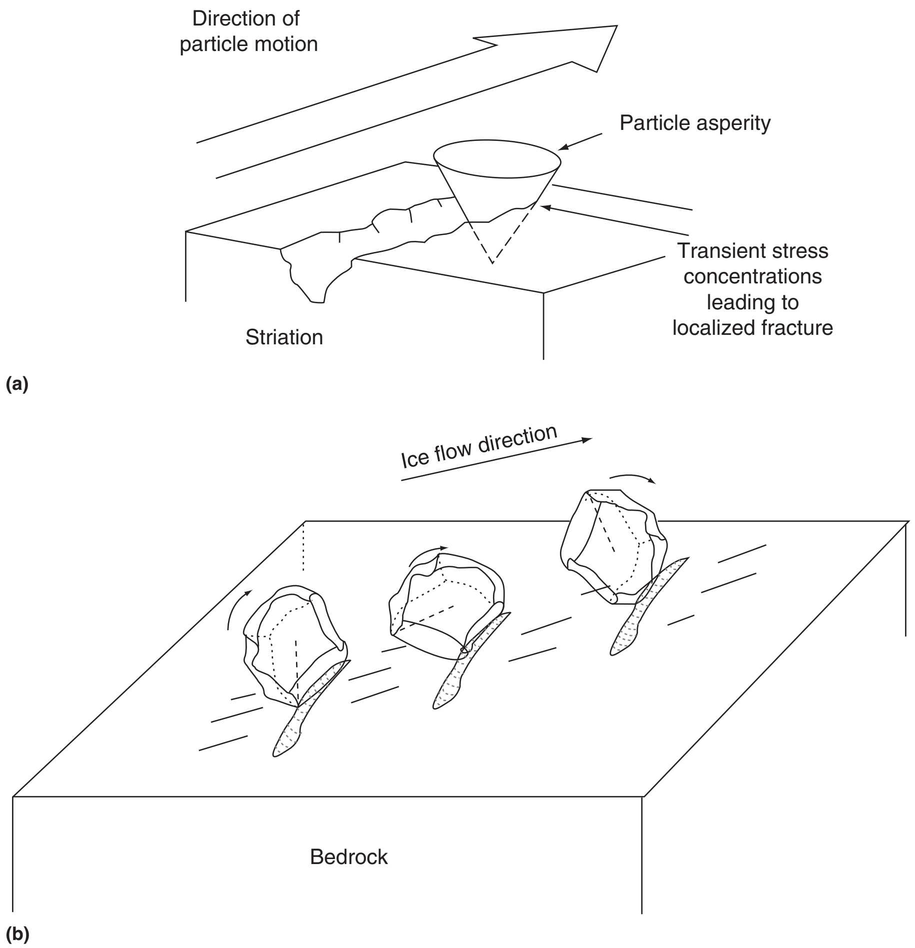

Figure 6 (a) Simple model of the striation process. (b) Set of chattermarks produced by a rolling pattern of movement of a clast across bedrock. Reproduced from Benn, D.I., Evans, D.J.A., 1998. Glaciers and Glaciation. Edward Arnold, London, 734 pp (modified from Drewry 1986).

This produces an indentation in the bedrock if the bedrock lithology is weaker than that of the clast. As the clast is moved across the bedrock, it produces an elongated indentation called a striation. The striation depth is controlled by the relative hardness of the bedrock and the clast, and by the magnitude of the force directed on the clast through the contact point toward the bed. The length of the striation produced per unit time is then controlled by the bed-parallel velocity of the clast. In such a model, the abrasion rate *Ab* is a function of:

- the force pressing the clast against the bed, F
- the velocity of the clast relative to the bed, U_p
- the concentration of clasts in basal ice, C_o
- the relative hardness of the clast and the bedrock, ΔHd

As the clast is moved across the bed by glacier motion, wear products, changes in the shape of the asperity, and increased friction can all inhibit the movement of the clast causing lodgment or rotation. When a clast rotates (**Figure 5(b)**), successive contact points between the clast and the bed produce distinct small fracture points, crescents or striations, called chattermarks.

