8.8 Erosional Features

MJ Munro-Stasiuk, Kent State University, Kent, OH, USA
J Heyman and J Harbor, Purdue University, West Lafayette, IN, USA

Glossary

Chattermark Crescentic-shaped fractures caused by glacial erosion of bedrock surfaces; these features are concave in the down-ice direction and are oriented perpendicular to ice movement.
Cirque Amphitheater-shaped depression cut in to mountainsides by small ice patches (cirque glaciers), with an overdeepened section commonly filled with a small lake (tarn) after deglaciation.
Crag and tail An elongated streamlined hill that is smoothed by abrasion on its up-glacier end and with a tail of preserved existing sediment or bedrock, or deposited sediment on its down-flow end.
Knock and lochain Extensive areas of subglacially eroded bedrock that includes rock knobs, roches moutonnées and rock basins; from Scots Gaelic words for 'knoll' and 'small lake.'
Roche moutonnée An asymmetric bedrock knob or hill with a smoothed surface on the up-glacier side and a plucked, quarried surface in the down-flow direction.
s-Form Smoothed and sculpted bedrock surface resulting from glacial erosion.
Striation A scratch or small elongated groove on a rock surface resulting from glacial abrasion.
Tunnel valley Long (10^1 km), wide (10 km), flat-bottomed, overdeepened (10^1–10^2 m), radial or anabranching valley system incised into bedrock or sediment that terminates in an ice marginal fan.

Abstract

The wide range of features produced by glacial erosion, over scales from millimeters to kilometers, results from both the complexity of the processes and their glaciological controls, and interactions with heterogeneous substrates. Small-scale features include striations or striae and chattermarks that are produced by clasts in basal ice being forced against underlying bedrock, as well as smoothed and sculpted rock surfaces (s-forms). Intermediate scale such as whalebacks and rock drumlins are smoothed, subglacially formed, elongate bedrock landforms produced by glacial abrasion, and somewhat similar are roche moutonnées which have a smoothed up-glacier surface and a plucked, quarried surface in the down-flow

Munro-Stasiuk, M.J., Heyman, J., Harbor, J., 2013. Erosional features. In: Shroder, J. (Editor in Chief), Giardino, R., Harbor, J. (Eds.), Treatise on Geomorphology. Academic Press, San Diego, CA, vol. 8, Glacial and Periglacial Geomorphology, pp. 83–99.

direction. Other classic landforms that result from a combination of erosion and deposition include crag and tails, drumlins and flutings. Larger scale characteristic features of glaciated landscapes reflect patterns of erosion controlled by ice dynamics and glacial history, and include glacial troughs and fjords, over-deepened rock basins, cirques, and large areas of extensively scoured bedrock.

8.8.1 Introduction

Erosion by glacial ice has had a profound effect on high latitude and high elevation landscapes, wearing down existing topographic highs, carving out deep valleys, and shaping complex landforms as a result of the interplay between glacial processes and geological materials. The properties that affect glacial erosion are complex, and relate to the nature of the ice (e.g., warm-based vs. cold-based; drainage systems in the ice), the nature of the substrate (e.g., hard beds vs. soft beds; permeability; general rheological properties), the nature of the topography (slope, bed roughness, pre-existing valleys), and the time available for all these processes to occur. The nature of erosion can also differ depending on whether the ice is of valley origin or continental origin, or whether erosion is under, to the side, or in front of glacial ice. As reviewed in detail in sections below, erosion occurs in one of three forms:

1. Abrasion – the general wearing down of rock surfaces by ice and its entrained clasts;
2. Fracturing and quarrying – rock fracture, loosening of rock fragments, and subsequent evacuation of those fragments; and
3. Meltwater erosion in, under and in front of the ice.

Understanding that the environment of erosion is complex, this chapter provides an overview of the major erosional forms organized by landform scale.

8.8.2 Small-Scale Erosional Forms

Small-scale forms typically range from a few millimeters to a few tens of meters in size (e.g., Kor et al., 1991; Munro-Stasiuk et al., 2005). Common forms related to ice erosion include striations, grooves, and various fractures, whereas multiple s-forms (also known as p-forms) have been attributed to erosion by ice and/or meltwater.

8.8.2.1 Striations and Chattermarks

Striations or striae are scratches or small elongated grooves in bedrock or on clasts that are the product of abrasion (**Figure 1**). Clasts protruding out of basal sliding ice are dragged along bedrock surfaces producing the marks. The rate of abrasion depends on the effective force with which individual clast fragments are pressed against the bed, the flux of fragments over the bed, and the relative hardness of rocks in the ice and of the bed (Hallet, 1979). Hallet (1979) also noted that where geothermal heat flow or frictional heating are high, or where the ice is extending, the rate of abrasion should be higher. He also noted that glacier thickness has no affect on abrasion, and hence on the nature of striation morphology.

The presence of striations is a reflection of the spatial and temporal variations in the stresses exerted by rock fragments entrained in basal ice, as well as a representation of glacier sliding (e.g., Boulton, 1974; Kamb et al., 1976; Hallet, 1979, 1981; Shoemaker, 1988; Iverson, 1991). The greater the stresses exerted, the greater the promotion of crack growth and brittle failure, and the deeper and wider the striae tend to be (Drewry, 1986). Typically striae widen in the down-glacier direction, although Iverson (1991) noted that the converse can be true. Most commonly, striae widen gradually (wedge striations) or abruptly at their terminus (nail-head striations), and thus are good indicators of flow direction. Abrasion commonly appears in the form of smooth or polished bedrock surfaces where there may be occasionally thin glacial striae on bedrock surfaces (Ericson, 2004). Polishing occurs once the asperities on the abrading rocks have also been worn down (Benn and Evans, 2010).

Chattermarks and fractures are the direct result of quarrying due to stresses exerted by the overlying ice resulting in rock fracture, loosening of rock fragments, and subsequent evacuation of those fragments (Iverson, 1991; Hallet, 1996). Chattermarks are crescentic-shaped fractures that are concave in the down-ice direction and are oriented perpendicular to ice movement. They typically occur in groups, each of a similar size, somewhat evenly spaced, and parallel to each other. Each fracture is believed to be the result of a single collision between a rock protruding down from the overlying ice onto a bedrock surface.

8.8.2.2 s-Forms (Also Known as p-Forms)

s-Form is the general name given to a suite of small-scale landforms that can be described as smooth and sculpted. Dahl (1965, p. 83) first introduced the term 'p-form' to reflect the 'plastically sculptured detail forms on rock surfaces.' Although he strongly advocated meltwater erosion for the forms, either through catastrophic or prolonged flows, his ideas and his term p-form have often been misrepresented as supporting erosion by abrasion below plastically deforming ice or till. Kor et al. (1991) introduced the descriptor 's-form,' reflecting sculpted form, to indicate that the forms are erosional and to de-emphasize the medium of erosion.

