UNIVERSITY OF THE ARCTIC

Module 5
The Land

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Key Terms and Concepts

- endogenic processes
- exogenic processes

Glaciers and Glaciology

- alpine (valley) glaciers
- cirque glaciers
- continental glaciers
- firn
- glacier ice
- glacier mass balance
    - – accumulation and accumulation zone
    - – ablation and ablation zone
    - – equilibrium line
- glacier motion
    - – internal deformation and Glen’s Flow Law
    - – basal sliding
    - – soft sediment deformation
    - – extending flow
    - – compressive flow
- glacial erosion
    - – abrasion
    - – plucking
    - – hydrolysis
    - – carbonation
    - – oxidation

- glacial transport
    - – traction
    - – regelation
    - – supraglacial
    - – englacial
    - – subglacial
- glacial deposition
    - – ablation (melt out)
    - – lodgement

Alpine Landforms

- erosional landforms
    - – striations
    - – roches moutonnées
    - – cirques
    - – glacial troughs
    - – fiords
- depositional landforms
    - – lateral moraines
    - – end moraines
    - – hummocky moraine
    - – ground moraine
- glaciofluvial sediments and landforms
    - – eskers
    - – braided stream channels and sandbars

Permafrost and Periglacial Landscapes

- permafrost
    - – permafrost table
    - – active layer
    - – level of zero amplitude
    - – talik
    - – continuous permafrost
    - – discontinuous permafrost

  - sporadic permafrost

Periglacial Processes and Landforms

- frost shattering
  - block fields
  - felsenmeer
  - grus
- ground ice
  - patterned ground
  - segregated ice
    - frost heave
    - cryoturbation
    - non-sorted circles
    - palsas and peat plateaux
  - frost cracking
    - low-centred ice wedge polygons
    - high-centred ice wedge polygons
  - mass movement
    - frost creep
    - gelifluction
    - solifluction lobes
    - stripes
  - intrusive ice
    - open-system pingos
    - closed-system pingos

Learning Objectives/Outcomes

This module should help you to

1. develop an understanding of the distribution of glaciers and permafrost in the circumpolar North.
2. develop an understanding of the processes that influence the mass balance and flow of glaciers.
3. develop an understanding of the processes of glacial erosion, transport, and deposition.
4. develop an understanding of the morphology, composition, and distribution of glacial landforms.

5. develop an understanding of glaciofluvial processes and landforms.

6. develop an understanding of the processes that contribute to the development and persistence of permafrost in northern environments.

7. develop an understanding of various periglacial processes and the landforms created by the processes.

8. provide concise definitions for the key words.

Reading Assignments

Young (1989), chapter 4, "Glaciers and Glaciology," in *To the Arctic: An Introduction to the Far Northern World*, 65–83.

Young (1989), chapter 5, "Polar Landscapes: Glacial Geology and Geomorphology," in *To the Arctic: An Introduction to the Far Northern World*, 85–109.

Young (1989), chapter 6, "The Periglacial Environment," in *To the Arctic: An Introduction to the Far Northern World*, 111–136.

Young (1989), chapter 7, "Ice Ages," in *To the Arctic: An Introduction to the Far Northern World*, 137–149.

Overview

Variations in the topographic relief of the Earth's surface reflect the constant interaction of *exogenic* and *endogenic* processes. Endogenic processes are responsible for creating topographic relief (i.e., variations in the elevation of the land surface relative to sea level). These processes include volcanism, earthquakes, and the folding and faulting of rocks in the Earth's crust. On the other hand, exogenic processes operate to reduce topographic relief via the weathering of rock and the subsequent redistribution of the weathering products, sediments, and dissolved minerals at or near the Earth's surface. The processes of weathering, erosion, transport, and deposition are collectively referred to as *denudation*. In this module we will focus on exogenic processes, and we will explore the roles of glaciers and glacial processes, as well as those of permafrost and periglacial processes, in shaping northern landscapes.

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Lecture

Glaciers and Glaciology

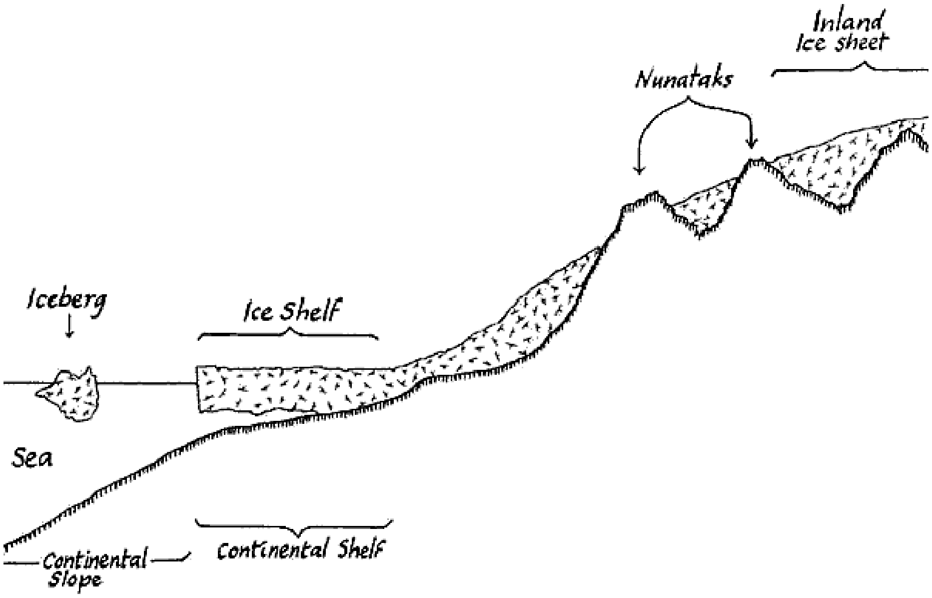
A glacier can be defined as a large, naturally occurring deposit of perennial ice, formed from the accumulation and recrystallization of snow, that is capable of flowing slowly under the pressure of its own weight and the force of gravity. Glaciers cover approximately 10% of the present surface of the Earth (see table 5.1); at the Last Glacial Maximum (ca. 30,000 BP) glaciers covered approximately 30% of the Earth's surface. Glaciers occupy highlands where winter snowfall is heavy and summer temperatures are not adequate to completely melt the snowpack. Glaciers serve as active geomorphic agents capable of eroding, transporting, and depositing immense quantities of rock debris in polar and alpine landscapes.

Table 5.1 The relative extent of modern glaciers (after Sugden and John 1976)

Region	**Area (km^2)**
South Polar Region	
Antarctic Ice Sheet	12,585,000
Subantarctic Islands	3,000
North Polar Region	
Greenland Ice Sheet	1,802,600
Canadian Arctic Archipelago	153,169
Iceland	12,173
Svalbard	58,016
Other Arctic Islands	55,658
North America	
Alaska	51,476
Canadian and American Cordillera	25,404
South America	26,500
Europe	
Scandinavia	3,810
Alps	3,600
Caucasus	1,805
Other Mountain Ranges	61
Asia	
Himalaya	33,200
Other Mountain Ranges	81,821
Pacific Region	
Japan and New Zealand	1,015

Types of Glaciers

Glaciers are generally classified into three types: alpine (valley) glaciers, piedmont glaciers, and continental glaciers. Alpine glaciers develop in highland regions and are constrained by topography, being confined to mountain valleys. Piedmont glaciers form when the lower reaches of alpine glaciers coalesce and spread over lowlands at the foot of a mountain range. These two forms of glaciers cover an area ranging from 5 to 10,000 km^2. Continental glaciers, on the other hand, cover a vast area exceeding 25,000 km^2. These large glaciers tend to inundate the underlying topography; however, the tallest mountain tops may protrude through the glaciers as *nunataks* (see fig. 5.1).



Source: After Pielou (1994)

Fig. 5.1 A longitudinal cross-section of an ice sheet discharging glacier ice to the sea via outlet valley glaciers