Although there is general agreement that abrasion occurs as a result of stress concentration, crushing, and fracture when a clast in ice comes in contact with the underlying bedrock at a small point of contact (**Figure 6(b)**), varying approaches have been developed for deriving F, the force pressing the clast against the bed. Boulton's (1974, 1976) model of glacial abrasion is based on the idea that clasts at the bed of a glacier support the weight of ice that overlies them and is sometimes described using the analogy of a piece of sandpaper (Drewry, 1986). For temperate glaciers, basal water pressure also supports some of the overlying weight of ice; thus in Boulton's model, F depends on effective pressure:

$$F = (\rho_i g h - P_w) A_i \qquad [2]$$

where ρ_i is ice density, P_w is water pressure, and A_i is a measure of the contact area between clasts and the bed. If eqn [2] is correct, then abrasion rates should be highest under thick ice with low basal water pressure. However, recognizing that with very high effective pressures, the velocity of the clast will be reduced by friction; in Boulton's model, the abrasion rate decreases beyond a maximum F value that is a function of ice velocity and for very large values of F, clast motion falls to zero (deposition). The model's implicit assumption of a fairly rigid contact between the glacier base and underlying bedrock has been observed in some field settings (Anderson et al., 1982; Souchez and Lorrain, 1987) and laboratory experiments (Iverson, 1993). However, in other cases, basal ice has been observed to deform around basal clasts, which led Hallet (1979, 1981) to the development of an alternate model for basal conditions and abrasion mechanics.

In Hallet's abrasion model, it is recognized that ice flows in a direction that includes components that are both down-glacier and toward the bed (where melting is occurring). To accommodate the component of flow toward the bed, ice must deform around the clasts that are in contact with the bed (Hallet, 1979, 1981) and as a result, viscous drag produces a bed-directed force on clast. In Hallet's model, clasts do not support the weight of the overlying ice. Instead, like any other object in a fluid, they have a buoyant weight that is a result of the difference between the density of the ice and the density of the clast. Thus, the tangential force on the clast contact point (F) in the direction of the bed is the combination of the buoyant weight (F_b) of the clast and a viscous drag (F_i) (eqns [3]–[5]):

$$F = F_b + F_i \qquad [3]$$

$$F_b = \frac{4}{3}\pi R^3(\rho_r - \rho_i) g \cos\theta \qquad [4]$$

$$F_i = \frac{f 4\pi\eta R^3}{R_*^2 + R^2} v_n \qquad [5]$$

where R is the asperity radius; ρ_r and ρ_i are the density of rock and ice, respectively; θ is the local down-glacier inclination of the bed; f is a parameter describing the bed influence on modifying the viscous drag for a clast in ice; η is the effective viscosity of the ice; R_* is the transition radius (which affects regelation and creep); and v_n is the ice velocity normal to the bed (Hallet, 1979). The buoyant force is typically insignificant except for very large clasts, and thus in most cases eqn [5] is the primary driver of the abrasion model. In applying this approach to a situation of multiple clasts in basal ice, the primary controlling factors become:

- Basal melt rate
- Clast concentration
- Clast properties: size and density

High abrasion rates occur when basal melting rates are high and clast concentrations are low enough that they do not impede ice flow toward the bed.

In comparing these different formulations, it is interesting to note that in eqn [2], F will be highest under thick ice with low water pressures; whereas in eqn [5], F is highest when the sliding velocity is high. However, although the sliding velocity of a glacier also increases with ice thickness, it decreases with low water pressure; thus there are situations where the different models of the force involved in abrasion (F) produce very different results in terms of predictions of spatial patterns of glacial abrasion. Velocity-based erosion models that are consistent with Hallet's model for glacial abrasion as well as the strong velocity-dependence of plucking models have been effective in predicting the development of glacial landforms (e.g., Harbor et al., 1988; Hooke, 1991; Harbor, 1992) and are now widely used in landscape development models (e.g., MacGregor et al., 2000; Tomkin, 2009; Egholm et al., 2009; Shuster et al., 2011).