Despite general agreement on a subglacial origin for s-forms, there are strongly opposing views on the processes responsible for creating them. Their formation has been attributed to glacial abrasion (e.g., Boulton, 1974, 1979; Goldthwait, 1979), erosion by subglacial slurries (Gjessing, 1965), and subglacial meltwater erosion (e.g., Dahl, 1965; Sharpe and Shaw, 1989; Kor et al., 1991; Shaw, 1994; Pair, 1997; Munro-Stasiuk et al., 2005). Commonly, s-forms are covered by striae that conform to the shape of the forms in some cases (e.g., Benn and Evans, 1998) and cross-cut others with disregard for form (e.g., Shaw, 1988). The presence of striae has been used as evidence by several researchers to support ice abrasion as the main mode of s-form

Figure 1 Glacial striations: (a) limestone surface at Glacial Grooves State Memorial, Kelleys Island, OH; (b) striated boulder on Skeiderarsandur, Iceland.

formation (e.g., Boulton, 1974, 1979; Goldthwait, 1979). In addition, at modern sites, observations of ice squeezed into grooves and cavettos are cited as evidence that ice can be responsible for eroding bedrock into the shape of s-forms (e.g., Boulton, 1974; Rea and Whalley, 1994).

In support of a meltwater hypothesis, several small-scale forms, such as muschelbrüche, sichelwannen, spindle flutes, and other scour marks have been reproduced in flumes (e.g., Allen, 1971; Shaw and Sharpe, 1987) and identical forms have been documented in purely fluvial environments (e.g., Maxson, 1940; Karcz, 1968; Baker and Pickup, 1987; Tinkler, 1993; Baker and Kale, 1998; Hancock et al., 1998; Wohl and Ikeda, 1998; Gupta et al., 1999; Richardson and Carling, 2005). The presence of sharp upper rims on many of these forms is highly representative of flow separation (e.g., Allen, 1982; Sharpe and Shaw, 1989), and the ever-present crescentic and hairpin furrows represent the generation of horseshoe vortices around obstacles encountered by the flow (Peabody, 1947; Dzulynski and Sanders, 1962; Karcz, 1968; Baker, 1973; Allen, 1982; Sharpe and Shaw, 1989; Shaw, 1994; Lorenc et al., 1994).

s-Forms have been classified by several researchers (e.g., Ljüngner, 1930; Kor et al., 1991; Richardson and Carling, 2005). Richardson and Carling (2005) divide strictly fluvial forms into three topological types: (1) Concave features that include potholes and furrows; (2) Convex and undulating forms such as hummocky and undulating forms; and (3) Composite forms such as compound obstacle marks. Kor et al. (1991) also grouped the forms into three types based on their orientation relative to the flow direction: (1) Transverse s-forms are commonly wider than they are long; (2) Longitudinal forms are generally longer than they are wide, with their long axes laying parallel to the flow direction; and (3) Non-directional forms have no obvious relationship to flow direction. The forms are presented using Kor et al.'s classification in Figures 2 and 3.

Nondirectional forms record no discernable flow direction and include undulating surfaces and potholes. Undulating surfaces are smooth, nondirectional, low-amplitude undulations occurring on gentle lee slopes of rock rises. These low amplitude features are very common to broad areas of

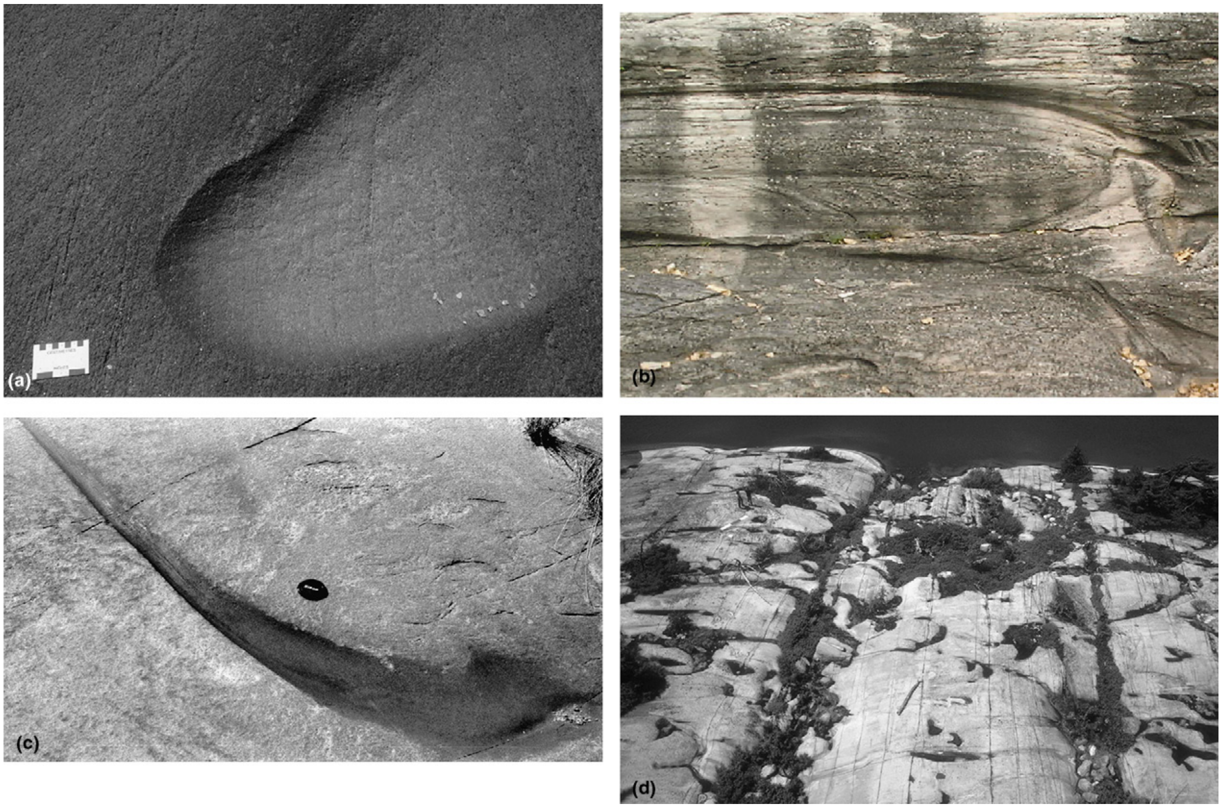


Figure 2 Transverse s-forms: (a) Muschelbrüche (singular muschelbrüch): shallow depressions which resemble the inverse casts of the shells of mussels, with sharp, convex upflow rims and indistinct, downflow margins merging imperceptibly with the adjacent rock surface. The proximal slope is steeper than the distal slope; (b) Sichelwannen (singular sichelwanne): sickle-shaped marks. They have sharp rims convex-up flow, and a crescentic main furrow, extending downflow into arms wrapped around a median ridge. Lateral furrows may flank the main furrow. In an en-echelon system, the ''arms'' merge and bifurcate downflow into other sichelwannen, or may extend downflow into comma forms; (c) Comma-forms (sometimes transitional from sichelwannen) have one arm instead of two; (d) Transverse troughs are relatively straight troughs arranged perpendicular to flow, with widths much greater than lengths. They commonly have a steep, relatively planar uplfow slope or lee face below a relatively straight rim. Potholes often occupy this upflow slope. The downflow or riser slope is gentler and normally eroded by shallow, stoss-side furrows, which produce a sinuous slope contour.

low-relief terrain, generally located peripheral to the main sheetflow events, and in association with the other s-forms described above. Potholes are near-circular, deep depressions that may show spiraling, descending flow elements inscribed on their walls (Gilbert, 1906; Alexander, 1932). Potholes typically occur in bedrock terrains on the summits, flanks and notably, on the lee sides of bedrock knolls and ridges. They range in size (diameter and depth) from a few centimeters to many meters, and commonly have a circular to slightly oval surface expression. The most spectacular potholes occur in the lee of bedrock obstacles or obstructions, and are commonly missing their distal walls. Potholes may occur singly or in small groups, or may occur in large concentrations, commonly showing an en echelon pattern. Some potholes contain a lag of boulders or other debris. Where potholes occur on the lee side of obstructions and they are associated with abundant sculpted forms on bedrock surfaces, they may be interpreted as resulting from catastrophic flood release of meltwater from beneath an active ice sheet (e.g., Embleton and King, 1975; Sugden and John, 1977; Kor et al., 1991). In this association, potholes are considered to be a type of s-form (Kor et al., 1991).