Glacier Mass Balance

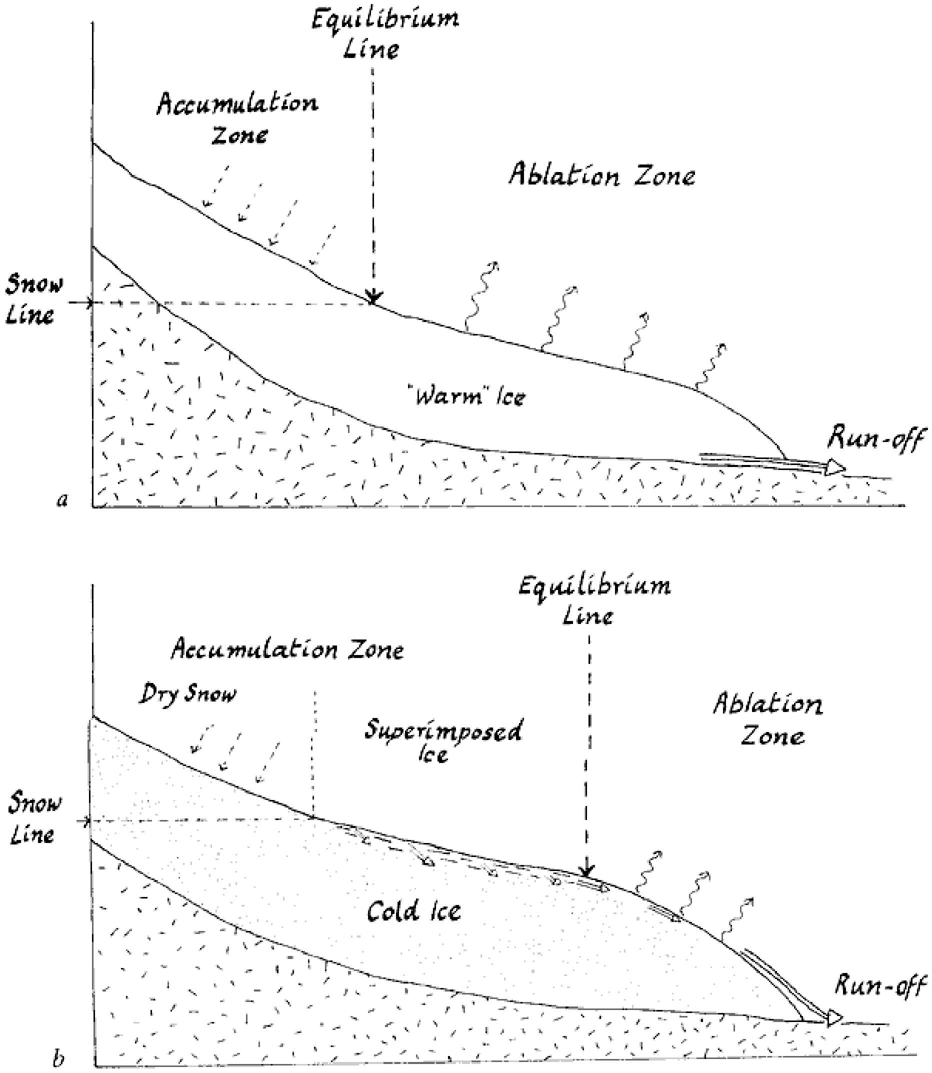
The movement of a glacier, hence its ability to serve as a geomorphic agent, is influenced by the interaction of two suites of processes: accumulation and ablation. Accumulation processes include direct precipitation as snow, the refreezing of meltwater, and sublimation (i.e., the physical transformation of water vapour into ice). These processes jointly serve to increase the mass and volume of glaciers. Ablation processes include the melting of snow and ice via insolation, friction associated with internal deformation of the glacier, the flow of geothermal heat, sublimation (i.e., the physical transformation of glacier ice into water vapour), and the calving of icebergs. These processes jointly serve to reduce the mass and volume of glaciers.

Glacier mass balance refers to the annual variations in the mass and volume of glaciers that results from the interaction of accumulation and ablation processes. Stated simply,

Mass balance = accumulation – ablation

When the volume of snow and ice that accumulates on a glacier exceeds the volume of snow and ice removed by ablation (i.e., when accumulation is greater than ablation), the glacier is said to exhibit a positive mass balance. The glacier thickens and develops a steep surface slope, and the glacier margin advances (see Glacier Ice Movement later in this lecture). Conversely, when the volume of snow and ice that accumulates on a glacier is less than the volume of snow and ice removed by ablation (i.e., when ablation is greater than accumulation), the glacier is said to exhibit a negative mass balance. The glacier thins and develops a gentle surface gradient, and the glacier margin retreats.

The surface of a glacier can be divided into zone of accumulation (i.e., positive mass balance) at higher elevations and a zone of ablation (i.e., negative mass balance) at lower elevations separated by the equilibrium line (i.e., accumulation equals ablation) (see fig. 5.2).



Source: After Pielou (1994)

Fig. 5.2 Longitudinal cross-sections of “warm-based” and “cold-based” alpine glaciers illustrating the position of the accumulation zone, the ablation zone, and the equilibrium line

Origins of Glacier Ice

Glacier ice originates primarily as snow deposited on the glacier surface. Over time, snow is gradually transformed into *firn* via the combined action of compaction and *sintering*. Firn is eventually converted into glacier ice. The transformation of fresh snow into glacier ice involves the progressive compression of air bubbles and a reduction in porosity, resulting in an increase in the density of the material. (See table 5.2.)

Table 5.2 Density of snow, firn, and glacier ice

	Density (kgm^{-3})
Fresh snow	50–100
Firn	400–800
Glacier ice	830–910

Glacier Ice Movement

Glaciers serve as a transportation system moving snow, ice, and rock debris from the *accumulation zone* and disposing of these materials in the *ablation zone*. Three main processes contribute to the motion of glacier ice:

- *Internal deformation.* This process involves the slippage of ice crystals along planes within the glacier in response to an applied shear stress.
- *Basal sliding.* This process involves the glacier ice sliding along the underlying bedrock surface on a thin layer of meltwater. The process operates only where the temperature of the basal ice lies close to the melting point for ice (i.e., a "warm-based" glacier), allowing for the presence of liquid water at the base of the glacier.
- *Soft sediment deformation.* This process involves glacier ice sliding on a layer of deforming saturated, fine-grained sediments at the glacier bed. Shear stress exerted by the moving glacier exceeds the shear strength of subglacial sediments; deformation occurs within the sediments allowing the glacier to slide forward across its bed. As is the case for basal sliding, this process operates only beneath "warm-based" ice.

Ice temperatures influence all three processes. In *warm-based* or temperate glaciers, ice temperatures near the bed are near the melting point for ice. Warm ice deforms easily, resulting in a higher rate of strain. Warm ice also permits meltwater generated at the surface or within the glacier to move through the glacier to lubricate the bed, thus facilitating basal sliding and soft sediment deformation. *Cold-based* or *polar* glaciers, on the other hand, exhibit basal ice temperatures well below the melting point of ice; these glaciers are frozen to their beds. Motion of these glaciers is via internal deformation only.

Internal Deformation

A glacier flows in response to pressure generated by the weight of the glacier itself and shear stress generated by the force of gravity. The magnitude of the shear stress (τ_o) at any depth within a glacier is expressed as:

$$\tau_o = \rho_i g h \sin\beta$$

Where: ρ_i = ice density (kgm^{-3})
g = acceleration due to gravity (ms^{-2})
h = ice thickness (m)
β = slope of the glacier surface

An equation known as *Glen's Flow Law* relates the shear stress that glacier ice experiences to the rate of internal deformation (i.e., strain, ε). Glen's Flow Law is expressed as:

$$\varepsilon = A_i \tau_o^{\,n}$$

Where: ε = rate of strain (internal deformation)
A_i = creep co-efficient
n = creep exponent; value ranges from 1.5 to 4

The magnitude of the creep coefficient, A_i, is influenced by ice temperature and the quantity of sediment frozen within the ice. The creep exponent, n, is influenced primarily by the nature of the underlying bedrock topography. It is important to note that Glen's Flow Law is represented by an exponential equation. This means that a small change in shear stress (e.g., associated with a change in annual mass balance) will initiate a disproportionately large change in the rate of ice motion via internal deformation.

Rates of Glacier Movement

Ice thickness and velocity increase from the head of the glacier towards the equilibrium line within the accumulation zone. Acceleration of ice in this portion of a glacier is referred to as *extending flow*. Ice thickness and velocity decrease from the equilibrium line towards the ice margin within the ablation zone. Deceleration of ice in this portion of a glacier is referred to as *compressive flow*. Ice velocity also varies across the width and depth of a glacier. Transverse ice velocities (i.e., observed across the width of a glacier) are greatest near the centre line of a glacier and decrease towards the ice margin. Longitudinal ice velocities (i.e., observed throughout the depth of a glacier) tend to increase away from the bed but remain fairly constant as the surface of the glacier is approached. The deceleration of ice near the bed and margins of the glacier reflects the influence of friction between the moving glacier and the rocks and/or sediments that underlie the base of the glacier. Average ice velocities range from 10 to 200 m annually but can exceed 1,000 m annually in exceptional circumstances, such as in *glacial surges*.

Glacial Processes

Glacier Erosion

Erosion of bedrock by glacier ice involves two processes, *abrasion* and *plucking*. Abrasion is a mechanical weathering process. Glaciers use rock clasts at their base as an abrasive. Rock clasts are moved in traction at the bed, their sharp edges cutting *striations* into bedrock, releasing fine-grained rock debris (i.e., *rock flour*) that polishes bedrock surfaces (see fig. 5.3). The rate of abrasion is influenced by variations in ice pressure and velocity, the concentration of rock clasts near the base of the glacier, and the hardness of the rock clasts relative to the bedrock substrate. Thick, fast-flowing ice armed with hard (e.g., igneous and metamorphic) rock clasts promotes rapid and deep abrasion of the bedrock substrate.

Source: Natural Resources Canada. GSC Photo Number 2002-508, http://sts.gsc.nrcan.gc.ca/clf/landscapes_details.asp?numero=756.
Photograph by Lynda Dredge

Fig. 5.3 Glacial striations in granitoid gneiss at Scarpa Lake, Melville Peninsula, Nunavut

Plucking is a mechanical weathering process that involves the removal of large blocks of rock from the glacier bed. Plucking occurs where glacier ice is forced to traverse bedrock obstacles at the bed. In this situation ice pressures exerted on bedrock obstacles are greater on the upstream side of the obstacle than ice pressures exerted on the downstream side of the obstacle, causing the rock to fracture on the downstream side of the obstacle. Higher ice pressures on the upstream side of the obstacle promote pressure melting of basal ice and the generation of meltwater. This meltwater lubricates the glacier bed and allows the glacier to slide over the obstacle. On the downstream side of the obstacle, the meltwater refreezes, a process known as *regelation*, and loosened blocks of rock are removed by freezing to the base of the glacier (see roches moutonnées later in this lecture).

Meltwater Erosion

Glacier meltwater also contributes to mechanical and chemical weathering of bedrock substrates. Beneath warm-based glaciers, turbulent, fast-flowing water—carrying large sediment loads derived from the melting of debris-rich basal ice—can actively abrade bedrock surfaces. High rates of erosion via abrasion are most likely to occur during periods of high meltwater discharge in the summer (see Module 3).

Chemical weathering of bedrock occurs via the processes of *hydrolysis*, *carbonation*, and *oxidation*. Examples of these processes are presented below.

