Chapter 16 Glaciation

Learning Objectives

After reading this chapter, completing the exercises within it, and answering the questions at the end, you should be able to:

- Describe the timing and extent of Earth's past glaciations, going as far back as the early Proterozoic.
- Describe the important geological events that led up to the Pleistocene glaciations and how the Milankovitch orbital variations along with positive feedback mechanisms have controlled the timing of those glaciations.
- Explain the differences between continental and alpine glaciation.
- Summarize how snow and ice accumulate above the equilibrium line and are converted to ice.
- Explain how basal sliding and internal flow facilitate the movement of ice from the upper part to the lower part of a glacier.
- Describe and identify the various landforms related to alpine glacial erosion, including U-shaped valleys, arêtes, cols, horns, hanging valleys, truncated spurs, drumlins, roches moutonées, glacial grooves, and striae.
- Identify various types of glacial lakes, including tarns, finger lakes, moraine lakes, and kettle lakes.
- Describe the nature and origins of lodgement till, ablation till, and glaciofluvial, glaciolacustrine, and glaciomarine sediments.

Figure 16.0.1 Glaciers in the Alberta Rockies: Athabasca Glacier (centre left), Dome Glacier (right), and the Columbia Icefield (visible above both glaciers). The Athabasca Glacier has prominent lateral moraines on both sides.

A **glacier** is a long-lasting body of ice (decades or more) that is large enough (at least tens of metres thick and at least hundreds of metres in extent) to move under its own weight. About 10% of Earth's land surface is currently covered with glacial ice, and although the vast majority of that is in Antarctica and Greenland, there are many glaciers in Canada, especially in the mountainous parts of B.C., Alberta, and Yukon and in the far north (Figure 16.0.1). At various times during the past million years, glacial ice has been much more extensive, covering at least 30% of the land surface at times.

Glaciers represent the largest repository of fresh water on Earth (~69% of all fresh water), and they are highly sensitive to changes in climate. In the current warming climate, glaciers are melting rapidly worldwide, and although some of the larger glacial masses will last for centuries more, many smaller glaciers, including many in western Canada, will be gone within decades, and in some cases, within years. That is much more than just a troubling thought and a scenic loss for western Canadians because we rely on glacial ice for our water supplies—if not for water to drink, then for water to grow food. Irrigation systems in B.C. and across Alberta and Saskatchewan are replenished by meltwater originating from glaciers in the Coast Range and the Rocky Mountains.

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16.1 Glacial Periods in Earth's History

We are currently in the middle of a **glacial period** (although it's less intense now than it was 20,000 years ago) but this is not the only period of glaciation in Earth's history; there have been many in the distant past, as illustrated in Figure 16.1.1. In general, however, Earth has been warm enough to be ice-free for much more of the time than it has been cold enough to be glaciated.

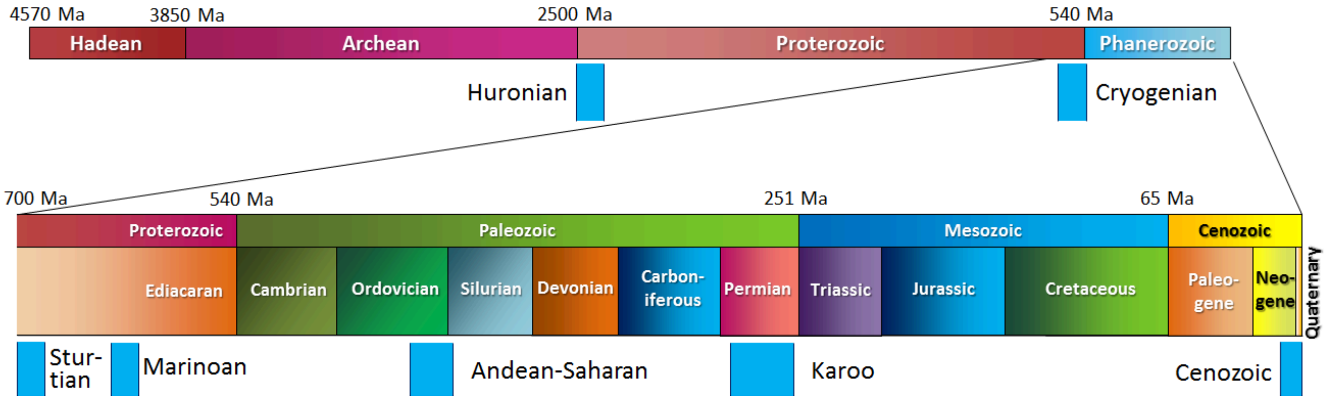

Figure 16.1.1 The record of major past glaciations during Earth's history.

The oldest known glacial period is the Huronian. Based on evidence of glacial deposits from the area around Lake Huron in Ontario and elsewhere, it is evident that the Huronian Glaciation lasted from approximately 2,400 to 2,100 Ma. Because rocks of that age are rare, we don't know much about the duration, intensity or global extent of this glaciation.

Late in the Proterozoic, for reasons that are not fully understood, the climate cooled dramatically and Earth was seized by what appears to be its most intense glaciation. The glaciations of the Cryogenian Period (*cryo* is Latin for icy cold) are also known as the "Snowball Earth" glaciations, because it is hypothesized that the entire planet was frozen—even in equatorial regions—with ice up to 1 km thick on the oceans. A visitor to our planet at that time might not have held out much hope for its inhabitability, although life still survived in the oceans. There were two main glacial periods within the Cryogenian, each lasting for about 20 million years: the Sturtian at around 700 Ma and the Marinoan at 650 Ma. There is also evidence of some shorter glaciations both before and after these. The end of the Cryogenian glaciations coincides with the evolution of relatively large and complex life forms on Earth. This started during the Ediacaran Period, and then continued with the so-called explosion of life forms in the Cambrian. Some geologists think that the changing environmental conditions of the Cryogenian are what actually triggered the evolution of large and complex life.

There have been three major glaciations during the Phanerozoic (the past 540 million years), including the Andean/Saharan (recorded in rocks of South America and Africa), the Karoo (named for rocks in southern Africa), and the Cenozoic glaciations. The Karoo was the longest of the Phanerozoic glaciations, persisting for much of the time that the supercontinent Gondwana was situated over the South Pole (~360 to 260 Ma). It covered large parts of Africa, South America, Australia, and Antarctica (see Figure 10.1.3). As you might recall from Chapter 10, this widespread glaciation, across continents that are now far apart, was an important component of Alfred Wegener's evidence for continental drift. Unlike the Cryogenian glaciations, the Andean/Saharan, Karoo, and Cenozoic glaciations only affected parts of the Earth. During Karoo times, for example, what is now North America was near the equator and remained unglaciated.

Earth was warm and essentially unglaciated throughout the Mesozoic. Although there may have been some alpine glaciation at this time, there is no longer any record of it. The dinosaurs, which dominated terrestrial habitats during the Mesozoic, did not have to endure icy conditions.

A warm climate persisted into the Cenozoic; in fact there is evidence that the Paleocene (~50 to 60 Ma) was the warmest part of the Phanerozoic since the Cambrian (Figure 16.1.2). A number of tectonic events during the Cenozoic contributed to persistent and significant planetary cooling since 50 Ma. For example, the collision of India with Asia and the formation of the Himalayan range and the Tibetan Plateau resulted in a dramatic increase in the rate of weathering and erosion. Higher than normal rates of weathering of rocks with silicate minerals, especially feldspar, consumes carbon dioxide from the atmosphere and therefore reduces the greenhouse effect, resulting in long-term cooling.