8.8.3 Intermediate-Scale Forms

8.8.3.1 Roche Moutonnées, Whalebacks and Rock Drumlins

Roche moutonnées, whalebacks and rock drumlins are smoothed, subglacially formed, elongate bedrock landforms (**Figure 4**). Whalebacks and rock drumlins are smoothed across their entire surfaces, whereas roche moutonnées have a smoothed up-glacier surface which contrasts with a plucked, quarried surface in the down-flow direction. These features occur in many regions that were formerly glaciated, including Scandinavia, the British Islands, North America, Greenland, and South America (Linton, 1963; Rudberg, 1973; Glasser and Warren, 1990; Evans, 1996; Glasser and Harrison, 2005; Kerr and Eyles, 2007; Roberts and Long, 2005) and range in size from meter scale (e.g., Glasser and Warren, 1990; Knight, 2009) to tens of kilometers (e.g., Kerr and Eyles, 2007). A rock drumlin is a smooth, elongate bedrock form with an asymmetrical long profile and shapes similar to those of drumlins composed of till (cf. Spagnolo et al., 2011). The more commonly used term, whaleback, refers to smooth bedrock forms

Figure 3 Longitudinal s-forms: (a) Rat-tails are positive residual bedforms defined by hairpin scours (crescentic scours with arms extending downflow) wrapped around their stoss ends which then extend downflow. As the lateral scours widen and shallow downflow, the rat tail tapers and becomes lower. They range from millimeters up to several kilometers long (e.g. Allen, 1982); (b) Spindle flutes are narrow, shallow, spindle-shaped marks much longer than they are wide and with sharp rims bounding the upflow side and, in some cases the downflow margins. They are pointed in the upflow direction and broaden downflow. Whereas open spindle flutes merge indistinctly downflow with the adjacent rock surface, closed spindles have sharp rims closing at both the upflow and downflow ends. Spindle flutes may be asymmetrical, with one rim more curved than the other; (c) Cavettos are curvilinear, undercut channels eroded into steep, commonly vertical or near-vertical rock faces. The upper lip is usually sharper than the lower one; (d) Furrows are linear troughs, much longer than wide that carry a variety of s-forms and remnant ridges on their beds and walls. Rims are remarkably straight when viewed over the full length of furrows but are usually sinuous in detail, due to sculpting into the trough walls by smaller s-forms.

Figure 4 Whalebacks on Brandsfjället, northwestern Swedish mountains. Ice flow was from the upper left to the lower right of the picture.

with more symmetrical long profiles. Following Stokes et al. (2011), we use the term whaleback for all intermediate-scale, subglacially formed, smooth bedrock forms.

Whalebacks commonly have striated surfaces, and the formation of whalebacks is generally thought to involve pure glacial abrasion without quarrying, because of the absence of plucked lee sides that are characteristic of roche moutonnées. The smooth surface of whalebacks has been explained as a result of continuous ice-bed contact and an absence of lee-side separation (Glasser and Warren, 1990; Evans, 1996). Glasser and Warren (1990) argued that basal ice pressure is a critical control on the development of whalebacks, and that these medium-scale landforms are formed by small-scale, ice-bed processes controlled by meso-scale basal ice conditions. Evans (1996) investigated whalebacks in the Coast Mountains of British Columbia and argued that these landforms form under thick (and rapidly sliding) ice, keeping the ice pressure high to prevent the formation of plucked roche moutonnées. Evans (1996) proposed that whalebacks can develop under ice a few hundred meters thick, but that an ice thickness of 1–2 km is more favorable for whaleback formation. Glasser and Harrison (2005) investigated two whalebacks and surrounding sediments outside the North Patagonian Icefield, and suggested that the bedrock was eroded by clasts entrained in basal ice rather than by till sliding over bedrock.

The importance of bedrock lithology and structure for the formation of whalebacks has been stressed in several studies. Although Sugden and John (1976) noted that whalebacks most commonly seem to form in crystalline bedrock, Evans (1996) reported whalebacks formed in multiple bedrock lithologies, and Krabbendam and Glasser (2011) argued for abrasion occurring more commonly in softer bedrock. Bedrock structure generally seems to influence the long axis orientation of whalebacks (Evans, 1996; Roberts and Long, 2005; Kerr and Eyles, 2007), and Kerr and Eyles (2007) described whalebacks with long axes parallel to bedrock strikes and independent of varying ice flow directions. Evans (1996) noted the importance of bedrock structure, but found that this alone cannot explain the distinction between whalebacks and roche moutonnées in British Columbia.

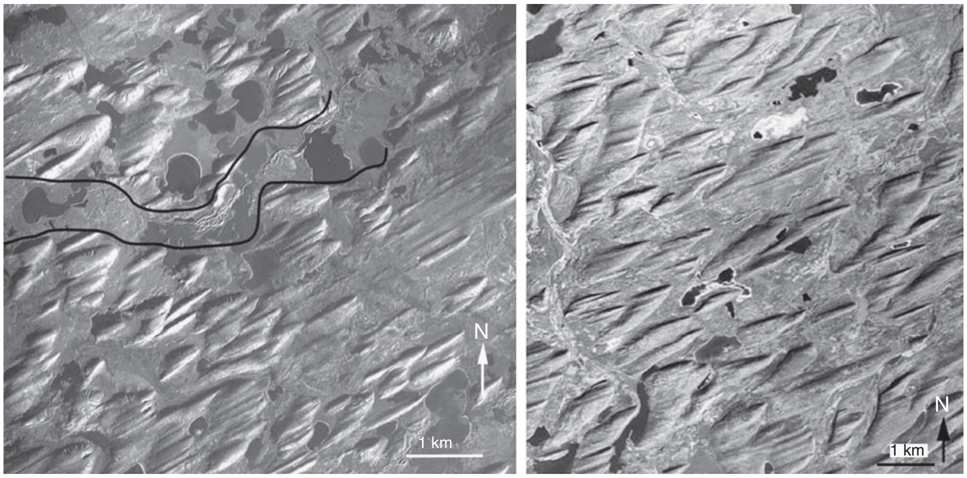
Alternative hypotheses for whaleback formation include erosion by subglacial meltwater and pre-glacial weathering. Based on bedrock surfaces marked by erosional forms interpreted as fluvially formed, Kor et al. (1991) argued that whalebacks in Ontario, North America, were created by fluvial erosion in association with catastrophic subglacial drainage. Bedrock forms similar to whalebacks, formed by pre-glacial weathering and preserved under non-erosive, cold-based ice (cf. Kleman, 1994) have been reported from southern Scandinavia and Minnesota, North America (Lindström, 1988; Lidmar-Bergström, 1997; Olvmo et al., 1999; Patterson and Boerboom, 1999; Johansson et al., 2001), and also from non-glaciated regions in Africa and Australia (Lindström, 1988). This has been presented as an argument in support of the idea that very limited glacial erosion may be enough to produce whalebacks in areas where pre-glacial weathering produces smooth weathering fronts.