- **Hydrolysis:** This is the reaction between rock minerals and the H^+ and OH^- ions in water.

1.
$$\underset{\text{albite}}{4NaAlSi_3O_8} + \underset{\text{water}}{6H_2O} \rightarrow \underset{\text{kaolinite}}{Al_4Si_4O_{10}(OH)_8} + \underset{\substack{\text{sodium}\\\text{ions}}}{4Na^+} + \underset{\substack{\text{hydroxyl}\\\text{ions}}}{4OH^-} + \underset{\text{silica}}{8SiO_2}$$

- **Carbonation:** This is the reaction between rock minerals and carbonic acid (H_2CO_3) in solution. Carbonic acid is produced when carbon dioxide from the atmosphere or soil dissolves in water.

2.
$$\underset{\text{albite}}{4NaAlSi_3O_8} + \underset{\text{water}}{6H_2O} + \underset{\substack{\text{carbon}\\\text{dioxide}}}{4CO_2} \rightarrow \underset{\text{kaolinite}}{Al_4Si_4O_{10}(OH)_8} + \underset{\substack{\text{sodium}\\\text{ions}}}{4Na^+} + \underset{\substack{\text{bicarbonate}\\\text{ions}}}{4HCO_3^-} + \underset{\text{silica}}{8SiO_2}$$

3.
$$\underset{\text{limestone}}{CaCO_3} + \underset{\text{water}}{H_2O} + \underset{\substack{\text{carbon}\\\text{dioxide}}}{CO_2} \rightarrow \underset{\substack{\text{calcium}\\\text{ions}}}{Ca^{2+}} + \underset{\substack{\text{bicarbonate}\\\text{ions}}}{2HCO_3^-}$$

- **Oxidation:** This is the reaction between rock minerals and water. Oxidation occurs when an atom or ion loses an electron and takes on a positive charge. Oxygen is a common oxidizing agent.

4. $$\underset{\text{pyrite}}{\mathrm{2FeS_2}} + \underset{\text{water}}{\mathrm{7H_2O}} + \underset{\text{oxygen}}{\mathrm{15O}} \rightarrow \underset{\substack{\text{iron}\\\text{hydroxide}}}{\mathrm{2Fe(OH)_3}} + \underset{\substack{\text{sulphuric}\\\text{acid}}}{\mathrm{4H_2SO_4}}$$

Some of these weathering products (e.g., silica, iron hydroxides) can be precipitated onto the downstream side of bedrock obstacles where subglacial meltwater refreezes. However, many of the weathering products (e.g., Na^+, HCO_3^-) are transported as dissolved ions in meltwater flowing in channels beneath the base of the glacier to be discharged into proglacial stream channels.

Glacier Transport

Glacial erosion continues only if rock debris is entrained and transported away by moving glacier ice. In the *supraglacial environment* (i.e., on the surface of the glacier), rock debris is delivered via mass wasting (e.g., avalanches) and aeolian processes onto the surface of the glacier. In the accumulation zone of the glacier, this *supraglacial debris* moves towards the glacier bed under *extending flow* as the surface is progressively buried by subsequent snowfalls. This transport of rock debris through the glacier occurs within the *englacial environment*. Englacial transport of rock debris serves to provide a continuous supply of new cutting tools to the base of the glacier, thus facilitating abrasion of the bedrock substrate. In the ablation zone, rock debris delivered to the glacier surface is transported at the surface towards the glacier margin (see fig. 5.4).

Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5.4 Supraglacial debris accumulating on the snout of outlet valley glacier

In the *subglacial environment* (i.e., at the base of the glacier) rock debris is entrained via a traction force and/or regelation. Deformation of moving glacier ice applies a *tractive force* to rock clasts on the bed. If this force is sufficient to overcome friction between the clast and the bed, the particle will be transported in *traction* across the bed of the glacier. The alternate thawing and freezing of basal ice as the glacier moves over obstacles on the bed allows for debris to be entrained via regelation. Beneath the ablation zone of a glacier, basal debris-rich ice moves towards the surface of the glacier under *compressive flow*. The englacial transport of *subglacial debris* away from the bed, coupled with the melting of glacier ice via insolation, serves to increase the quantity of supraglacial debris present in the ablation zone (see fig. 5.4).

Glacier Deposition

Rock debris being transported by glaciers will eventually be deposited beneath and adjacent to the glaciers via one of two processes, *ablation* or *lodgement*. Ablation (melt out) involves the melting of glacier ice that releases supraglacial and subglacial rock debris in transport. The energy required to melt glacier ice is derived from insolation, internal friction, and geothermal heat. Lodgement occurs when traction cannot overcome friction between a rock clast and the underlying bedrock surface beneath thick ice and the clast comes to rest on the glacier bed. Sediment deposited directly from glacier ice via melt out or lodgement is referred to as *glacial till* or *diamicton*, and the landforms consisting of these deposits are known as *moraines*.

Alpine Glacial Landscapes

Erosional Landforms

A variety of landforms develop largely in response to the abrasion and plucking operating beneath glaciers. These features include striations, roches moutonnées, cirques, glacial troughs, and fiords (fjords).

Striations are small, linear grooves cut into bedrock and oriented parallel to the direction of ice flow that are produced by abrasion (see fig. 5.3). The orientation of these features on bedrock surfaces provides a sense of the direction of ice flow across the landscape.

Roches moutonnées are asymmetric, streamlined bedrock forms produced by the combined action of abrasion and plucking (see fig. 5.5). The gently sloping upstream side of these features bears evidence of intense abrasion in the form of abundant striations and a strongly polished surface. The steeply sloping downstream side bears evidence of plucking in the form of fractured rock and rugged topography. Like striations, these features are oriented parallel to the direction of ice flow and provide a sense of ice motion across the landscape.

Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5. 5 A roche moutonnée developed in crystalline bedrock. The shape of this erosional landform indicates that the direction of glacier flow was from right to left in this photograph

Cirques are amphitheater-shaped depressions excavated in bedrock that extend from a steeply sloping headwall to a low rim (see fig. 5.6). The combined action of glacial abrasion and plucking and the frost shattering of rock produce these erosional features.

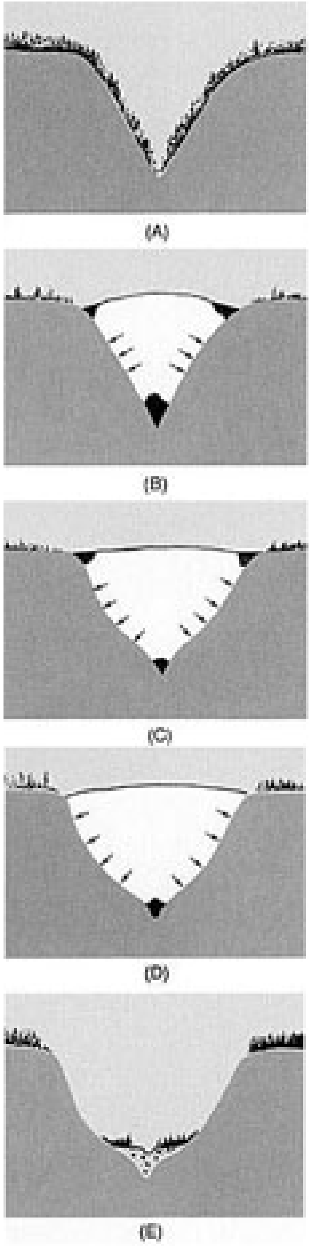
Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5. 6 A large cirque basin developed in sedimentary bedrock. Arêtes mark the margin of the cirque basin at the skyline in this photograph. A small cirque glacier partially occupies the cirque basin

The increase in insolation during the late spring and summer warms snow and ice at the surface of glaciers, leading to the generation of meltwater. Some of this meltwater will infiltrate the snow and percolate through glacier ice and into the underlying bedrock. Here, the meltwater freezes and expands, thus contributing to the frost shattering of rock.

Erosion of cirque floors occurs largely by glacial abrasion. Supraglacial debris and frost-shattered rock entrained by traction at the base of glaciers provides the cutting tools required to abrade the bedrock surface. The rotary motion of cirque glaciers associated with extending flow beneath the accumulation zone and compressive flow beneath the ablation zone actively abrade the bedrock substrate and deepen the cirque floor. Cirque growth may progress over time to the point where narrow ridges of rock known as *arêtes* separate adjacent cirques (see fig. 5.6). Where three or more arêtes converge, a sharply pointed mountain peak known as a *horn* develops.

Glacial troughs develop where the flow of glacier ice is confined to mountain valleys. Thick ice over the valley floor flows more vigorously than thin ice near the glacier margin that overlies the valley walls (i.e., think about Glen's Flow Law). Glacial erosion is more effective beneath the centre of an alpine glacier than near the ice margins. The end result is to produce a deep, U-shaped trough as illustrated in figure 5.7. Where alpine glaciers extend to the coastline, glacial troughs may be eroded several hundreds of metres below sea level. Deglaciation of the coastline allows the sea to flood into and partially drown the glacial trough to create a *fiord* (see fig. 5.8).



Source: Ritter et al. (1998)

Fig. 5.7 A conceptual model for the development of glacial troughs

Source: Natural Resources Canada. Canadian Landscapes [online], http://sts.gsc.nrcan.gc.ca/clf/landscapes_details.asp?numero=266

Fig. 5. 8 Photograph of a fiord developed in crystalline bedrock in the Canadian Arctic

Depositional Landforms

Glacial tills are among the most common surficial materials represented in present-day and formerly glaciated landscapes. Moraines are depositional landforms composed of glacial till.

Lateral moraines are composed largely of coarse, angular rock debris delivered to the glacier margin or surface via mass wasting. Till is deposited largely by ablation of the glacier surface to form ridges along each side of a glacier (see fig. 5.9). If adjacent alpine glaciers coalesce, their lateral moraines join to form a medial moraine.

Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5.9 A prominent lateral moraine, composed of supraglacial debris, developed against the margin of an outlet valley glacier

End moraines are composed of rock debris transported supraglacially and englacially and deposited in distinct ridges at the margin of glaciers (see fig. 5.10). Till is deposited by supraglacial and subglacial melt out. In the latter case, squeezing of water-soaked till from beneath the glacier margin contributes to the formation of end moraines.

Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5.10 A prominent end moraine, composed of supraglacial and subglacial tills, developing at the snout of an outlet valley glacier

In contrast to the till that composes lateral moraines, rock clasts present in the till that composes end moraines exhibit a greater degree of roundness—the product of abrasion and chemical weathering in a subglacial environment. End moraines mark the maximum extent of glaciers in present-day and formerly glaciated landscapes.

Hummocky moraine develops where supraglacial debris is deposited over stagnant ice at the glacier margin. Stagnant ice refers to glacier ice that has become so thin via ablation that it is no longer capable of flowing. Slow and uneven melt out of the buried stagnant ice creates a landscape characterized by rolling topography associated with till ridges interspersed with shallow, water-filled basins known as *kettle lakes* (see fig. 5.11).

Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5.11 Hummocky moraine and kettle lakes

Ground moraine is composed of poorly sorted glacial till deposited beneath the glacier by a combination of lodgement and basal melting of the glacier. The till is deposited in sheets of variable thickness over broad areas of present-day and formerly glaciated landscapes.

Glaciofluvial Processes and Sediments

Meltwater is discharged across and beneath glaciers and enters stream channels in the *proglacial environment* beyond the ice margin (see Module 3). During periods of high meltwater production and/or direct precipitation, subglacial and proglacial stream discharge increases. Stream power increases, contributing to channel erosion in both the subglacial and proglacial environments and increases in stream load. Stream load refers to the quantity of sediment that a stream is capable of transporting at a given discharge. Rock debris of all sizes from boulders to rock flour is entrained and transported by streams during these periods of high discharge.

During periods of low meltwater production and/or the absence of direct precipitation, subglacial and proglacial stream discharge decreases. Stream power decreases, contributing to channel aggradation in both the subglacial and proglacial environments as the stream load is deposited within the channel. In subglacial environments deposition of stratified sand and gravel within stream channels forms *eskers* (see fig. 5.12). In proglacial environments deposition of stratified sand and gravel within stream channels forms *bars* within *braided stream channels* (see fig. 5.13).

Source: Compics International Inc. (1997, *Glaciers and Glaciation*)

Fig. 5.12 An esker traverses a tundra landscape in northern Canada. Note the coarse texture of the glaciofluvial sediments that compose the esker exposed in the foreground of this photograph.

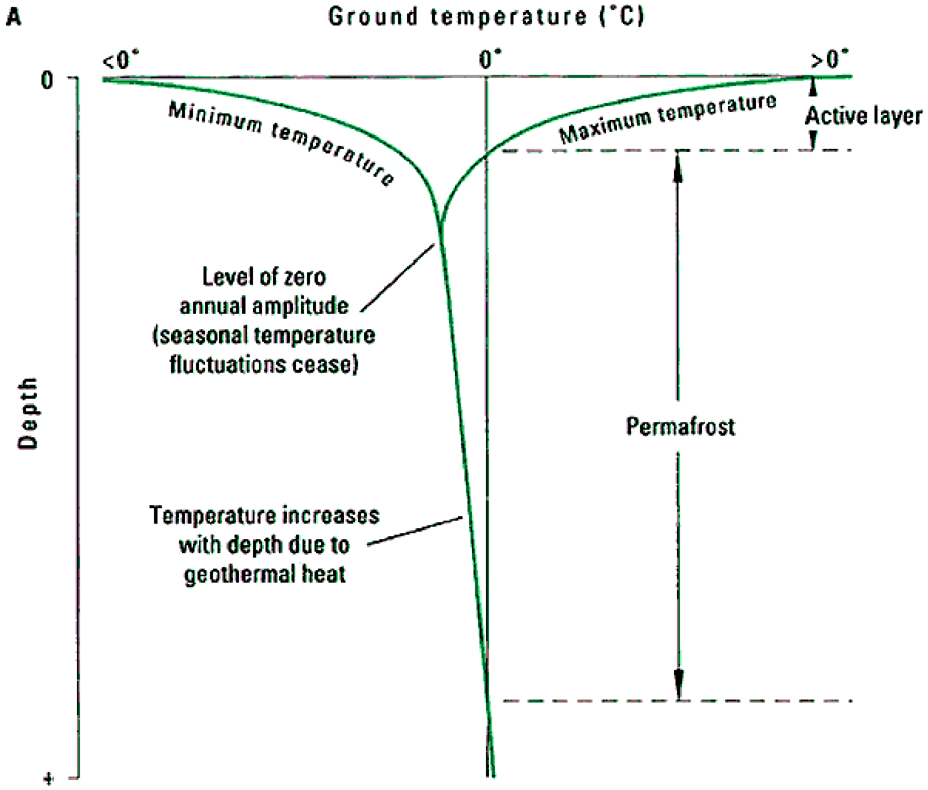
Fig. 5.13 A large northern stream illustrating the development of sandbars that create a braided channel pattern

Permafrost and Periglacial Landscapes

Permafrost

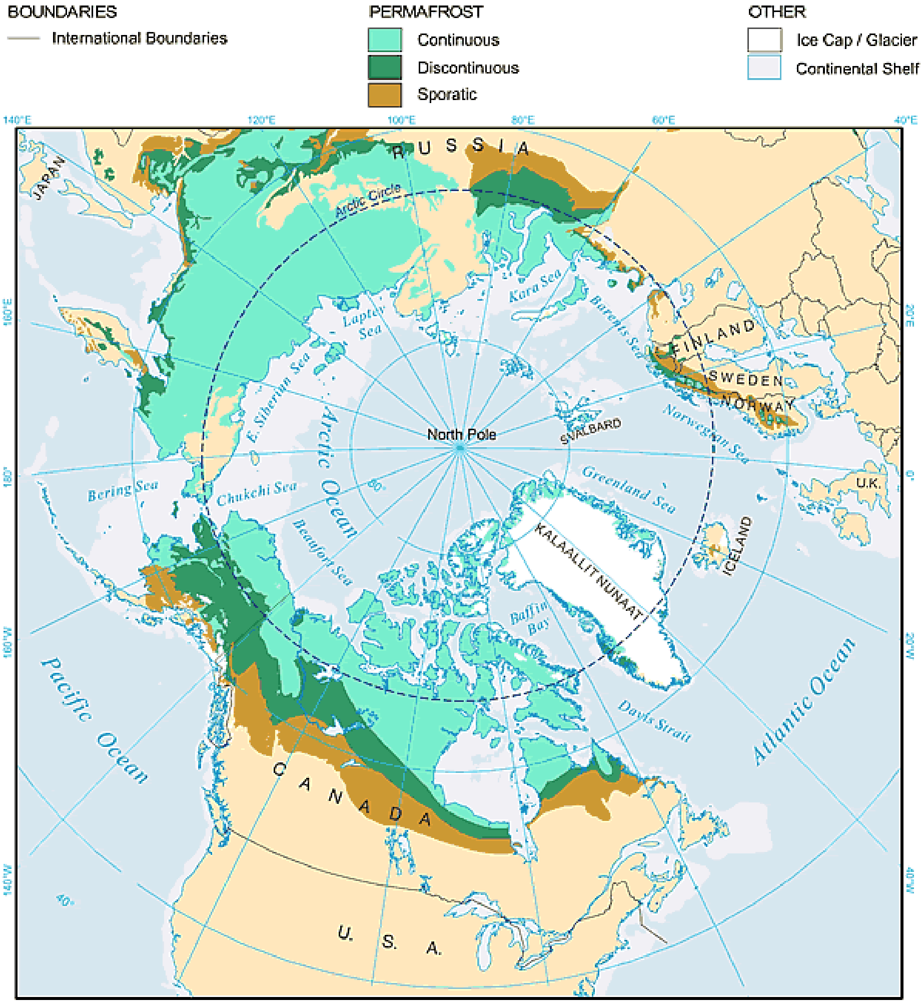
In regions where the depth of frost penetration into the ground during the fall and winter is greater than the depth of ground thawing in the summer, a zone of permanently frozen ground, known as *permafrost*, persists throughout the year. (See fig. 5.14.) Permafrost refers to a thermal condition observed in soils, peat, and rocks in which ground temperatures remain below 0°C for two or more consecutive years. The upper surface of the permanently frozen ground is known as the *permafrost table*. The ground above the permafrost table is referred to as the *active layer*; this layer thaws in the summer. Seasonal variations in ground temperatures decrease with depth towards the *level of zero amplitude* that defines the permafrost table. Ground temperatures do not fluctuate at greater depths, but they gradually increase with depth (i.e., 1°C per 50 metres) in response to the flow of geothermal heat towards the ground surface, reaching the freezing point at the base of the permafrost layer (see fig. 5.14). *Taliks* are areas of unfrozen ground that are situated within and/or below the base of the permafrost layer. (We examined the influence of taliks on the movement of groundwater within permafrost in Module 3.) Geographers recognize several types of permafrost:

- Continuous permafrost refers to an environment where more than 80% of the ground surface is underlain by permafrost. The southern limit of continuous permafrost corresponds closely to the –8°C mean annual isotherm (derived from air temperatures).
- Discontinuous permafrost refers to an environment where 30–80% of the ground surface is underlain by permafrost. The southern limit of discontinuous permafrost corresponds closely to the –1°C mean annual isotherm (derived from air temperatures).
- Sporadic permafrost refers to an environment where less than 30% of the ground surface is underlain by permafrost.