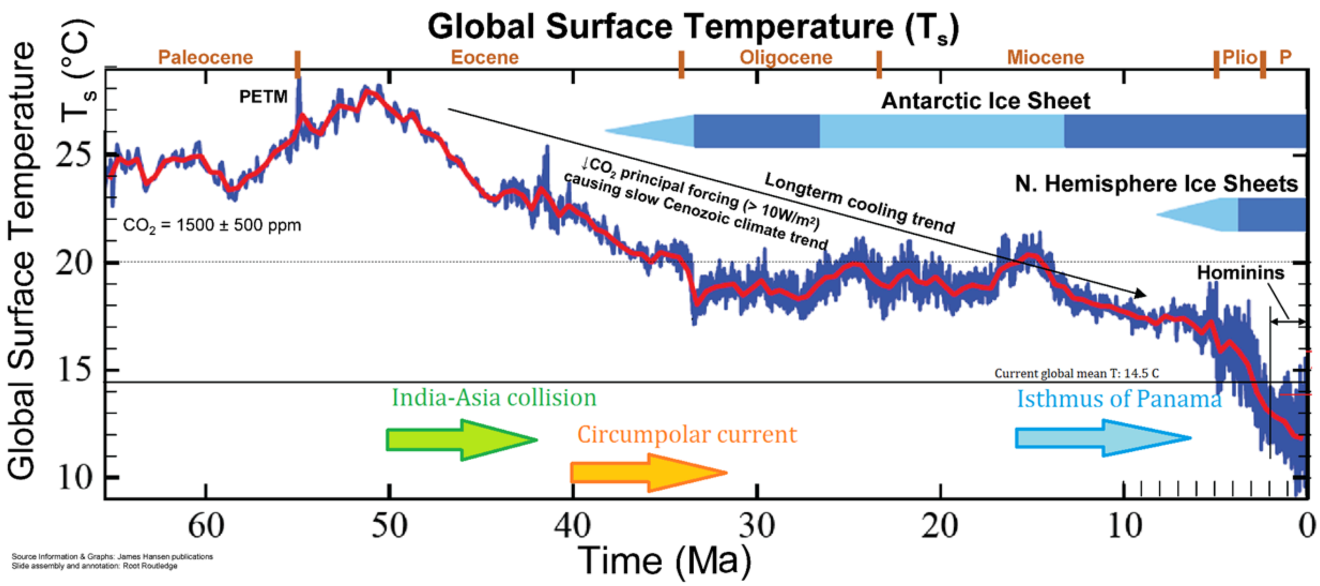

Figure 16.1.2 The global temperature trend over the past 65 Ma (the Cenozoic). From the end of the Paleocene to the height of the Pleistocene Glaciation, global average temperature dropped by about 14°C. (PETM is the Palecene-Eocene thermal maximum.)

At 40 Ma, ongoing plate motion widened the narrow gap between South America and Antarctica, resulting in the opening of the Drake Passage. This allowed for the unrestricted west-to-east flow of water around Antarctica, the Antarctic Circumpolar Current (Figure 16.1.3), which effectively isolated the southern ocean from the warmer waters of the Pacific, Atlantic, and Indian Oceans. The region cooled significantly, and by 35 Ma (Oligocene) glaciers had started to form on Antarctica

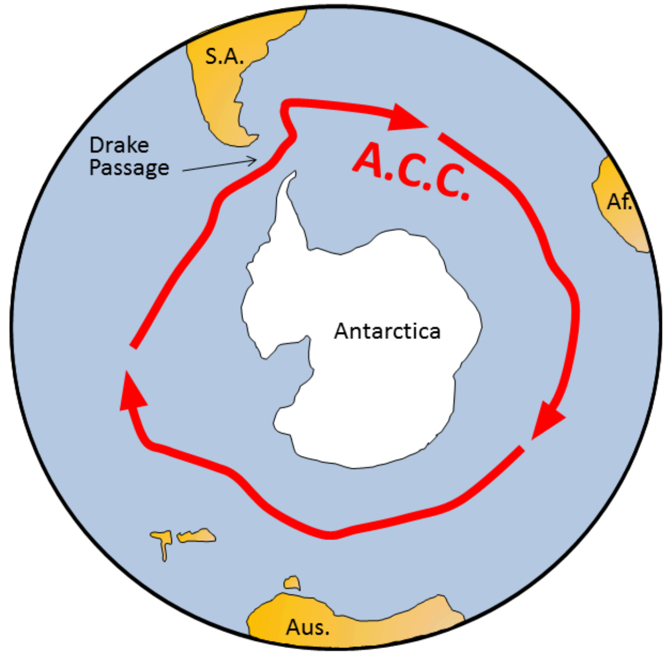

Figure 16.1.3 The Antarctic Circumpolar Current (red arrows) prevents warm water from the rest of Earth's oceans from getting close to Antarctica.

Global temperatures remained relatively steady during the Oligocene and early Miocene, and the Antarctic glaciation waned during that time. At around 15 Ma, subduction-related volcanism between central and South America created the connection between North and South America, preventing water from flowing between the Pacific and Atlantic Oceans. This further restricted the transfer of heat from the tropics to the poles, leading to a rejuvenation of the Antarctic glaciation. The expansion of that ice sheet increased Earth's reflectivity enough to promote a positive feedback loop of further cooling: more reflective glacial ice, more cooling, more ice, etc. By the Pliocene (~5 Ma) ice sheets had started to grow in North America and northern Europe (Figure 16.1.4). The most intense part of the current glaciation — and the coldest climate — has been during the past million years (the last one-third of the Pleistocene), but if we count Antarctic glaciation, it really extends from the Oligocene to the Holocene, and will likely continue into the future.

The Pleistocene has been characterized by significant temperature variations (through a range of almost 10°C) on time scales of 40,000 to 100,000 years, and corresponding expansion and contraction of ice sheets. These variations are attributed to subtle changes in Earth's orbital parameters (Milankovitch cycles), which are explained in more detail in Chapter 21. Over the past million years, the glaciation cycles have been approximately 100,000 years; this variability is visible in Figure 16.1.4.