8.8.3.2 Drumlins, Crag and Tails, and Large-Scale Flutings

Classical drumlins (**Figure 9**) are streamlined and are arranged in fields, some containing tens of thousands of individual landforms. They are typically asymmetrical in plan, highest at their proximal blunt ends, and taper in a downflow direction on their lee sides. There have been several theories of drumlin formation proposed including, but not limited to, subglacial deformation (Boulton, 1987), dilatancy (Smalley and Unwin, 1968), subglacial bedform formation and subglacial meltwater erosion and deposition (Shaw et al., 1989). The meltwater erosion hypothesis for drumlin formation holds that drumlins are created either by direct fluvial erosion of the bed (Beverleys), or by fluvial infilling of cavities formed as erosional marks in glacier beds (Livingstones) (e.g., Shaw, 1996) (**Figures 5** and **6**).

It is generally thought that drumlin streamlining represents minimum resistance to flow, but this is only true for flows of high Reynolds Numbers, i.e., turbulently flowing water (Shapiro, 1961). Hairpin furrows or horseshoe-shaped scours are commonly wrapped around the proximal ends of Beverleys and crag and tails (**Figure 7**) (e.g., Shaw, 1994), in a similar fashion to the s-forms previously described. Analogous forms in turbulent flow are produced by vortex erosion around bridge piers (Dargahi, 1990) and on the upstream side of scour remnant ridges (Allen, 1982). Thus, streamlining and hairpin troughs support the hypothesis of drumlin formation by turbulent meltwater.

Crag and tails are elongated streamlined hills (**Figure 7**) that are the result of erosion by ice on their upflow end and preservation of existing sediment or bedrock, or deposition of sediment on their downflow end. A classic example of a crag and tail is Castle Rock, Edinburgh, Scotland, the site of the famous Edinburgh Castle (**Figure 7**). Castle Rock is a plug of resistant basalt that was originally a volcanic vent. As glacial ice moved over the plug from west to east, it preferentially eroded the upflow facing slope by abrasion and quarrying, but the less resistant sedimentary rocks behind the crag were left as



(a) (b)

Figure 5 (a) Parabolic and spindle shaped drumlins (Livingstones), Livingstone Lake area, northern Saskatchewan. A tunnel channel containing eskers is indicated by the black lines. Flow from northeast. (b) Transverse asymmetrical drumlins, Livingstone Lake area, northern Saskatchewan. Flow from northeast.

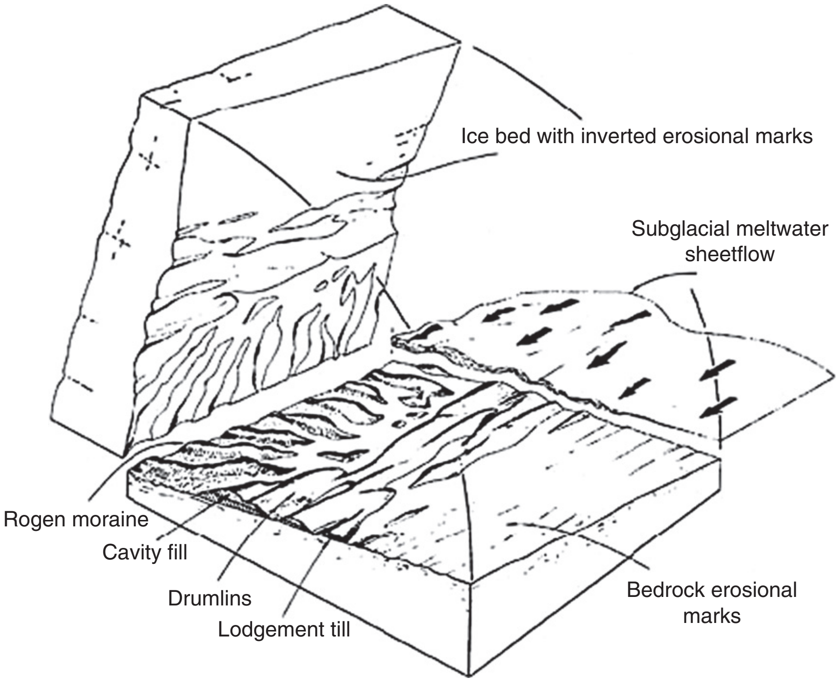


Figure 6 Landform formation by broad, subglacial meltwater flow. Note erosional and depositional features.