Source: Adapted from French (1996)

Fig. 5.14 Annual variation in ground temperatures within permafrost



Map Source: Base Map; Atlas of Canada (www.atlas.gc.ca), Government of Canada. Data Source: UNEP/GRID-Arendal. Digital Version: Circum-arctic Map of Permafrost and Ground-Ice Conditions, ver. 1.0. In: International Permafrost Association, Data and Information Working Group, comp. 1998. Circumpolar Active-Layer Permafrost System (CAPS), version 1.0. CD-ROM available from National Snow and Ice Data Center, nsidc@kryos.colorado.edu. Boulder, Colorado: NSIDC, University of Colorado at Boulder. Map produced by: GIServices, University of Saskatchewan 2003. Projection: Azimuthal Equidistant. Latitude of origin: 75° N. Central meridian: 90° W. All latitudes north of equator.

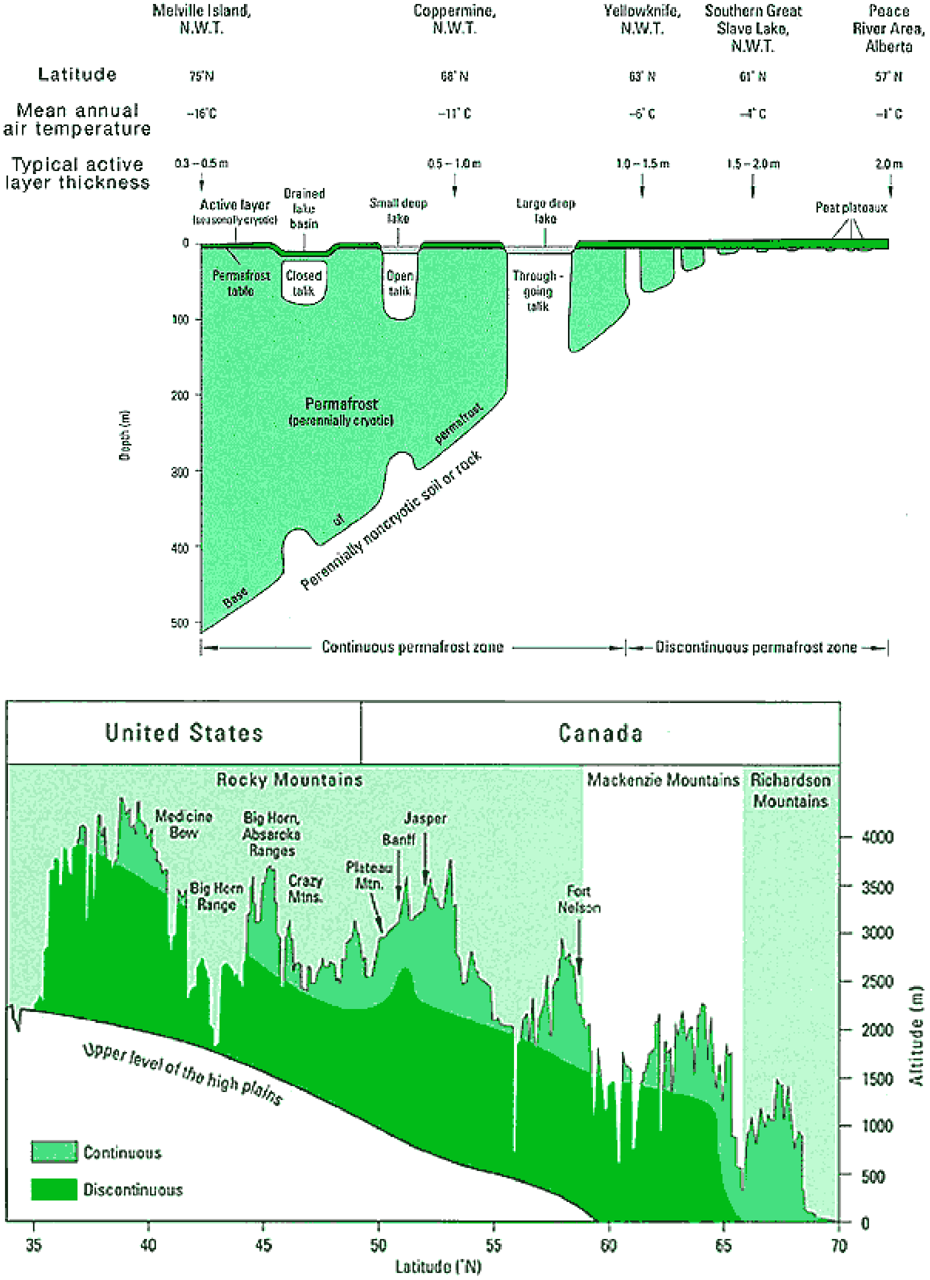
Fig. 5.15 Permafrost zones in the circumpolar North

Permafrost occurs extensively within the northern hemisphere (see table 5.3). The boundaries of the various permafrost zones in the circumpolar North are illustrated in figure 5.15. It is apparent that the boundaries of the permafrost zones dip southwards moving from west to east across Canada. This phenomenon reflects the presence of Hudson Bay; a large inland sea situated squarely in the centre of the Canadian land mass. The surface of Hudson Bay remains covered by sea ice well into July and by the time the cover of sea ice is completely removed in September it is too late for the surface waters to be warmed by insolation prior to fall freeze-up. The cool surface waters of Hudson Bay keep summer air temperatures cool as well and promote the development of permafrost in the regions adjacent to the bay. Students are encouraged to view the climographs for Kuujuuarapik and Churchill in Module 2 that demonstrate the cooling influence of Hudson Bay at these coastal locations.

Table 5.3 The relative extent of permafrost in the northern hemisphere (after French 1996)

	Area (million km^2)	**Percentage Underlain by Permafrost (%)**
Major Permafrost Regions		
Russian Federation	16.84	50
Canada	9.38	50
China	9.38	22
Greenland	2.18	100
United States of America (Alaska)	1.52	82
Northern Hemisphere		
Continuous permafrost	7.6	
Discontinuous permafrost	17.3	
Alpine permafrost	2.3	

The depth of the active layer and the permafrost layer varies with latitude and altitude (see fig. 5.16). Depths depend largely upon the intensity of the cold, the thermal and physical properties of the soil and rock, and the nature of the overlying vegetation cover. The active layer extends to depths of approximately 20 cm in the Canadian Arctic Archipelago and as much as 15 m in the Western Cordillera. Permafrost extends to depths of 50–100 m at the southern boundary of the continuous permafrost zone and to 500–600 m in the Canadian Arctic Archipelago.



Source: Adapted from Briggs et al. (1993)

Fig. 5.16 Latitudinal variations in the extent of permafrost in North America

The seasonal cycle of ground freezing and thawing plays an important role in shaping northern landscapes. The rate of spring thawing influences the magnitude of the nival freshet (see Module 3), and the rate of fall freeze-up influences frost heaving and ice segregation within soils. Most permafrost is impervious to the percolation of water; hence, there is little or no subsoil drainage of water from the active layer. Given this situation, the active layer consists of materials saturated with water and exhibiting low *shear strength*. These materials are susceptible to mass movement in the form of *frost creep* and *gelifluction*. These processes are examined in more detail further in this lecture.

Periglacial Processes and Landforms

Frost Action

Frost action refers to processes associated with the freezing and thawing of soil and rocks. These processes include:

- an increase in the volume of water upon freezing that contributes to the *frost shattering* of rocks
- the formation of *segregated ice* that contributes to the *frost heaving* of the ground surface
- thermal contraction of the ground, leading to the development of *frost cracks* and *wedge ice*

Frost Shattering

The expansion of water upon freezing can generate high pressures capable of exceeding the tensile strength of rock and resulting in the shattering of rocks. Effective frost action can only occur in environments with a plentiful supply of water required to saturate the rocks and frequent freeze-thaw cycles. Given these criteria, it stands to reason that frost action will operate with greater efficacy in Subarctic and alpine environments compared to High Arctic environments. This combination of environmental conditions is rarely encountered under natural field conditions; hence, the efficacy of this process in northern environments is presently being questioned.

Source: Natural Resources Canada. Canadian Landscapes [online], http://sts.gsc.nrcan.gc.ca/clf/landscapes_details.asp?numero=239

Fig. 5.17 Tors and adjacent block field developed by frost shattering of sedimentary bedrock in the Canadian Arctic

Fine-grained sedimentary rocks, such as shale and limestone, and fine-grained metamorphic rocks, such as slate and phyllite, are particularly susceptible to frost action. On upland surfaces, frost action produces accumulations of angular rock clasts surrounding pitted, irregular bedrock outcrops referred to as *tors* (see fig. 5.17). The debris-mantled surface adjacent to tors is referred to as a *block field* or *felsenmeer* (see fig. 5.17). Prolonged exposure of felsenmeer to frost action produces a material known as *grus* consisting of mineral grains released from the surfaces of rock clasts. Frost action operating along the walls of alpine valleys supplies rock clasts to the surface of alpine glaciers where they are entrained as supraglacial debris.

Ground Ice

Ground ice refers to all types of ice formed within freezing and frozen ground. Landforms characterized by a variety of geometric shapes, including circles, polygons, and stripes, are associated with the presence of ground ice in periglacial landscapes. These features are referred to collectively as *patterned ground*.

Segregated Ice, Frost Heave, and Non-sorted Circles

Freezing of groundwater within pores, fractures, and joints in soil and rock contributes to the formation of segregated ice crystals. As freezing progresses, segregated ice crystals grow and coalesce to form ice lenses. The growth of these ice lenses is sustained by the continuous movement of groundwater towards the freezing plane within soil or rock until the moisture supply is depleted. The growth of segregated ice lenses displaces the soil surface upwards—a process referred to as *frost heaving*.