8.7.5 Rates of Glacial Erosion

In addition to research on the mechanisms and spatial patterns of glacial erosion, there has been considerable debate since the latter half of the nineteenth century about the erosive power of glaciers and their ability to substantially modify the landscape (reviewed briefly in Harbor, 1989). In contrast to Tyndall's (1862) argument that alpine valleys were solely the product of glacial erosion, the significance of glacial erosion in the development of major landforms was doubted by several influential geologists in the early twentieth century (e.g., Bonney, 1910; Garwood, 1910). For example, Garwood (1910, p. 311) wrote that "ice, on the whole, erodes less rapidly than other denuding agents, and … under certain conditions, it may act relatively as a protective agent." However, based on evidence of large-scale deposition related to glaciations and extensive areas with features identified as resulting from glacial erosion, by the late 1880s, it was generally accepted that glaciers were capable of significant erosion (Davies 1969). Today, most textbooks of physical geography and geology describe glaciers as agents of considerable erosion, and in a recent paper Shuster et al. (2011), echoing Tyndall's (1862) assertion, used thermochronometry, and numerical modeling to show that valleys in the archetypal glacial landscape of the Fiordland region of New Zealand were carved almost entirely during the last 2 million years, a period dominated by glaciations.

Several challenges involved in determining rates of glacial erosion exist, given that we are typically trying to determine when and how quickly rock was removed from a location in the past, on the basis of what is left behind or where and how quickly the material arrived at another location. These challenges include defining what is actually meant by glacial erosion (given that glaciers also transport material produced by processes other than basal erosion), widely varying measurement techniques, and the natural variability of glacial erosion temporally and spatially. Most glaciated landscapes have been produced by phases of both glacial and nonglacial erosion over long time periods, and most areas of glaciation studied today include a mix of glacial and nonglacial processes that are difficult to unravel (Harbor and Warburton, 1993; Delmas et al., 2009). At the most general scale, it is reasonable to argue that what we are seeking to distinguish between are rates of erosion in areas affected by glaciation versus those that are unaffected, and thus it is reasonable to include approaches that cannot distinguish between processes under ice and other processes impacted by glaciation. However, as Delmas et al. (2009, p.487) lamented "A major problem is ..(to).. distinguish between debris that were directly quarried by the glacier from the underlying bedrock, debris that were supplied to the glacier by supraglacial hillsopes, and the preglacial – and perhaps predominantly nonglacial – regolith that the glacier cleared away during ice advance." Thus, there is