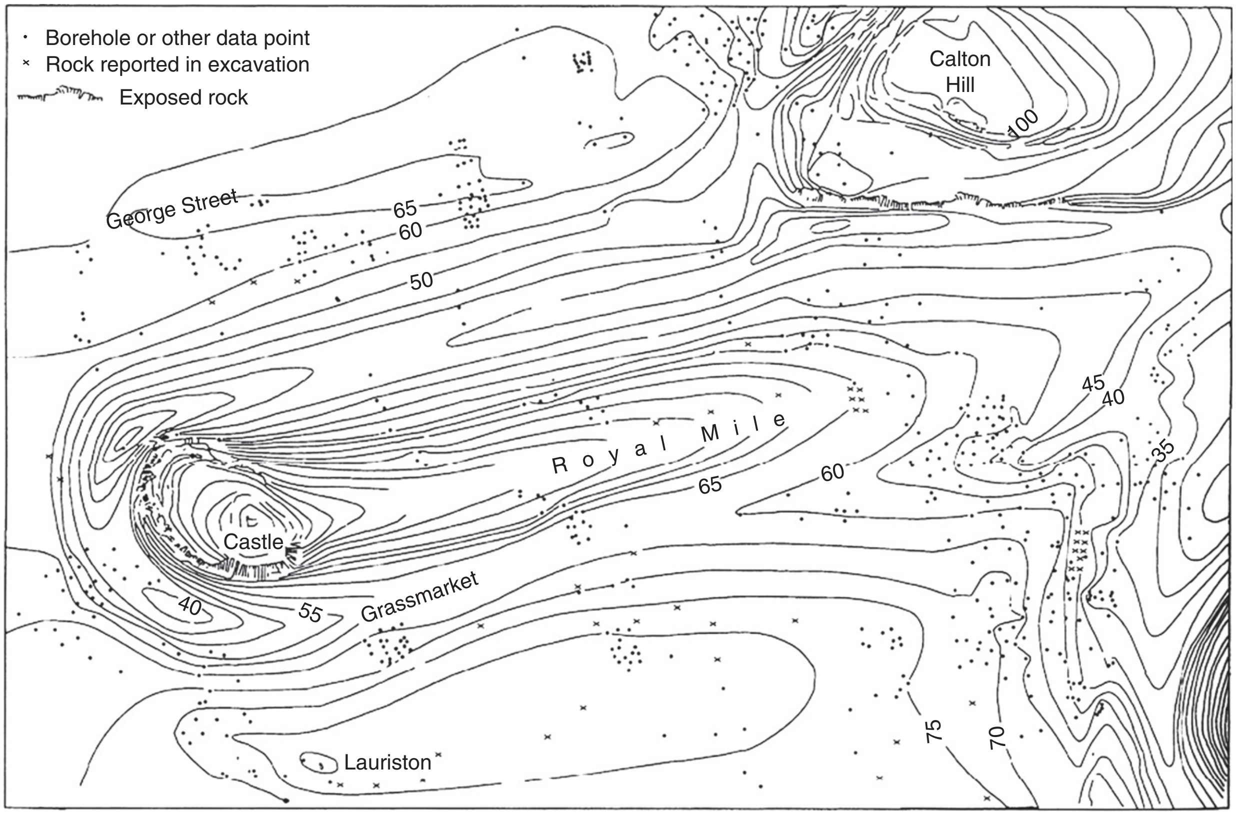


Figure 7 Map showing location of Castle Rock, Edinburgh, Scotland (the location of Edinburgh Castle). The western slope represents the steep erosional crag; the eastern long portion of the landform represents the depositional tail. An erosional trough is also present at the bottom of the western slope.

a residual tail. In several locations tails are composed of sediment deposited into a cavity in the lee of the crag.

Sometimes called mega-scale glacial lineations, large-scale fluting can be tens of kilometers long and a few hundred meters wide (Boulton, 1976; Shaw et al., 2000). Flutings are commonly transitional from drumlins in the same field and as such they are part of a continuum. They have been attributed to various factors including sediment deformation (e.g., Boulton and Clark, 1990), meltwater erosion (e.g., Shaw et al., 2000), ice-keel plowing (Clark et al., 2003) and squeezing into basal cavities (e.g., Bluemle et al., 1993). The deformation hypothesis, perhaps the most popular, states that where glaciers or ice sheets come into contact with soft beds that are easily deformed, the ice will accelerate over those beds and pervasively deform them and mould them into fluted terrain (e.g., Eyles et al., 1999; Evans, 2000). However, since these landforms are sometimes composed of *in-situ* bedrock or pre-existing undeformed sediment (e.g., Munro-Stasiuk and Shaw, 2002) many of these are interpreted as the product of erosion instead.

A theory of erosion by glacial ice has been attributed to ice keels that plow through the sediments, carving elongate grooves, and deforming material up into intervening ridges (Clark et al., 2003). Clark et al. indicated that this is strong evidence of basal ice streaming over long distances. An alternative theory is erosion by subglacial meltwater. This originates from two main observations. First, high ground downstream from forward facing slopes or steps is commonly fluted (Pollard et al., 1996; Shaw et al., 2000; Munro-Stasiuk and Shaw, 2002) and the internal structure of fluting is typically truncated by the land surface and a boulder lag commonly lies on this surface (Shaw et al., 2000; Munro-Stasiuk and Shaw, 2002).

8.8.3.3 Tunnel Channels

Tunnel channels/valleys are generally long (10^1 km), wide (10 km), flat-bottomed, overdeepened ($10^1 - 10^2$ m), radial or anabranching valley systems. They can be incised into bedrock or sediment, terminate in ice marginal fans, and be empty, partially filled or buried by sediments. In recent years, considerable debate has ensued as to the precise mechanism(s) by which such large valleys formed (e.g., Ó Cofaigh, 1996) and their associate implications for reconstructing ice-sheet hydrology. There are two dominant hypotheses for their formation: (1) the piping (bed deformation) hypothesis (e.g., Boulton and Hindmarsh, 1987) that states that tunnel valleys formed at below bankfull conditions in a headward progression as a saturated substrate dewatered and formed pipes at the ice margin. As sediment was flushed from these pipes and with ice margin retreat, tunnel valleys gradually evolved, growing deeper and longer; and (2) the channelized underburst hypothesis (e.g., Brennand and Shaw, 1994) that states that tunnel channels were incised under bankfull conditions by channelized underbursts (jökulhlaups or megafloods) draining a subglacial or supraglacial meltwater reservoir. Thermal conditions at the glacier sole may have facilitated reservoir growth and drainage (Cutler et al., 2002). Although the piping hypothesis may hold true for some sediment-walled tunnel valleys formed at the ice margin, many observations from tunnel channels/valleys associated with the Laurentide and Cordilleran ice sheets support the channelized underburst hypothesis (e.g., Brennand and Shaw, 1994; Sjogren and Rains, 1995; Beaney and Shaw, 2000; Cutler et al., 2002; Fisher et al., 2005). Regardless, they are the geomorphic expression of large subglacial meltwater flows that efficiently evacuated meltwater and sediment from beneath past ice sheets (e.g., Wright, 1973; Rampton, 2000).

There are many sites in formerly glaciated regions that contain major networks of tunnel valleys, including Germany (Piotrowski, 1997), Denmark (Jørgensen and Sanderson, 2006), Ireland (Eyles and McCabe, 1989), the North Sea Basin (Praeg, 2003; Lonergan et al., 2006; Kristensen et al., 2007), the Scotian Shelf (Boyd et al., 1988) and all along the southern margin of the former Laurentide Ice Sheet (Sharpe et al., 2004; Hooke and Jennings, 2006; Clayton and Attig, 1989). However, no modern observations of tunnel valley formation had been made until major outburst flood from Skeiderarjökull, Iceland in 1996. There, the channelization of a sheet flow underburst feeding the jökulhlaup resulted in tunnel channels (Russell et al., 2001). Although these were smaller than their ancient counterparts, their characteristics nonetheless demonstrated that they formed oblique to local ice flow direction, and both topography and proglacial hydrology played a role in their formation. One channel was 160 m long and ascended 11.5 m over that distance where it joined the apex of an ice-contact fan. Russell et al. (2001) determined that the channel formed via hydro-mechanical erosion of unconsolidated glacial sediment, evidenced by an enormous volume of rip-up clasts in the ice-contact fan.

8.8.4 Large-Scale Erosional Forms

8.8.4.1 Glacial Troughs and Fjords

Glacial troughs and, in particular, fjords are generally the largest and most spectacular landforms created by glaciers (**Figure 8**). Glacial troughs are valleys eroded by glaciers, with smoothed and generally steep valley sides lacking spurs, and rounded 'U-shaped' cross-sections (**Figures 8(a)** and **8(b)**). Fjords are glacial troughs reaching out into the ocean (**Figure 8(c)**), and they typically represent the largest glacial troughs. Glacial troughs range in size from hundred-meter scale troughs in mountain regions to the largest fjords with lengths of several hundred kilometers and maximum depths of a few thousand meters. Glacial troughs occur all around the globe in mountain regions where glacier erosion has been sufficient to re-shape the landscape. Fjords occur in Northern Europe (Norway, Iceland), Greenland, North America (northeastern and northwestern coasts), South America (Chile), Antarctica, and New Zealand, where extensive glaciers have reached the ocean.