Source: Compics International Inc. (1997, *Canadian Arctic*)

Fig. 5.18 Non-sorted circles developed in a poorly sorted mixture of weathered sedimentary bedrock and fine-grained sediments

Circular patches of bare, fine-grained soils that are bordered by tundra vegetation are common features observed in periglacial landscapes (see fig. 5.18). These features are referred to by a variety of terms including non-sorted circles, frost boils, mud boils, and earth circles. For our purposes we will use the term *non-sorted circles* when referring to these features. They may occur singly or in groups, and commonly vary from 0.5 to 3.0 metres in diameter. The origin of these periglacial landforms remains unclear, although numerous theories have been proposed to account for their development. We will examine two hypotheses for the development of non-sorted circles.

- cryostatic pressure hypothesis

  According to this hypothesis, freezing of the active layer occurs in two directions: upward from the permafrost table, and downward from the ground surface. As freezing progresses, segregated ice lenses form within the soil, trapping bodies of saturated unfrozen sediments above the permafrost table. The continued growth of segregated ice exerts pressure (i.e., cryostatic pressure) on the water present within the pores of the unfrozen sediment. This pressure is relieved by the upward flow of the saturated sediments towards the ground surface (see fig. 5.19). This hypothesis has been challenged on the basis that cryostatic pressures measured in naturally occurring soils are not sufficient to overcome the strength of the overlying frozen sediment. Given this situation, it is unlikely that saturated fine-grained sediments can be injected through freezing sediments to reach the ground surface.

- soil convection hypothesis

  According to this hypothesis, freezing of the active layer occurs in two directions: upward from the permafrost table, and downward from the ground surface. As freezing progresses, segregated ice lenses form within the soil, drawing water out of the middle of the active layer towards freezing planes at the top and bottom of the active layer. The growth of segregated ice within the active layer results in frost heaving of the ground surface (see fig. 5.20); vertical displacements up to 20 cm can be achieved in this manner. Subsequent thawing of segregated ice within the active layer results in subsidence of the ground surface. At the same time, pore water pressures within the thawing sediments increase as the melting of ice lenses generates meltwater. The increase in pore pressure reduces the strength of the unfrozen sediments in the middle of the active layer. The pressure exerted by the weight of the overlying sediments results in the upward displacement of saturated sediments towards the ground surface (see fig. 5.20). Recent field experiments support the occurrence of soil convection within the active layer and its role in producing non-sorted circles.

The mixing of soil that accompanies the seasonal freezing and thawing of the active layer is referred to as *cryoturbation*. The depth of mixing affects the rooting zone for vegetation and nutrient cycling within soils in periglacial landscapes.

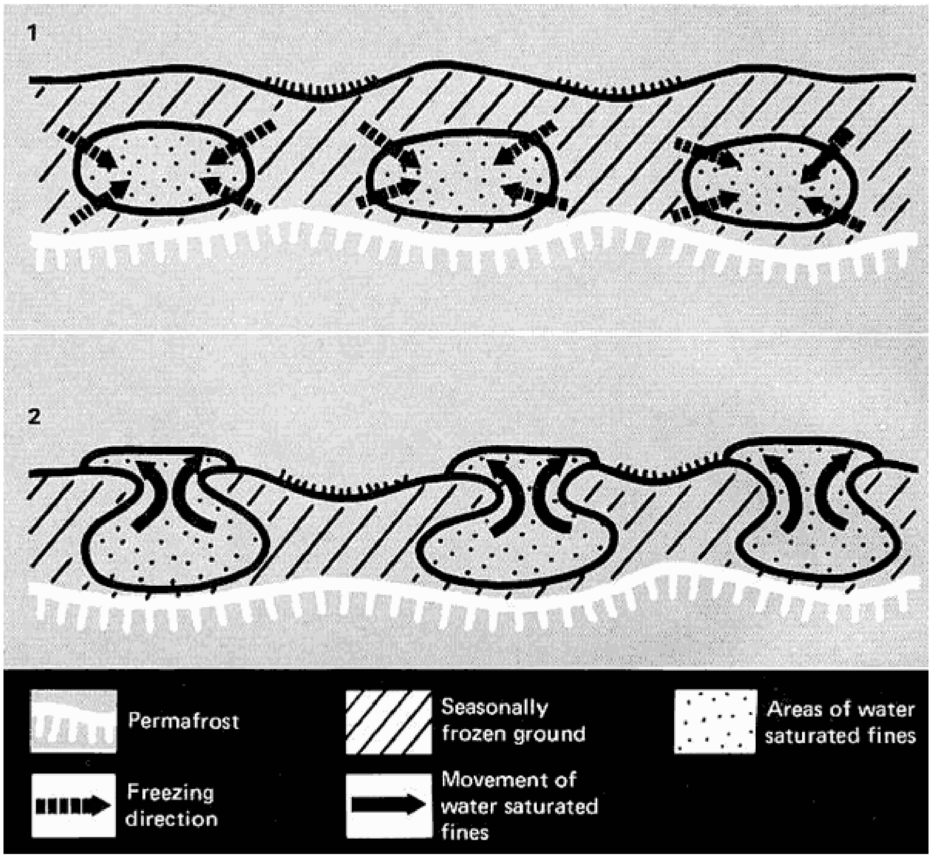
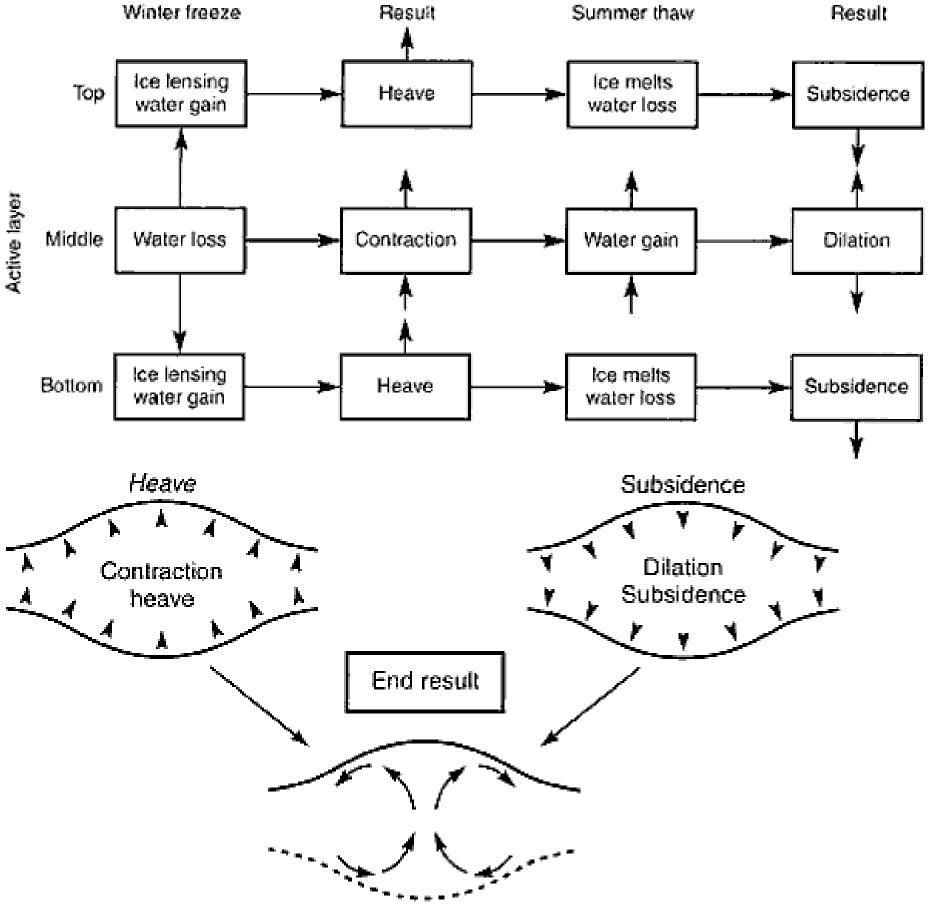


Fig. 5.19 A conceptual model of the cryostatic pressure hypothesis for the origin of non-sorted circles (French 1996)