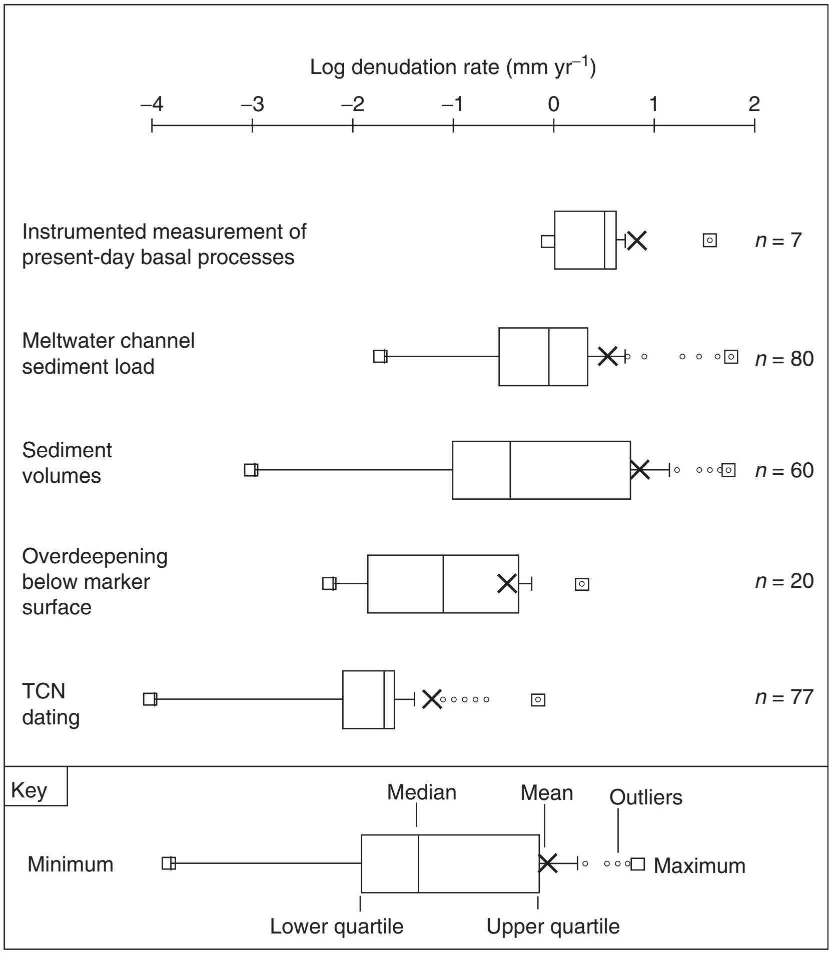

Figure 7 Box and whiskers plots highlighting variability in glacial erosion rates based on five different estimation methods. Reproduced from Delmas, M., Calvet, M., Gunnell, Y., 2009. Variability of Quaternary glacial erosion rates – a global perspective with special reference to the Eastern Pyrenees. Quaternary Science Reviews, 28, 484–498.

considerable interest in unraveling the details of the sediment budgets of glaciated areas, including the assessments of actual rates of rock loss under active ice. However, in making present-day measurements, we are limited to very short timescales and settings that represent relatively little glacial extent compared with periods of past glacial maximum conditions, although several studies have attempted to compare watersheds with a range of levels of glaciations as a way of trying to determine the glacial component of erosion (e.g., Hicks et al., 1990, but also see Harbor and Warburton, 1993).

In summarizing a wide range of studies of glacial erosion rates, and building on the review provided by Hallet et al. (1996), Delmas et al. (2009) identified five main approaches to reconstructing glacial erosion rates (**Figure 7**). These approaches range from very short timescale studies of active processes to many long-term attempts to reconstruct erosion rates based on the volume of rock lost in creating erosional features or the volume of sediment deposited associated with glaciation:

- Instrumented measurement of present-day basal processes, typically abrasion under temperate glaciers accessed by cavity systems or tunnels. This approach generally yields relatively high erosion rates for short time periods at very specific locations that are chosen because it is believed that they are sites of active (thus measurable) erosion. Boulton (1974), for example, recorded erosion rates of 3–30 mm yr^{-1}.
- Watershed-averaged erosion from meltwater sediment fluxes, typically suspended sediment load for short-term studies of current glaciers. This approach combines all glacial and nonglacial processes in the watershed above the measurement point and yields values that typically range between 10^{-2} and 10^{1} mm yr^{-1}. Given the storage of sediment in the glacial system, there may be very significant temporal variations in sediment fluxes over a range of timescales, and Koppes and Montgomery (2009) suggested that erosion rates calculated from sediment yields are higher during glacial retreat because warmer basal temperatures are enhancing flow at the bed and increased basal melt may be remobilizing sediments stored at the glacier bed.
- Terrestrial cosmogenic nuclide reconstructions of erosion depths based on incomplete resetting of the cosmogenic record due to erosion depths <3 m. This approach provides high precision measurements for specific, well-constrained study sites where erosion depths have been low, but only a minimum estimate for larger erosion depths. This approach yields the lowest glacial erosion rates, typically between 10^{-4} and 10^{-1} mm yr^{-1}, and only provides minimum values for areas with high erosion rates.
- Estimates of glacial incision relative to a preglacial or nonglaciated surface. This approach yields rates that are much lower than direct instrumental measurements, and that average glacial and nonglacial processes over very long timescales.
- Watershed-averaged erosion derived from total sediment volumes in deposits, such as offshore sediments, moraines, till sheets, and lake sediments. This approach also combines all processes operating in a watershed during the period of study (ranging from individual years for varve studies to the entire Quaternary) and typically yields rates from 10^{-3} to 10 mm yr^{-1}.