The cross-section shape of glacial troughs, being distinctly different from fluvial valleys, has been the focus of several studies investigating glacial erosion. Although glacial troughs are commonly described as U-shaped, it has frequently been pointed out that the cross-section is commonly asymmetrical and better represented by various numerical functions

Figure 8 (a) Glacial trough in New Zealand; (b) glacial trough in the Jotunheimen mountains, central Norway; (c) the outer region of Ranafjorden, Norway.

(Svensson, 1959; Graf, 1970; Hirano and Aniya, 1988, 1989; James, 1996; Li et al., 2001; Morgan, 2005). A commonly used expression to describe glacial trough cross-section shape is the parabolic function $y = ax^b$ (Svensson, 1959; Graf, 1970) where y is height, x is distance to the mid-point of the valley floor, and a and b are constants. In addition to overall shape, a number of researchers have emphasized the need to combine this with the depth/width ratio (form ratio; Graf, 1970), with values for measured cross-sections ranging from 0.1 to 0.45 (Li et al., 2001; Benn and Evans, 2010). Extremely shallow glacial troughs with depth/width ratios less than 0.02 have been identified on the northeastern Tibetan Plateau based on digital elevation model analysis (Heyman et al., 2008).

Hirano and Aniya (1988, 1989) explained the glacial trough cross-section form as a result of glacial erosion working towards minimizing basal friction, and argued that alpine valley glaciers tend to deepen their troughs whereas ice sheets widen their troughs. Harbor et al. (1988) and Harbor (1992) used a physically based numerical ice flow model coupled to an erosion model driven by basal ice velocity to simulate the development of glacial trough cross-sections. The simulations demonstrated that in a perfect V-shaped (pre-glacial) valley, the basal velocity and erosion rate is highest some distance up on the valley sides, and that this can explain how the U-shape is formed (**Figure 9**). Once the valley has attained a U-shaped form the basal velocity will be highest in the mid-valley floor position, and continuous glacial erosion will deepen the valley.

The time required to form glacial trough shapes has been investigated using a number of inventive methods. Nesje et al. (1992) estimated the total volume of glacial erosion in Sognefjord, Norway, and used an estimated duration of ice coverage of 600 000 years to derive erosion rates ranging from 1 to 3.3 mm yr^{-1}. With a coupled ice flow–erosion model 10 000 years of continuous glaciation is required to produce a recognizable glacially eroded trough from a V-shaped form (Harbor, 1992), and to form a steady state U-shaped valley 100 000 years of continuous glaciation is required (Harbor et al., 1988). Similar results were presented by Jamieson et al. (2008) using a 3D ice sheet model to simulate glacial landscape evolution. Brook et al. (2006, 2008) used a marine oxygen isotope record as a proxy for ice extent in New Zealand. Integrating the time of ice coverage for a set of glacial trough cross sections they arrived at durations 400 000–600 000 years of glacial occupancy required for glacial form development.

The importance of the pre-glacial landscape for glacial trough evolution has long been recognized, and the origin of glacial troughs is generally thought of as involving glaciers carving pre-glacial fluvial valleys (e.g., Linton, 1963; Sugden, 1968; Sugden and John, 1976; Haynes, 1977; Benn and Evans, 2010). Mapped tectonic lineaments on Ellesmere Island and around Sognefjord, Norway, have been shown to correlate well with glacial trough orientation (Randall, 1961; England, 1987; Nesje and Whillans, 1994), indicating a lithological control on glacial trough development. Glasser and Ghiglione (2009) mapped an extensive area of the Patagonian Andes and showed that fjord orientations of South America are largely similar to tectonic lineaments. Based on this they concluded that the primary control on fjord development in glaciated areas is geological rather than glaciological. Similarly, field measurements of rock mass strength in New Zealand and the northern British islands have been shown to correlate with glacial trough cross-section form (Augustinus, 1992a, 1995;

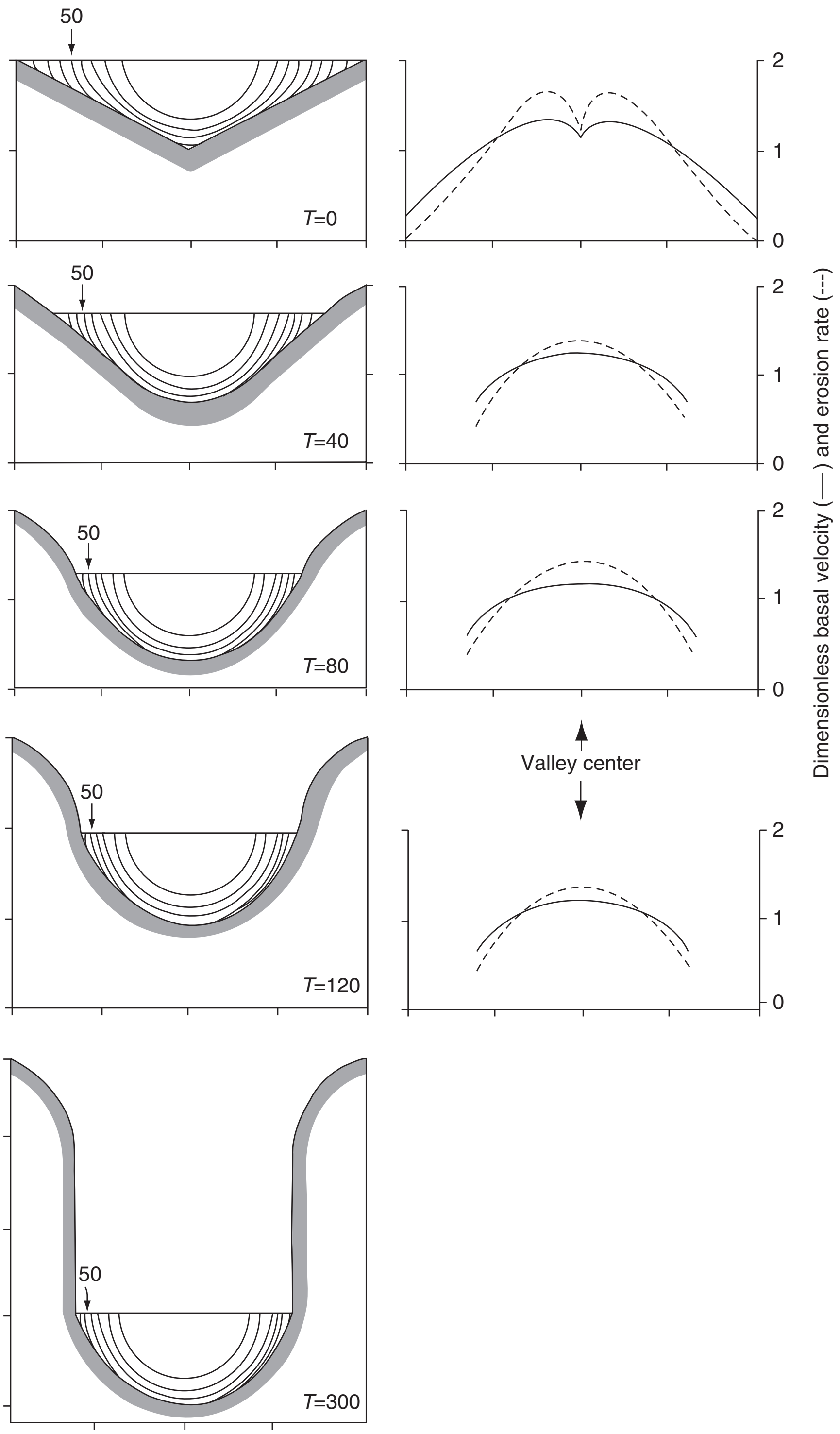


Figure 9 Glacial erosion remolding a V-shaped valley into a glacial trough (Harbor, 1992). In an ideal V-shaped valley, the basal ice velocity (shown as percentage of maximum with 10% interval lines), and thus the erosion rate, is higher some distance up the valley side. This causes widening of the lower parts of the valley and development of a U-shaped parabolic cross-section.