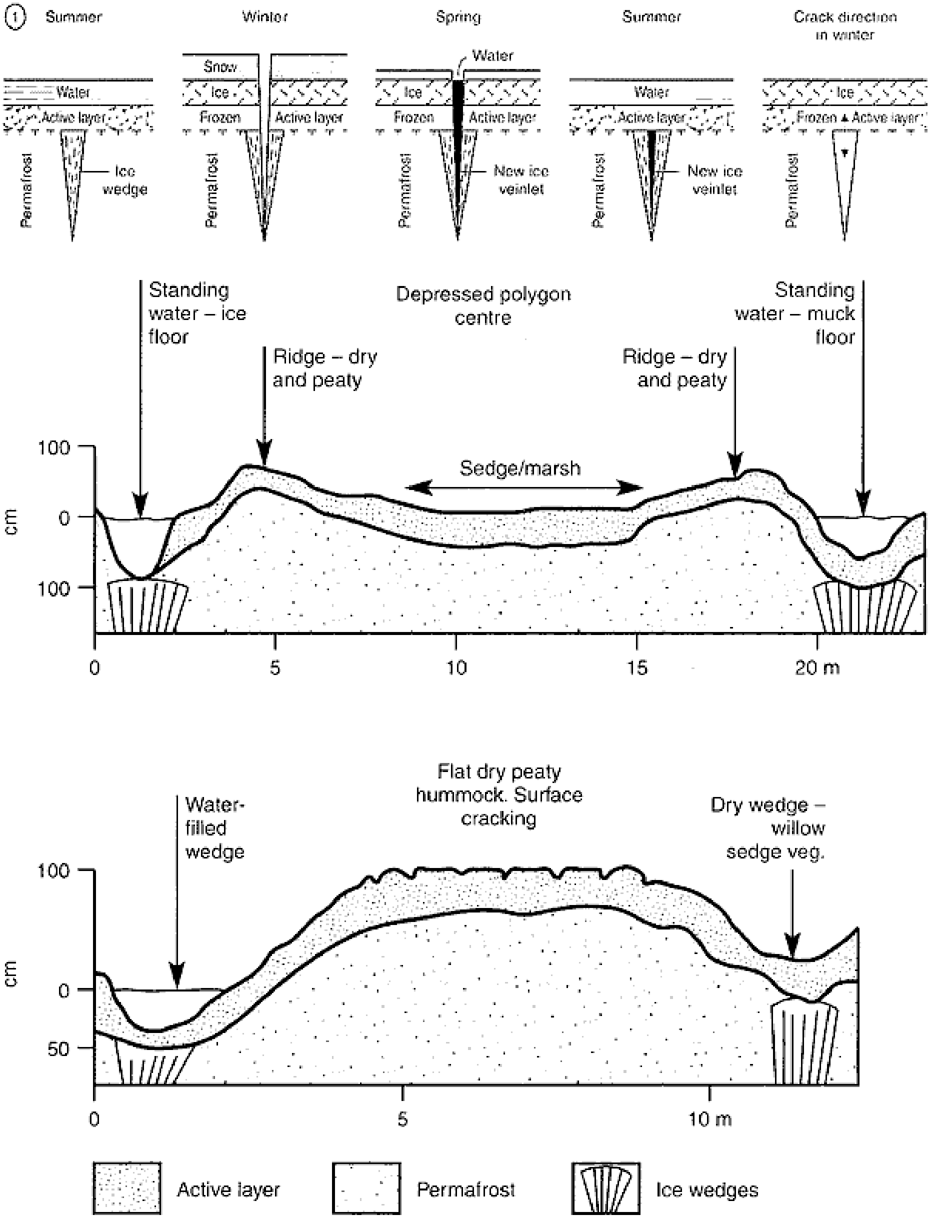
Source: Adapted from French (1996)

Fig. 5.20 A conceptual model of the soil convection hypothesis for the origin of non-sorted circles

Frost Cracking, Wedge Ice, and Polygons

Ice wedge polygons are common landforms observed in periglacial landscapes. The conditions favourable for the development of these landforms occur in poorly drained lowlands within the continuous permafrost zone. The sequence of events leading to the development of ice wedge polygons is initiated by the prolonged freezing of ice-rich soil at temperatures below –15°C during the winter. In response to this intense freezing, the ground contracts and cracks: a process known as *frost cracking*. The cracks outline the borders of polygons visible on the ground surface.

In the following spring and early summer, meltwater infills the cracks and freezes, creating narrow, wedge-shaped veins of ice that penetrate to depths below the active layer (see fig. 5.21). As the active layer develops over the summer, the soil warms and expands. However, soil expansion is constrained by the presence of the ice wedges so that the soil is forced to expand upward, forming a slightly raised rim of soil adjacent to the ice-filled frost cracks.



Source: Adapted from French (1996)

Fig. 5.21 A conceptual model for the origins of ice wedge polygons

The ice wedges present within the frost cracks exhibit a lower tensile strength than the surrounding ice-rich soil; hence, they form lines of weakness within the ground. In the next winter, frost cracking occurs preferentially within the ice wedges and the cycle is repeated year after year, progressively enlarging the ice wedges so that they become wider and deeper (see fig. 5.21). As the ice wedges enlarge, the soil adjacent to the frost cracks continues to be thrust upward to form distinct ridges. The end result is a polygonal landform characterized by a marginal trough occupied by ice wedges and a raised rim of soil and vegetation surrounding a central depression; this landform is referred to as a *low-centred ice wedge polygon* (see fig. 5.22). The dimensions of these landforms vary considerably from place to place. The width of the polygons varies from 15 to 40 metres. The width and depth of ice wedges range from 1 to 4 metres and 3 to 10 metres, respectively. The raised rim may stand 0.5 to 1.5 metres above the central depression. Differences in the topography between the marginal rim and central depression cause noticeable differences in drainage and the composition of plant communities that inhabit the surface of these polygons. The central depression is often waterlogged in summer and inhabited by wetland vegetation dominated by mosses and sedges (*Carex* spp.). The rim exhibits better drainage and is inhabited by a variety of plants better suited to drier soils; these plants include dwarf willow (*Salix* spp.), dwarf birch (*Betula* spp.), crowberry (*Empetrum nigrum*), bearberry (*Arctostaphylos uva-ursi*), bilberry (*Vaccinium uliginosum*), grasses, and lichens.

Source: Compics International Inc. (1997, *Canadian Arctic*)
Fig. 5.22 Low-centred ice wedge polygons in a tundra landscape

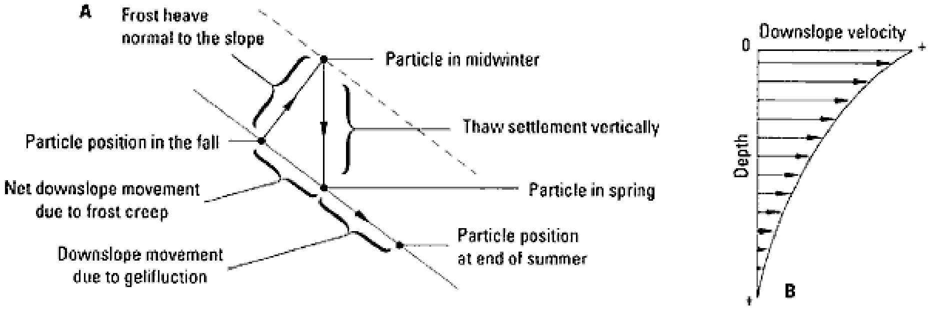
Over time, low-centred ice wedge polygons are transformed into high-centred ice wedge polygons. Several processes operating simultaneously bring about this transformation. Continued enlargement of ice wedges at the margins of these landforms gradually raise the ground surface in the interior of the polygons. At the same time, the gradual accumulation of plant debris and *aeolian* sediments also raise the ground surface within the central depression above the surrounding terrain. As the ground surface rises, drainage improves and the soil becomes progressively drier. Plants inhabiting the rim of the polygon expand and displace the wetland vegetation at the centre of the polygons. Water passing over this landscape is progressively diverted into marginal troughs. The flow of water within the troughs thaws the upper surface of ice wedges and erodes the margins of the polygons, resulting in the progressive widening and deepening of the troughs. The end result is a dome-shaped polygon referred to as a high-centred ice wedge polygon (see fig. 5.23).

Source: Compics International Inc. (1997, *Canadian Arctic*)

Fig. 5.23 High-centred ice wedge polygons in a tundra landscape

Mass Movement Processes

The saturation of soils in the active layer contributes to the mass movement of surficial materials. *Solifluction*, the action of slow flow in water-saturated soils, is a ubiquitous process operating on hillslopes in periglacial landscapes. Movement of material occurs via two processes: *frost creep* and *gelifluction* (see fig. 5.24).



Source: Adapted from French (1996).

Fig. 5.24 The origin of solifluction lobes via frost creep and gelifluction

Frost creep refers to the downslope movement of particles in response to the expansion and contraction of surficial materials associated with frost heaving of the ground surface. Materials moving in this manner can be displaced downslope several millimetres to several centimetres each year. Gelifluction refers to the flow of surficial materials associated with the presence of permafrost. Most permafrost is impervious to the percolation of water; hence, there is little or no drainage of water from the active layer. Given this situation, the active layer consists of materials saturated with water and exhibiting low *shear strength*. These materials are susceptible to mass movement under the influence of the force of gravity. The combination of frost creep and gelifluction of surficial materials produces distinct lobate landforms known as *solifluction lobes* (see fig. 5.25).

Source: Compics International Inc. (1997, *Canadian Arctic*)
Fig. 5.25 Active solifluction lobes in a tundra landscape

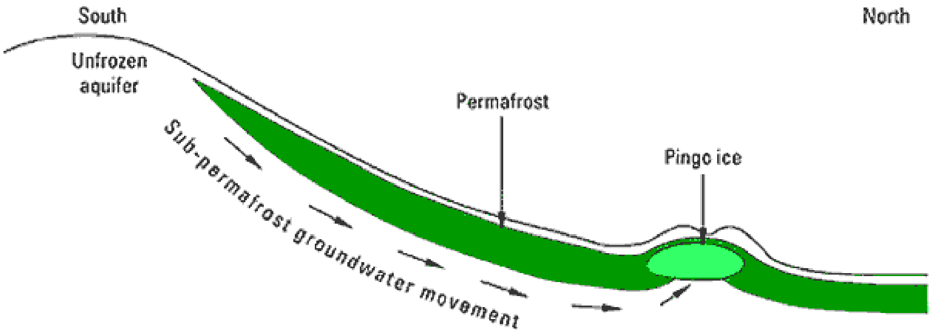
Gelifluction can transform non-sorted circles and ice wedge polygons present on upland surfaces into *stripes* on hillslopes, as mass movement becomes more important on steeper slopes (see fig. 5.26).

Source: Compics International Inc. (1997, *Canadian Arctic*)
Fig. 5.26 Active soil stripes in a tundra landscape

Intrusive Ice and Pingos

Pingos are large ice-cored hills ranging from several metres to almost 50 metres in height and from about 30 to 600 metres in diameter. The ice cores consist largely of *intrusive ice* derived from the injection of water under pressure into sediments beneath the ground surface. Geographers distinguish between open- (hydraulic) and closed- (hydrostatic) system pingos according to the source of the water that forms the ice core.