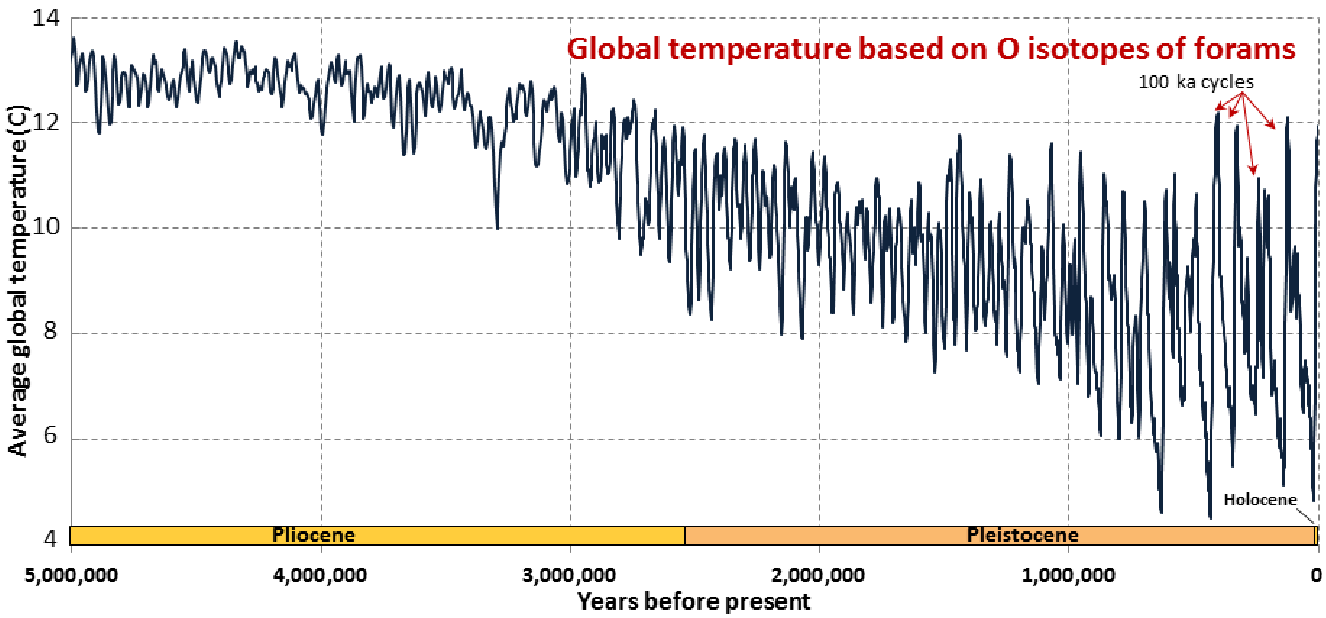

Figure 16.1.4 Foram oxygen isotope record for the past 5 million years based on O isotope data from sea-floor sediments shows a decrease in global temperture.

Exercise 16.1 Pleistocene glacials and interglacials

This diagram (Figure 16.1.5) shows the past 500,000 years of global temperature variations based on the same data used for Figure 16.1.4. The last five glacial periods are marked with snowflakes. The most recent one, which peaked at around 20 ka, is known as the Wisconsin Glaciation. Describe the nature of temperature change that followed each of these glacial periods.

The current interglacial (Holocene) is marked with an H. Point out the previous five interglacial periods.

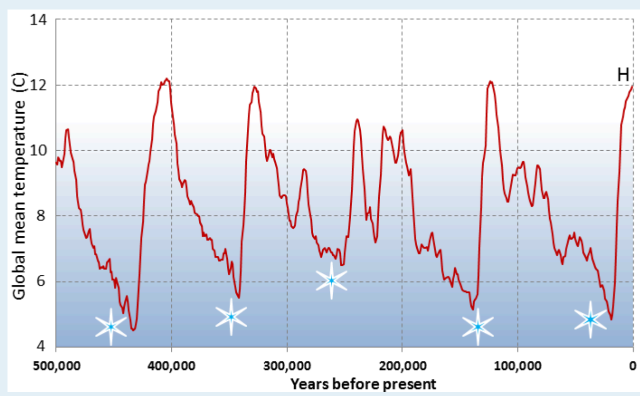

Figure 16.1.5

See Appendix 3 for Exercise 16.1 answers.

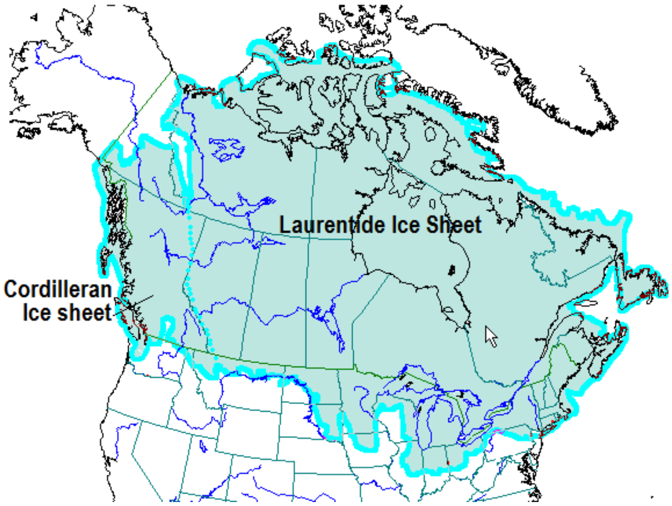

Figure 16.7 The extent of the Cordilleran and Laurentide Ice Sheets near the peak of the Wisconsin Glaciation, around 15 ka.

At the height of the last glaciation (**Wisconsin Glaciation**), massive ice sheets covered almost all of Canada and much of the northern United States (Figure 16.1.6). (In fact glacial ice extended well south of Wisconsin, into Illinois, Indiana and Ohio.) The massive **Laurentide Ice Sheet** covered most of eastern Canada (and adjacent USA), as far west as the Rockies, and the smaller **Cordilleran Ice Sheet** covered most of the western region. At various other glacial peaks during the Pleistocene and Pliocene, the ice extent was similar to this, and in some cases, even more extensive. The combined Laurentide and Cordilleran Ice Sheets were comparable in volume of ice to the current Antarctic Ice Sheet.

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- Figure 16.1.6: © Steven Earle. CC BY. Based on a NOAA Map.

16.2 How Glaciers Work

There are two main types of glaciers. **Continental glaciers** cover vast areas of land in extreme polar regions, including Antarctica and Greenland (Figure 16.2.1). **Alpine glaciers** (a.k.a. valley glaciers) originate on mountains, mostly in temperate and polar regions (Figure 16.0.1), but even in tropical regions if the mountains are high enough.

Figure 16.2.1 Part of the continental ice sheet in Greenland, with some outflow alpine glaciers in the foreground.

Earth's two great continental glaciers, on Antarctica and Greenland, comprise about 99% of all of the world's glacial ice, and approximately 68% of all of Earth's fresh water. As is evident from Figure 16.2.2, the Antarctic Ice Sheet is vastly bigger than the Greenland Ice Sheet; it contains about 17 times as much ice. If the entire Antarctic Ice Sheet were to melt, sea level would rise by about 80 m and most of Earth's major cities would be completely submerged.

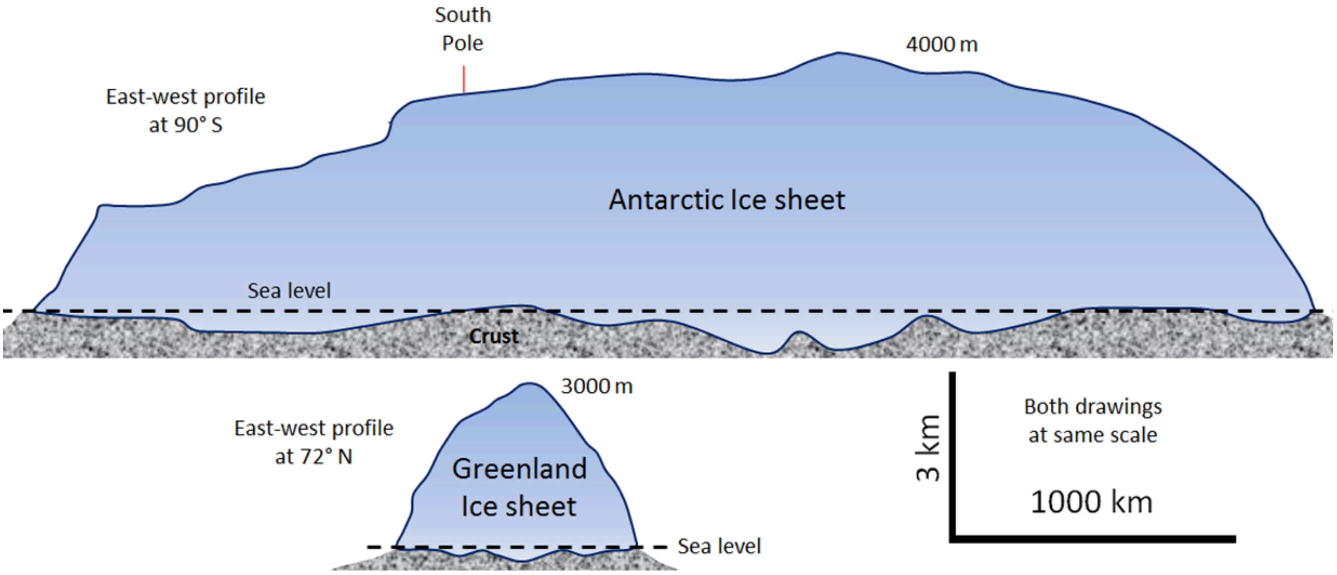

Figure 16.2.2 Simplified cross-sectional profiles the continental ice sheets in Greenland and Antarctica – both drawn to the same scale. [Image Description]