Brook et al., 2004), with deep narrow valleys being more easily developed in bedrock with high rock mass strength whereas a low rock mass strength is favorable for development of shallow wide U-shaped valleys. Harbor (1995) simulated the development of glacial trough cross sections under variable resistance to erosion (rock mass strength dependent) and illustrated that varying erodibility can result in a wide range of cross-section forms. The importance of topography and relief has also frequently been stressed in studies of trough development (e.g., Sugden, 1978; Kleman and Stroeven, 1997;

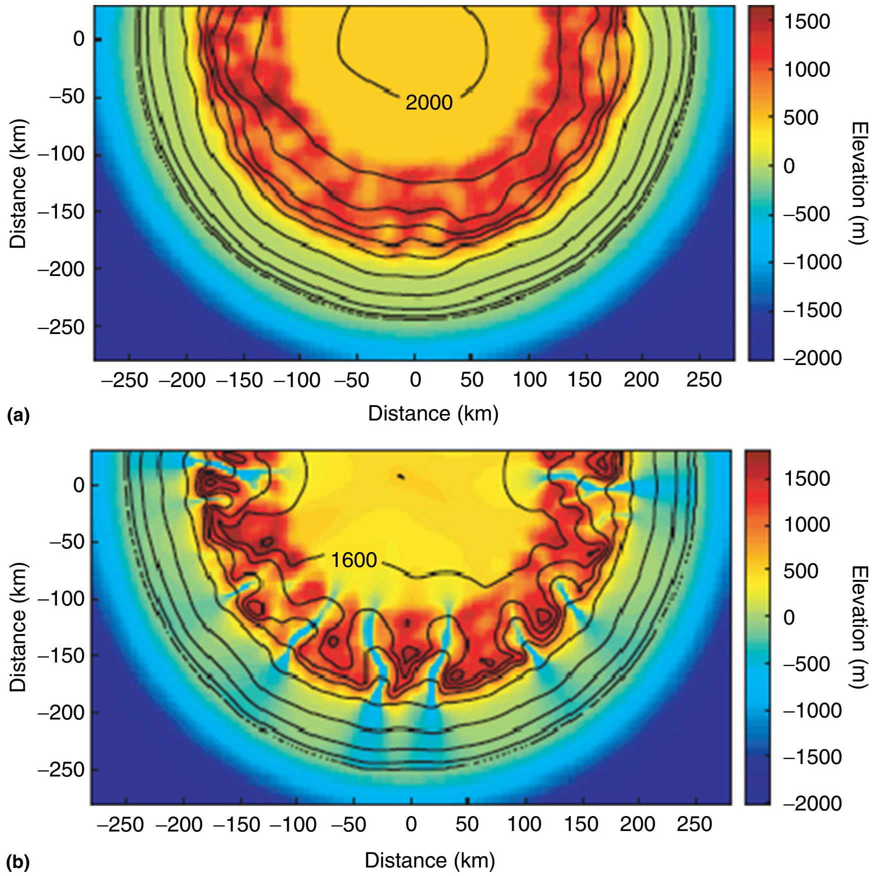


Figure 10 Simulated formation of fjords by topographic steering of ice. Reproduced from Kessler, M.A., Anderson, R.S., Briner, J.P., 2008. Fjord insertion into continental margins driven by topographic steering of ice. Nature Geoscience 1, 365–369: (a) Initial non-glacial topography with a mountain range separating a low relief high elevation area and the sea; (b) Topography after 1.2 million years of ice sheet glaciation. The topographic steering of ice over/through the mountain range into the ocean, with highest ice flow and erosion in the lower regions of the mountains, has created deeply incised glacial troughs and fjords.

Näslund, 1997; Jamieson et al., 2010), with the following basal thermal regime feedback chain enhancing the original relief: low lying area – thicker ice – warmer basal ice (more likely melted) – more glacial erosion. For Greenland, British Columbia, and New Zealand the area contributing ice to a glacial trough has been shown to correlate with glacial trough size (Haynes, 1972; Roberts and Rood, 1984; Augustinus, 1992b). Using a simple ice sheet–erosion model, Kessler et al. (2008) demonstrated that for ice flowing over/through a low relief mountain range into the ocean topographic steering alone can explain the formation of deep fjords (Figure 10).

The development of new chronological tools has allowed some hypotheses about glacial erosion and trough development to be tested. Cosmogenic exposure dating has been used to quantify or constrain glacial erosion, with multiple studies confirming low erosion rates at glacial valley interfluves and much higher erosion depths on glacial valley floors (e.g., Stroeven et al., 2002; Fabel et al., 2002; Li et al., 2005; Sugden et al., 2005; Phillips et al., 2006; Briner et al., 2006, 2008). Swift et al. (2008) used thermochronology to argue that the fjords in east Greenland have been controlled by geology and pre-glacial topography. Thermochronology data and numerical modeling has also been used to argue for Quaternary exhumation and relief growth in the Alps (Glotzbach et al., 2011) and for headward propagation of glacial erosion in New Zealand (Shuster et al., 2011).

8.8.4.2 Rock Basins

Rock basins are some of the most impressive and characteristic features of landscapes that are shaped by glacial erosion. As many rock basins are now partially filled with water (e.g., Figure 11), they are distinctive features of the landscape and may also contain sediments that can be used to constrain the minimum age of the last glaciation to occupy the basin, as well as to reconstruct post-glacial variations in geomorphic processes and climate.

In the context of glacial geomorphology, rock basins are bedrock surfaces where the topography has a local low point or area as a result of preferential erosion. They range in scale from small depressions on the order of meters in length, to the Great Lakes with lengths of hundreds of kilometers. Over the past two centuries considerable debate and research has focused on the mechanisms and controls that produce local overdeepening as a result of glacial erosion. From this work it is clear that there are a number of controls that are important,

Figure 11 Glaciated valley in the Northern Swedish Mountains with a large overdeepened rock basin now occupied by a lake.