Open-system pingos have been observed throughout the circumpolar North but are particularly numerous in unglaciated regions of the Yukon and Alaska. These landforms tend to develop as isolated features or in small groups on valley floors adjacent to south-facing valley slopes underlain by discontinuous permafrost. Water can infiltrate into the ground and circulate downslope within taliks. Near the base of the slope, hydraulic (artesian) pressure within taliks can generate sufficient force to drive water upward to the ground surface. Open-system pingos develop where groundwater rises to the surface and freezes to form a body a massive ground ice that heaves the overlying sediments (see fig. 5.27).



Source: Adapted from Briggs et al. (1993)

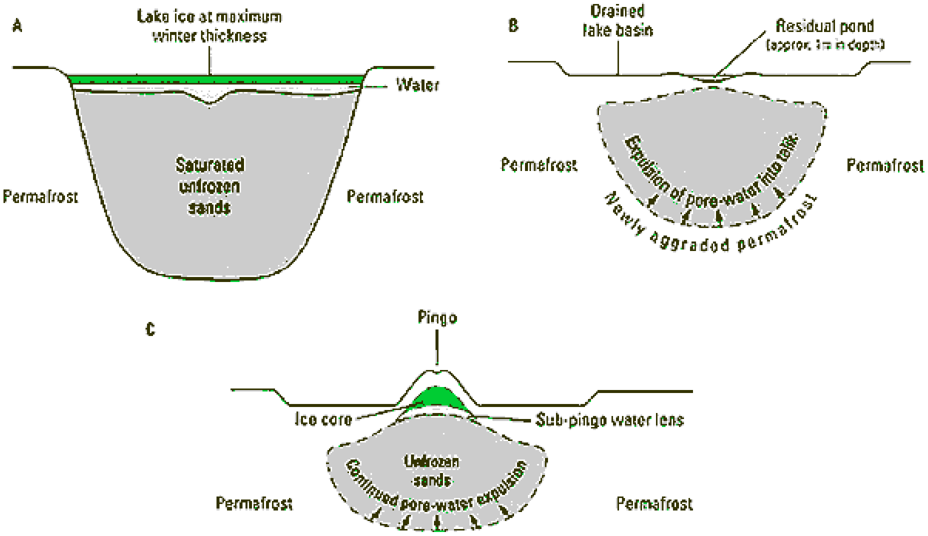
Fig. 5.27 The origin of an open-system pingo via the discharge of sub-permafrost groundwater at the base of a valley slope

Closed-system pingos occur almost exclusively in the zone of continuous permafrost and are commonly associated with alluvial lowlands that exhibit little topographic relief. The largest concentration of closed-system pingos occurs along the coastal plain of the Mackenzie Delta, the Yukon coastal plain, western Victoria Island, and Banks Island. They usually occur singly within small shallow lake basins (see fig. 5.28).

Source: Compics International Inc. (1997, *Canadian Arctic*)

Fig. 5.28 A closed-system pingo developed in a shallow lake basin on the delta plain of the Mackenzie River in Arctic Canada

Closed-system pingos form by the heaving of frozen ground in response to high pore water pressure, the result of pore water expulsion during the aggradation of permafrost, typically in a closed talik beneath a drained lake basin or river channel. The development of closed-system pingos occurs in stages, as illustrated in figure 5.29.



Source: Adapted from Briggs et al. (1993)

Fig. 5.29 The origin of a closed-system pingo following drainage of shallow tundra lake

The initial stage of pingo development depends on the presence of a deep talik surrounded by permafrost (i.e., a closed talik). This talik owes its existence to the overlying water body. The great capacity of water to store heat facilitates the thawing of the ground beneath the lake basin or river channel. Over many years, the heat supplied by the water body can penetrate deeply into the ground to create a saturated talik.

The draining of the lake basin or river channel exposes the ground surface, and the talik begins to refreeze. Freezing proceeds inwards from the permafrost table and downwards from the top of the newly exposed ground surface. As freezing progresses, water in the saturated talik expands and forces liquid water out of the freezing sediments; a process referred to as *pore water expulsion*. As the liquid water rises through the overlying sediments, it freezes to form segregated ice and the ground surface heaves slowly upward. Continued contraction of the talik causes pore water pressures to increase until the remaining liquid water is injected into the overlying sediments faster than it can freeze, producing a large body of intrusive ice that contributes to the rapid growth of the pingo.

Summary

Glaciers cover approximately 10% of the present surface of the Earth and cover an area of approximately 2 million km^2 in the circumpolar North. Glaciers occupy highlands where winter snowfall is heavy and summer temperatures are not adequate to completely melt the snowpack. Over time, and under favourable climatic conditions, these highland glaciers expand to continental proportions to form ice sheets. These moving bodies of ice serve as active geomorphic agents capable of eroding, transporting, and depositing immense quantities of rock debris in polar and alpine landscapes. A suite of landforms, characterized by striations, roches moutonnées, cirques, fiords, and moraines, bear witness to the activity of glaciers in northern landscapes.

Periglacial landscapes are closely linked to regions characterized by low annual temperatures, low annual precipitation, intense frost action, and the presence of permafrost. Frost action and permafrost-related processes are dominant geomorphic processes in these landscapes. Permafrost refers to a thermal condition observed in soils, peat, and rocks in which ground temperatures remain below 0^{o}C for two or more consecutive years. The upper surface of the permanently frozen ground is known as the permafrost table. The ground above the permafrost table is referred to as the active layer; this layer thaws in the summer. The depth of the active layer and the permafrost layer varies with latitude and altitude. Depths depend largely upon the intensity of the cold, the thermal and physical properties of the soil and rock, and the nature of the overlying vegetation cover. A suite of landforms, characterized by tors, solifluction lobes, patterned ground, and pingos, are produced by periglacial processes operating in northern landscapes.

Study Questions

A. Multiple Choice Questions

1. Which of the following terms is used to describe the overall “health” of a glacier?
   a. accumulation
   b. ablation
   c. water balance
   d. mass balance

2. Roches moutonnées form as a result of ____________________.
   a. ice plucking on both the upstream and downstream side of a rock outcrop
   b. abrasion on both the upstream and downstream side of a rock outcrop
   c. abrasion on the upstream side and ice plucking on the downstream side of a rock outcrop
   d. ice plucking on the upstream side and abrasion on the downstream side of a rock outcrop

3. Climate influences the movement of glaciers through its control of ____________________.
   a. the topographic relief of the landscape
   b. the supply of geothermal energy
   c. the mass balance of the glacier
   d. the rate of chemical weathering of bedrock
   e. all of the above

4. ____________________ contributes to the deposition of ____________________ at the ____________________ of alpine glaciers.
   a. Lodgement; glacial till; base
   b. Lodgement; glaciofluvial sediments; front
   c. Lodgement; glaciofluvial sediments; base
   d. Lodgement; glacial till; front
   e. none of the above

5. This glacial landform is composed largely of supraglacial debris: ____________________.
   a. roche moutonnée
   b. lateral moraine
   c. cirque
   d. glacial trough

6. Rapid and large variations in stream discharge associated with snowmelt are responsible for the development of ____________________ in proglacial streams.
   a. sandbars
   b. eskers
   c. lateral moraines
   d. end moraines
   e. none of the above

7. The flow of cold-based glaciers is influenced by____________________.
   a. internal deformation
   b. basal sliding
   c. soft sediment deformation
   d. a. and b.
   e. all of the above

8. Mass wasting of Earth materials contributes significantly to the development of this periglacial landscape feature: ____________________.
   a. open-system pingo
   b. lateral moraine
   c. solifluction lobes
   d. sorted stone circles

9. The term “talik” describes ____________________.
   a. a body of massive ground ice within the interior of pingos
   b. a large rock basin produced by glacial erosion
   c. a body of unfrozen ground within permafrost
   d. an accumulation of coarse rock debris on mountain slopes

10. These landforms are observed uniquely in the discontinuous permafrost zone: ____________________.
    a. closed-system pingos
    b. braided stream channels
    c. ice wedge polygons
    d. open-system pingos

11. ____________________ is associated with the presence of ____________________ ice in periglacial environments.
    a. Cryoturbation; wedge
    b. Gelifluction; wedge
    c. Cryoturbation; segregated
    d. Gelifluction; intrusive

12. These landforms are observed uniquely in the continuous permafrost zone: ____________________.
    a. sorted stone circles
    b. ice wedge polygons
    c. braided stream channels
    d. open-system pingos

13. ____________________ pressure within taliks in permafrost contributes to the development of ____________________ ice and ____________________.
    a. Artesian; wedge; ice wedge polygons
    b. Hydrostatic; intrusive; open-system pingos
    c. Artesian; intrusive; closed-system pingos
    d. Hydrostatic; intrusive; closed-system pingos
    e. Artesian; segregated; sorted stone circles

14. The process of ____________________ contributes to the development of ____________________.
    a. gelifluction; solifluction lobes
    b. frost cracking; felsenmeer
    c. frost shattering; ice wedge polygons
    d. gelifluction; non-sorted circles

Answers to Multiple Choice Questions

1. d
2. c
3. c
4. a
5. b
6. a
7. a
8. c

9. c

10. d

11. c

12. b

13. d

14. a

B. Essay Questions

1. Describe the movement of rock debris, sediments, and water in glaciated alpine landscapes in the discontinuous permafrost zone of northern Canada. Your answer should include brief descriptions of characteristic landforms associated with the transfer of sediment and water through these alpine landscapes.

2. How do the following environments affect the development of periglacial landforms?
    a. well-drained, fine-grained parent materials in the discontinuous permafrost zone
    b. poorly drained, fine-grained parent materials in the continuous permafrost zone

Glossary of Terms

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