Continental glaciers do not flow "downhill" because the large areas that they cover are generally flat. Instead, ice flows from the region where it is thickest toward the edges where it is thinner, as shown in Figure 16.2.3. This means that in the central thickest parts, the ice flows almost vertically down toward the base, while in the peripheral parts, it flows out toward the margins. In continental glaciers like Antarctica and Greenland, the thickest parts (4,000 m and 3,000 m respectively) are the areas where the rate of snowfall and therefore the rate of ice accumulation are highest.

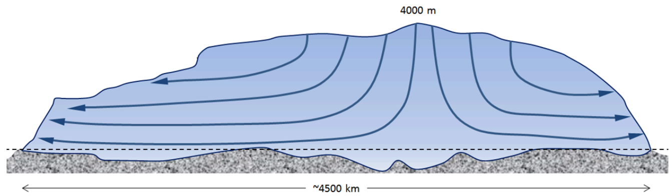

Figure 16.2.3 Schematic ice-flow diagram for the Antarctic Ice Sheet.

The flow of alpine glaciers is primarily controlled by the slope of the land beneath the ice (Figure 16.2.4). In the **zone of accumulation**, the rate of snowfall is greater than the rate of melting. In other words, not all of the snow that falls each winter melts during the following summer, and the ice surface is always covered with snow. In the **zone of ablation**, more ice melts than accumulates as snow. The **equilibrium line** marks the boundary between the zones of accumulation (above) and ablation (below).

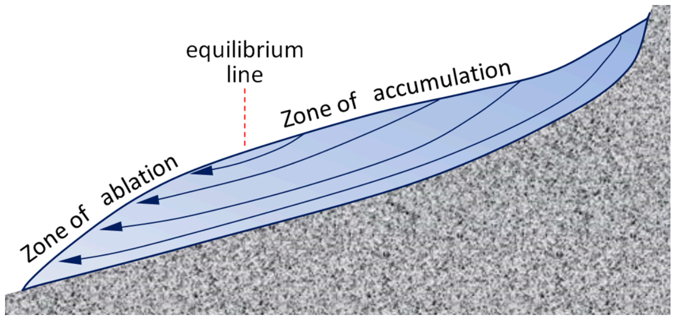

Figure 16.2.4 Schematic ice-flow diagram for an alpine glacier.

Above the equilibrium line of a glacier, not all of the winter snow melts in the following summer, so snow gradually accumulates. The snow layer from each year is covered and compacted by subsequent snow, and it is gradually compressed and turned into **firn** within which the snowflakes lose their delicate shapes and become granules. With more compression, the granules are pushed together and air is squeezed out. Eventually the granules are "welded" together to create glacial ice (Figure 16.2.5). Downward percolation of water from melting taking place at the surface contributes to the process of ice formation.

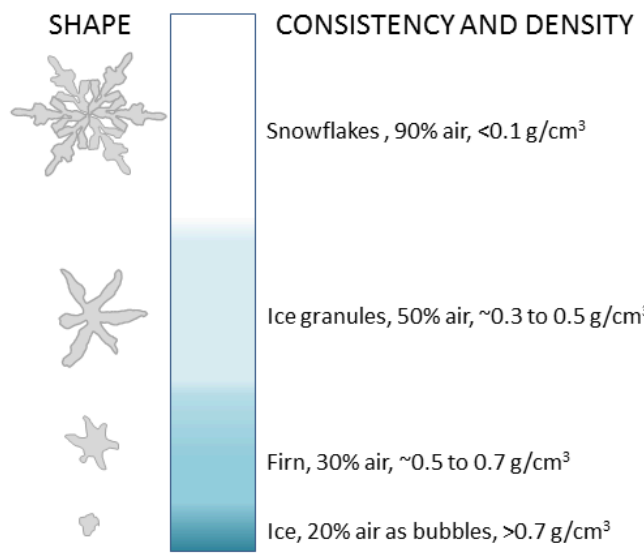

Figure 16.2.5 Steps in the process of formation of glacial ice from snow, granules, and firn. [Image Description]

The equilibrium line of a glacier near Whistler, B.C., is shown in Figure 16.2.6. Below that line, in the zone of ablation, bare ice is exposed because last winter's snow has all melted; above that line, the ice is still mostly covered with snow from last winter. The position of the equilibrium line changes from year to year as a function of the balance between snow accumulation in the winter and snowmelt during the summer. More winter snow and less summer melting obviously favours the advance of the equilibrium line (and of the glacier's leading edge), but of these two variables, it is the summer melt that matters most to a glacier's budget. Cool summers promote glacial advance and warm summers promote glacial retreat.

Glaciers move because the surface of the ice is sloped. This generates a stress on the ice, which is proportional to the slope and to the depth below the surface. As shown in Figure 16.2.6, the stresses are quite small near the ice surface but much larger at depth, and also greater in areas where the ice surface is relatively steep. Ice will deform, meaning that it will behave in a plastic manner, at stress levels of around 100 kilopascals; therefore, in the upper 50 m to 100 m of the ice (above the dashed red line), flow is not plastic (the ice is rigid), while below that depth, ice is plastic and will flow.

Figure 16.2.6 The approximate location of the equilibrium line (red) in September 2013 on the Overlord Glacier, near Whistler, B.C.

When the lower ice of a glacier flows, it moves the upper ice along with it, so although it might seem from the stress patterns (red numbers and red arrows) shown in Figure 16.2.7 that the lower part moves the most, in fact while the lower part deforms (and flows) and the upper part doesn't deform at all, the upper part moves the fastest because it is pushed along by the lower ice.