In combination, these methods yield results spread over five orders of magnitude (10^{-4}–10^{1} mm yr^{-1}), which is not surprising given the variability of types and scales of glaciation over space and time and the fact that the methods range from short-term measurements of active processes in locations where we would expect glacial erosion to be high, to methods that are best suited to areas with little erosion. Overall, the wide range of values suggests not only that the glacial erosion rates can be extremely high in certain situations but also that the rates of glacial erosion can be very low in cases where a glacier or ice sheet is frozen to its bed. This complexity makes it difficult to generalize about relative rates of glacial and nonglacial erosion, and explains why some areas affected by glaciations quickly develop characteristic glacial landforms and landscapes, and other areas retain nonglacial features despite being in locations that have undergone repeated glaciations (Stroeven et al., 2002).

8.7.6 Conclusion

Erosion associated with glaciation includes a wide variety of processes, ranging from the most basic interactions between ice, water, and bedrock or sediment at the base of a glacier, to other processes in a watershed that are affected by glaciation, including slope and fluvial processes in areas above and below the ice itself. Glacial abrasion and plucking are generally viewed as the processes most clearly associated with glacial action alone and are related specifically to several landforms and landscape elements discussed in the following chapters. Considerable field, laboratory, numerical modeling, and theoretical work have established a detailed understanding of the variety of conditions that occur at the ice–rock interface, and that result in failure and fracturing of underlying rock. Several different conceptions of the mechanics of glacial abrasion and quarrying have been proposed, and it is likely that these in fact represent some of the range of mechanical conditions that exist at the base of glaciers. In all of these models, a key controlling variable is sliding velocity, and thus most models that seek to predict temporal or spatial patterns of glacial erosion include sliding velocity raised to some power in addition to measures of rock resistance to erosion.

Attempts to establish rates of glacial erosion have involved considerable creativity in developing novel methods to back calculate how much rock was removed and when. This has included comparisons to adjacent nonglaciated surfaces, measures of sediment generated by erosion, cosmogenic nuclide and thermochronometry approaches, and direct measurements of short-term erosion rates at glacier beds. This work has demonstrated that rates of glacial erosion vary widely, from essentially zero under frozen-bed conditions, to rates as high as tens of millimeters per year. However, most measures of glacial erosion do not specifically separate out the direct action of ice on underlying bedrock from other processes in glaciated areas, which can include transport of rockfall and avalanche debris on the glacier surface, remobilization of stored sediments, and enhanced fluvial activity related to pulses of meltwater.

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Biographical Sketch

Jonathan Harbor's early fascination with glacial landscapes acquired while hiking through the English Lake District has developed into a very enjoyable career in research and teaching. His early enthusiasm was encouraged and developed by great mentors at Cambridge University (BA), the University of Colorado (MA), and the University of Washington (PhD), and by exceptional collaborators internationally as well as at his home institution, Purdue University. Projects he has been involved in with his students have ranged from detailed field investigations of basal sliding and glacier internal structure; numerical modeling of ice flow and landform development; and using cosmogenic nuclides in conventional and novel ways to reconstruct glacial chronologies, extents, and erosion patterns. This has provided opportunities for a study in a wide range of locations, including many parts of North America, the Alps, northern Fennoscandia, the Tibetan Plateau, and central Asia.