Figure 12 Glacially scoured bedrock with small enclosed depressions now filled with water, Northern Sweden.

including glaciological variables such as thermal regime and basal stress and velocity patterns, and bedrock variables such as structure and lithology (reviewed, for example, in Sugden and John, 1976, and Benn and Evans, 2010). The glaciological variables control the potential for erosion (are the basal conditions suitable for plucking, abrasion, and the removal of erosion products, and are there spatial variations in potential erosion rate?) and the bedrock variables control the propensity for erosion (under given glaciological conditions, how easily is the rock material removed, and are there spatial variations in erodability?). Taken together, this produces both variations in the reasons for particular rock basins as well as a wide variability in the patterns and scales of rock basins.

8.8.4.3 Knock and Lochain

Glacial geomorphology is replete with interesting terms derived from very local names for landscape features, and this includes knock and lochain morphology (Linton (1963), from Scots Gaelic words for 'knoll' and 'small lake') which refers to large areas of subglacially eroded bedrock that include rock knobs, roches moutonnées and rock basins (**Figure 12**) with relief amplitudes typically less than $\sim$100 m (Benn and Evans, 2010). The overall location of areas of knock and lochain topography is controlled by large-scale ice sheet dynamics, as this topography is found in areas of areal scouring. This requires wet-based ice where flow is laterally extensive, rather than focused into preferential flow paths. Beyond this large-scale glaciological control, which is a function of ice sheet dynamics, the details of exactly where the bedrock is higher and lower in the landscape is controlled primarily by patterns of strength in the bedrock. Basins occur where the bedrock is locally less resistant to erosion, because of a weaker lithology and/or a higher joint density.

8.8.4.4 Glacial Lakes

At a larger scale, but in similar overall glaciological settings, glacial erosion by ice sheets can produce large groups and swarms of lakes in a landscape of pervasive glacial erosion. The Finger Lakes of New York State are a classic example of large, elongated lake basins preferentially eroded by ice sheet glaciation, and include seven major lakes up to 60 km in length. The Finger Lakes have their origins in a series of northward-flowing river valleys that were then preferentially eroded by southward flowing ice during one or more glaciations. Several examples of oriented swarms of lakes are given in Sugden and John (1976) from Sweden, Minnesota, and the Canadian Shield, and in each case the details of the locations and the elongation patterns of the lakes relate to lithological and structural factors (faults, joints, dikes, fracture patterns, and areas of less resistant rock or junctions between rock types). On an even larger scale, the Great Lakes of North America also developed as a result of the interaction of glaciological controls and bedrock resistance to erosion. As reviewed in Larson and Schaetzl (2001), the Great Lakes developed as a result of ice sheet and ice lobe flow being channeled along major pre-glacial valley systems, resulting in preferential glacial erosion in areas of relatively weak rock.

8.8.4.5 Cirques and Overdeepenings in Glacial Valleys

Mountain areas that have been shaped by past alpine glaciation also include frequent lakes. At valley heads and in amphitheater-shaped depressions cut in to mountainsides by small ice patches (cirques), overdeepening in association with past glacial action has produced depressions now commonly filled with small lakes (tarns), for example, **Figure 13**. Studies of active cirque glaciers (e.g., Grove, 1960a, b) have shown that overdeepening is enhanced by a rotational ice flow pattern in combination with a supply of tools for abrasion in the form of rockfall into the randkluft (a gap between the glacier ice and the bedrock making up the back wall of the cirque) or the bergschrund (a crevasse that forms at the point where moving cirque glacier ice separates from stagnant ice).

Glaciated valleys commonly have tarns at their heads, where former glaciers may have begun and finished their period of existence as cirque forms. In addition, there are generally overdeepenings along the length of a glaciated valley. Some overdeepened basins form where there is a localized increase in ice discharge where tributary glaciers enter a main valley glacier, or where ice velocities increase through narrow valley sections and then decrease as the valley widens. It is also clear that some overdeepenings are responses to localized

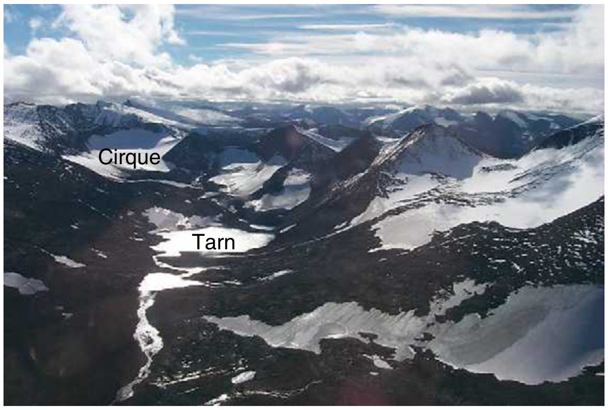


Figure 13 Alpine cirque glaciers and a valley head tarn, Northern Swedish Mountains.

areas of higher rock erodability that relate to bedrock structure and lithology; these areas may have been partially excavated by non-glacial processes before the onset of glaciation. Once an overdeepening has formed, positive feedback related to glaciological controls enhances the overdeepening. Based on observations from Storglaciaren in Sweden, Hooke (1991) argued that surface crevasses generally form just down-glacier from bumps (i.e., the up-glacier end of a depression) and this allows surface meltwater to reach the glacier bed, which will increase water pressure fluctuations that are critical for rock fracture and glacial plucking. Thus a depression creates a glaciological feedback that enhances erosion in the depression.

8.8.4.6 Streamlined Hills

Large streamlined hills lie within, or at the junction between, swales (elongated depressions) within fluted terrain in many regions. A good example is the Hand Hills in Alberta. Young et al. (2003) referred to the Hand Hills as a mega-drumlin because the hills, as a complex, resemble a drumlin in shape and profile. In this case the mega-drumlin rises 70 m above the prairie land surface and has a length of approximately 3 km (smaller than most streamlined hills, but significantly larger than drumlins). Like many drumlins of usual dimensions this mega-drumlin even has a large horseshoe-shaped scour at the base of its upflow-facing side.

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Biographical Sketch

Mandy J Munro-Stasiuk's research focuses on extreme environments. Munro-Stasiuk has worked predominantly on ancient and contemporary glacial landscapes in Canada, USA and Iceland, and in particular subglacial meltwater processes. Recently, she has started research in Mexico, on karst sinkholes and their use by the ancient Maya. Like most geographers, she has a skillset in geospatial technology, specifically, satellite remote sensing, terrain modelling and ground penetrating radar analysis as well as an avid interest in earth science education.

Captured by the beauty of glaciers and glacial landscapes on Svalbard in 2001–2002, Jacob Heyman finished his undergraduate thesis focusing on marginal moraines in the Swedish mountains at Stockholm University 2005. Heyman then shifted focus to the glacial history of the Tibetan Plateau which was the topic of his PhD project (2005–2010). Since February 2011 Heyman has been working as a postdoc at Purdue University focusing on glaciation and erosion of the Tibetan Plateau. The tools he has used to resolve the history of past glaciations include remote sensing (geomorphological mapping from satellite imagery and digital elevation models), field investigations, cosmogenic exposure dating, and numerical modelling.

An early fascination with glacial landscapes acquired from hiking through the English Lake District has turned in to a very enjoyable career in research and teaching. Jonathan Harbor's 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 Harbor 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 study in a wide range of locations, including many parts of North America, the Alps, northern Fennoscandia, the Tibetan Plateau and central Asia.