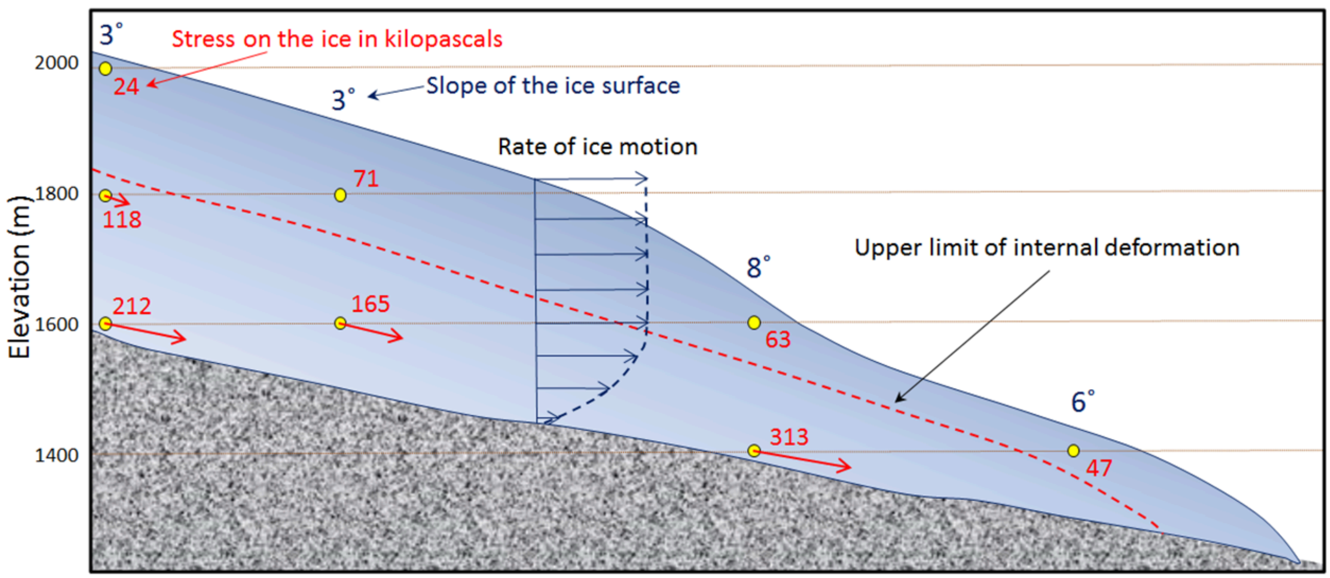

Figure 16.2.7 Stress within a valley glacier (red numbers) as determined based on the slope of the ice surface and the depth within the ice. The ice will deform and flow where the stress is greater than 100 kilopascals, and the relative extent of that deformation is depicted by the red arrows. Any deformation motion in the lower ice will be transmitted to the ice immediately above it, so although the red stress arrows get shorter toward the top, the ice velocity increases upward (blue arrows). The upper ice (above the red dashed line) does not flow, but it is pushed along with the lower ice.

The plastic lower ice of a glacier can flow like a very viscous fluid, and can therefore flow over irregularities in the base of the ice and around corners. However, the upper rigid ice cannot flow in this way, and because it is being carried along by the lower ice, it tends to crack where the lower ice has to flex. This leads to the development of **crevasses** in areas where the rate of flow of the plastic ice is changing. In the area shown in Figure 16.2.8, for example, the glacier is speeding up over the steep terrain, and the rigid surface ice has to crack to account for the change in velocity.

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Figure 16.2.8 Crevasses on Overlord Glacier in the Whistler area, B.C.

The base of a glacier can be cold (below the freezing point of water) or warm (above the freezing point). If it is warm, there will likely be a film of water between the ice and the material underneath, and the ice will be able to slide over that surface. This is known as **basal sliding** (Figure 16.2.9, left). If the base is cold, the ice will be frozen to the material underneath and it will be stuck—unable to slide along its base. In this case, all of the movement of the ice will be by internal flow.

One of the factors that affects the temperature at the base of a glacier is the thickness of the ice. Ice is a good insulator. The slow transfer of heat from Earth's interior provides enough heat to warm up the base if the ice is thick, but not enough if it is thin and that heat can escape. It is typical for the leading edge of an alpine glacier to be relatively thin (see Figure 16.2.7), so it is common for that part to be frozen to its base while the rest of the glacier is still sliding. This is illustrated in Figure 16.2.10 for the Athabasca Glacier. Because the leading edge of the glacier is stuck to its frozen base, while the rest continues to slide, the ice coming from behind has pushed (or thrust) itself over top of the part that is stuck fast.

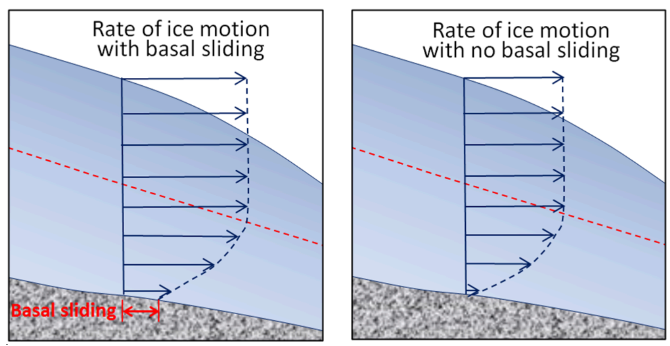

Figure 16.2.9 Differences in glacial ice motion with basal sliding (left) and without basal sliding (right). The dashed red line indicates the upper limit of plastic internal flow.

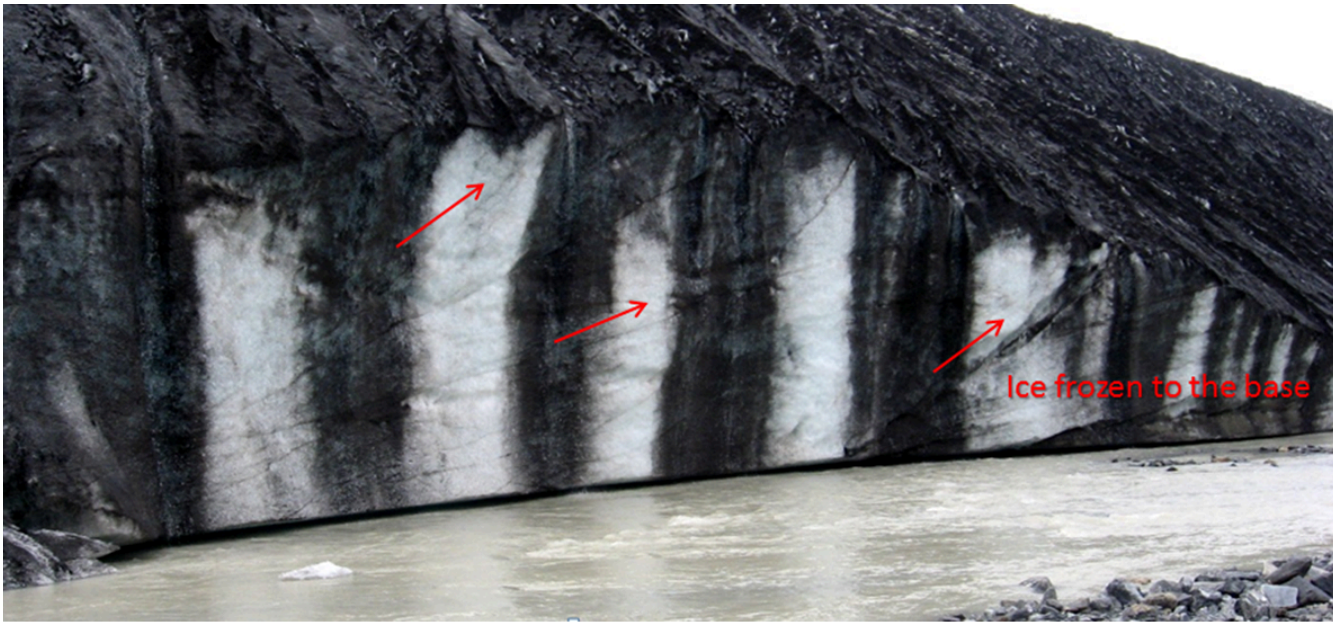

Figure 16.2.10 Thrust faults at the leading edge of the Athabasca Glacier, Alberta. The arrows show how the trailing ice has been thrust over the leading ice. (The dark vertical stripes are mud from sediments that have been washed off of the lateral moraine lying on the surface of the ice.)

Just as the base of a glacier moves more slowly than the surface, the edges, which are more affected by friction along the sides, move more slowly than the middle. If we were to place a series of markers across an alpine glacier and come back a year later, we would see that the ones in the middle had moved farther forward than the ones near the edges (Figure 16.2.11).

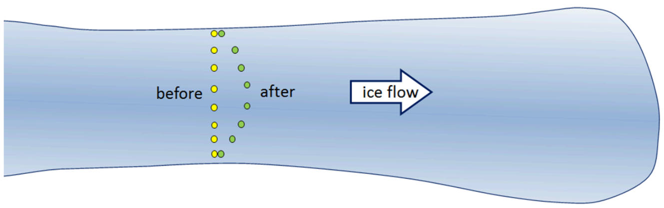

Figure 16.2.11 Markers on an alpine glacier move forward over a period of time with the ones in the middle moving faster than the ones on the edge.

Glacial ice always moves downhill, in response to gravity, but the front edge of a glacier is always either melting or **calving** into water (shedding icebergs). If the rate of forward motion of the glacier is faster than the rate of **ablation** (melting), the leading edge of the glacier advances (moves forward). If the rate of forward motion is about the same as the rate of ablation, the leading edge remains stationary, and if the rate of forward motion is slower than the rate of ablation, the leading edge retreats (moves backward).

Calving of icebergs is an important process for glaciers that terminate in lakes or the ocean. An example of such a glacier is the Berg Glacier on Mt. Robson (Figure 16.2.12), which sheds small icebergs into Berg Lake. The Berg Glacier also loses mass by melting, especially at lower elevations.

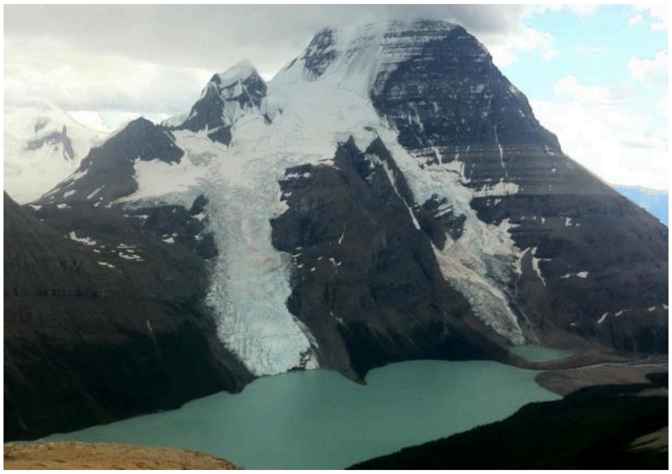

Figure 16.2.12 Mt. Robson, the tallest peak in the Canadian Rockies, Berg Glacier (centre), and Berg Lake. Although there were no icebergs visible when this photo was taken, the Berg Glacier loses mass by shedding icebergs into Berg Lake.

Exercise 16.2 Ice advance and retreat

These diagrams represent a glacier with markers placed on its surface to determine the rate of ice motion over a one-year period. The ice is flowing from left to right.

1. In the middle diagram, the leading edge of the glacier has advanced. Draw in the current position of the markers.
2. In the lower diagram, the leading edge of the glacier has retreated. Draw in the current position of the markers.

See Appendix 3 for Exercise 16.2 answers.

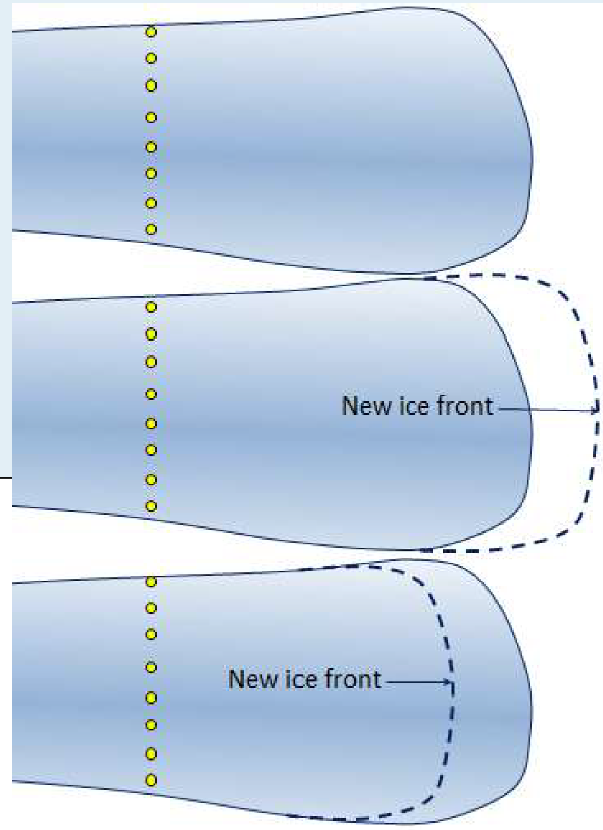

Figure 16.2.13

Image Descriptions

Figure 16.2.2 image description: An east to west cross section of the ice on Antarctica is over 4000 km long and up to 4 km high in some places. An east to west cross section of the ice on Greenland is less than 1000 km long and up to 3 km high. [Return to Figure 16.2.2]

Figure 16.2.5 image description: The comparative density of snowflakes, ice granules, firn, and ice.

Shape	**Amount of air**	**Density (grams per centimetres cubed)**
Snowflake	90%	Less than 0.1
Ice Granules	50%	From around 0.3 to 0.5
Firn	30%	From around 0.5 to 0.7
Ice	20% (as bubbles)	Greater than 0.7

[Return to Figure 16.2.5]

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16.3 Glacial Erosion

Glaciers are effective agents of erosion, especially in situations where the ice is not frozen to its base and can therefore slide over the bedrock or other sediment. In fact the ice itself is not particularly effective at erosion because it is relatively soft (Mohs hardness 1.5 at 0°C); instead, it is the rock fragments embedded in the ice and pushed down onto the underlying surfaces that do most of the erosion. A useful analogy would be to compare the effect of a piece of paper being rubbed against a wooden surface, as opposed to a piece of sandpaper that has embedded angular fragments of garnet.

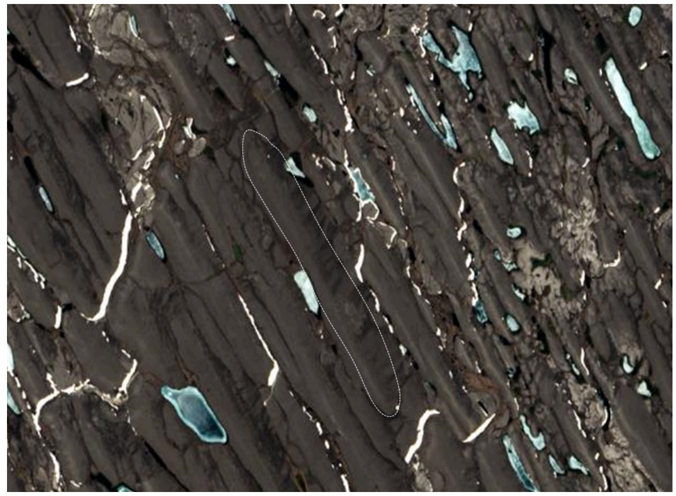

Figure 16.3.1 Drumlins — streamlined hills formed beneath a glacier, here made up of sediment — in the Amundsun Gulf region of Nunavut. The drumlins are tens of metres high, a few hundred metres across, and a few kilometres long. One of them is highlighted with a dashed white line.

The results of glacial erosion are different in areas with continental glaciation versus alpine glaciation. Continental glaciation tends to produce relatively flat bedrock surfaces, especially where the rock beneath is uniform in strength. In areas where there are differences in the strength of rocks, a glacier obviously tends to erode the softer and weaker rock more effectively than the harder and stronger rock. Much of central and eastern Canada, which was completely covered by the huge Laurentide Ice Sheet at various times during the Pleistocene, has been eroded to a relatively flat surface. In many cases the existing relief is due the presence of glacial deposits—such as drumlins, eskers, and moraines (all discussed below)—rather than to differential erosion (Figure 16.3.1).

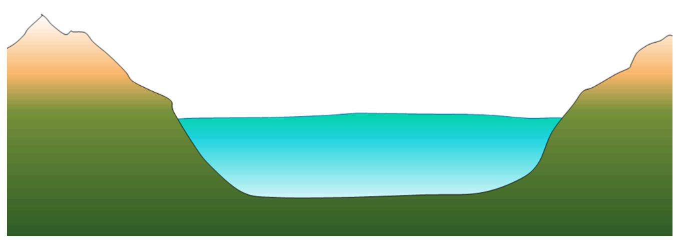

Figure 16.3.2 A depiction of a U-shaped valley occupied by a large glacier.

Alpine glaciers produce very different topography than continental glaciers, and much of the topographic variability of western Canada can be attributed to glacial erosion. In general, glaciers are much wider than rivers of similar length, and since they tend to erode more at their bases than their sides, they produce wide valleys with relatively flat bottoms and steep sides—known as **U-shaped valleys** (Figure 16.3.2). Howe Sound, north of Vancouver, was occupied by a large glacier that originated in the Squamish, Whistler, and Pemberton areas, and then joined the much larger glacier in the Strait of Georgia. Howe Sound and most of its tributary valleys have pronounced U-shaped profiles (Figure 16.3.3).

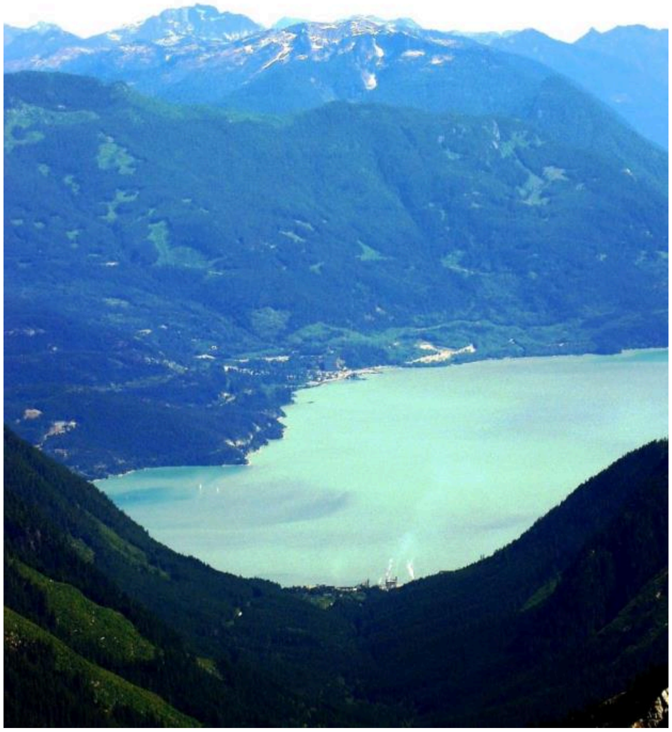

Figure 16.3.3 The view down the U-shaped valley of Mill Creek valley toward the U-shaped valley of Howe Sound (north of Vancouver BC), with the village of Britannia on the opposite side.

U-shaped valleys and their tributaries provide the basis for a wide range of alpine glacial topographic features, examples of which are visible on the International Space Station view of the Swiss Alps shown in Figure 16.3.4. This area was much more intensely glaciated during the past glacial maximum. At that time, the large U-shaped valley in the lower right was occupied by glacial ice, and all of the other glaciers shown here were longer and much thicker than they are now. But even at the peak of the Pleistocene Glaciation, some of the higher peaks and ridges would have been exposed and not directly affected by glacial erosion. A peak that extends above the surrounding glacier is called a **nunatuk**. In these areas, and in the areas above the glaciers today, most of the erosion is related to freeze-thaw effects.

Figure 16.3.4 is a Space Station view of a glaciated terrain in the Swiss Alps. Some of the important features visible are **arêtes**: sharp ridges between U-shaped glacial valleys; **cols**: low points along arêtes that constitute passes between glacial valleys; **horns**: steep peaks that have been glacially and freeze-thaw eroded on three or more sides; **cirques**: bowl-shaped basins that form at the head of a glacial valley; **hanging valleys**: U-shaped valleys of tributary glaciers that hang above the main valley because the larger main-valley glacier eroded more deeply into the terrain; and **truncated spurs** (a.k.a. "spurs"): the ends of arêtes that have been eroded into steep triangle-shaped cliffs by the glacier in the corresponding main valley.

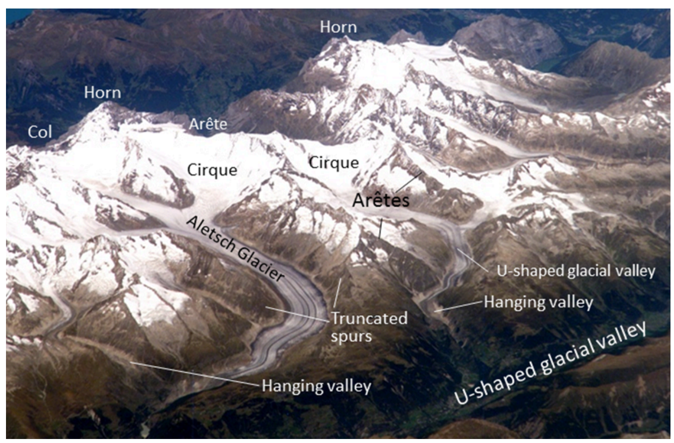

Figure 16.3.4 A view from the International Space Station of the Swiss Alps in the area of the Aletsch Glacier. The prominent peaks labelled "Horn" are the Eiger (left) and Wetterhorn (right). A variety of alpine glacial erosion features are labelled.

Some of these alpine-glaciation erosional features are also shown in Figure 16.3.5 in diagram form.

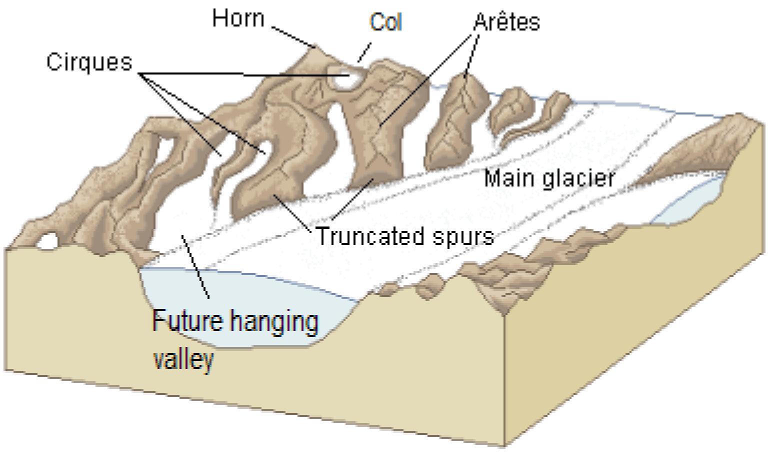

Figure 16.3.5 A diagram of some of the important alpine-glaciation erosion features.

Exercise 16.3 Identify alpine glacial erosion features

This is a photo of Mt. Assiniboine in the B.C. Rockies. What are the features at locations **a** through **e**? Look for one of each of the following: a horn, an arête, a truncated spur, a cirque, and a col. Try to identify some of the numerous other arêtes in this view, as well as another horn.

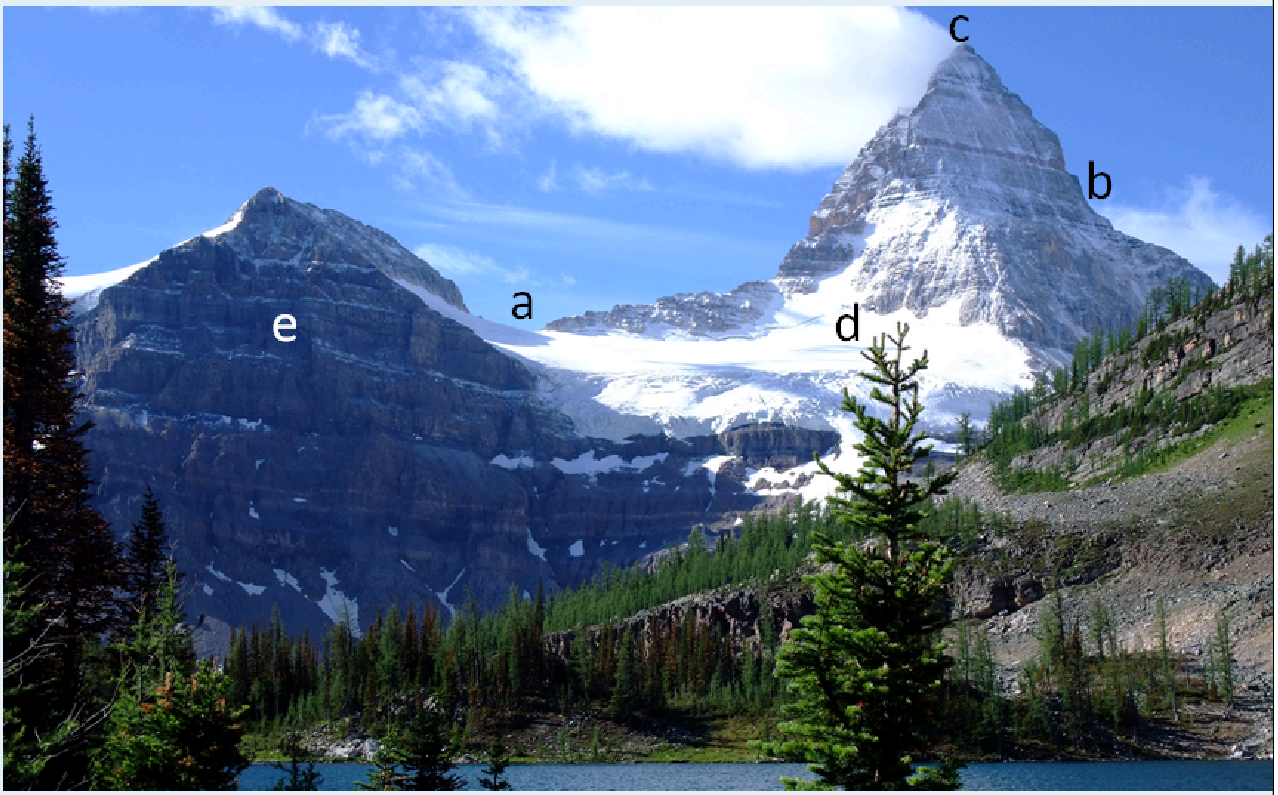

Figure 16.3.6 [Image Description]

See Appendix 3 for Exercise 16.3 answers.

A number of other glacial erosion features exist at smaller scales. For example, a **drumlin** is an elongated feature that is streamlined at the down-ice end. The one shown in Figure 16.3.7 is larger than most, and is made up almost entirely of rock. Drumlins made up of glacial sediments are very common in some areas of continental glaciation (Figure 16.3.1).

Figure 16.3.7 Bowyer Island, a drumlin in Howe Sound, B.C. Ice flow was from right to left.

A **roche moutonée** is another type of elongated erosional feature that has a steep and sometimes jagged down-ice end (Figure 16.3.8, left). On a smaller scale still, **glacial grooves** (tens of centimetres to metres wide) and **glacial striation** (millimetres to centimetres wide) are created by fragments of rock embedded in the ice at the base of a glacier (Figure 16.3.8, left and right). Glacial striae are very common on rock surfaces eroded by both alpine and continental glaciers.

Figure 16.3.8 Left: Roches moutonées with glacial striae near Squamish, B.C. Right: Glacial striae at the same location near Squamish. Ice flow was from right to left in both cases. The brown rock in the right-hand photo is a mafic dyke about 40 cm wide intruding into granite.

Glacial lakes

Lakes are common features in glacial environments. A lake that is confined to a glacial cirque is known as a **tarn** (Figure 16.3.9). Tarns are common in areas of alpine glaciation because the ice that forms a cirque typically carves out a depression in bedrock that then fills with water. In some cases, a series of such basins will form, and the resulting lakes are called **rock basin lakes** or **paternoster lakes**.

Figure 16.3.9 Lower Thornton Lake, a tarn, in the Northern Cascades National Park, Washington.

A lake that occupies a glacial valley, but is not confined to a cirque, is known as a **finger lake**. In some cases, a finger lake is confined by a dam formed by an end moraine, in which case it may be called a **moraine lake** (Figure 16.3.10).

Figure 16.3.10 Peyto Lake in the Alberta Rockies, is both a finger lake and a moraine lake as it is dammed by an end moraine, on the right.

In areas of continental glaciation, the crust is depressed by the weight of glacial ice that is up to 4,000 m thick. Basins are formed along the edges of continental glaciers (except for those that cover entire continents like Antarctica and Greenland), and these basins fill with glacial meltwater. Many such lakes, some of them huge, existed at various times along the southern edge of the Laurentide Ice Sheet. One example is Glacial Lake Missoula, which formed within Idaho and Montana in northwestern United States. During the latter part of the last glaciation (30 ka to 15 ka), the ice holding back Lake Missoula retreated enough to allow some of the lake water to start flowing out, which escalated into a massive and rapid outflow (over days to weeks) during which much of the volume of the lake drained along the valley of the Columbia River to the Pacific Ocean. It is estimated that this type of flooding happened at least 25 times over that period, and in many cases, the rate of outflow was equivalent to the discharge of all of Earth's current rivers combined. The record of these massive floods is preserved in the Channelled Scablands of Idaho, Washington, and Oregon (Figure 16.3.11).

Figure 16.3.11 Potholes Coulee near Wenatchee, Washington, one of many basins that received Lake Missoula floodwaters during the late Pleistocene. Here the water flowed from right to left, over the cliff and into this basin.

Another type of glacial lake is a **kettle lake**. These are discussed in section 16.4 in the context of glacial deposits.

Image Descriptions

Figure 16.3.6 image description:

1. A valley between two peaks.
2. A steep edge of a mountain.
3. A mountain peak.
4. An ice field in between two mountains.
5. A flat, rocky side of a mountain.

[Return to Figure 16.3.6]

Media Attributions

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- Figure 16.3.2: "Drumlin Field in Northern Canada" by NASA Earth Observatory. Public domain.
- Figure 16.3.3: "Woodf1a" © Keefer4. CC BY-SA**.**
- Figure 16.3.4: "Bernese Alps, Switzerland" by NASA Earth Observatory. Adapted by Steven Earle. Public domain.
- Figure 16.3.5: "Glacial landscape LMB" by Luis María Benítez. Public domain.

16.4 Glacial Deposition

Sediments transported and deposited during the Pleistocene glaciations are abundant throughout Canada and much of the northern USA. They are important sources of construction materials and are valuable as reservoirs for groundwater. Because they are almost all unconsolidated, they have significant implications for mass wasting.

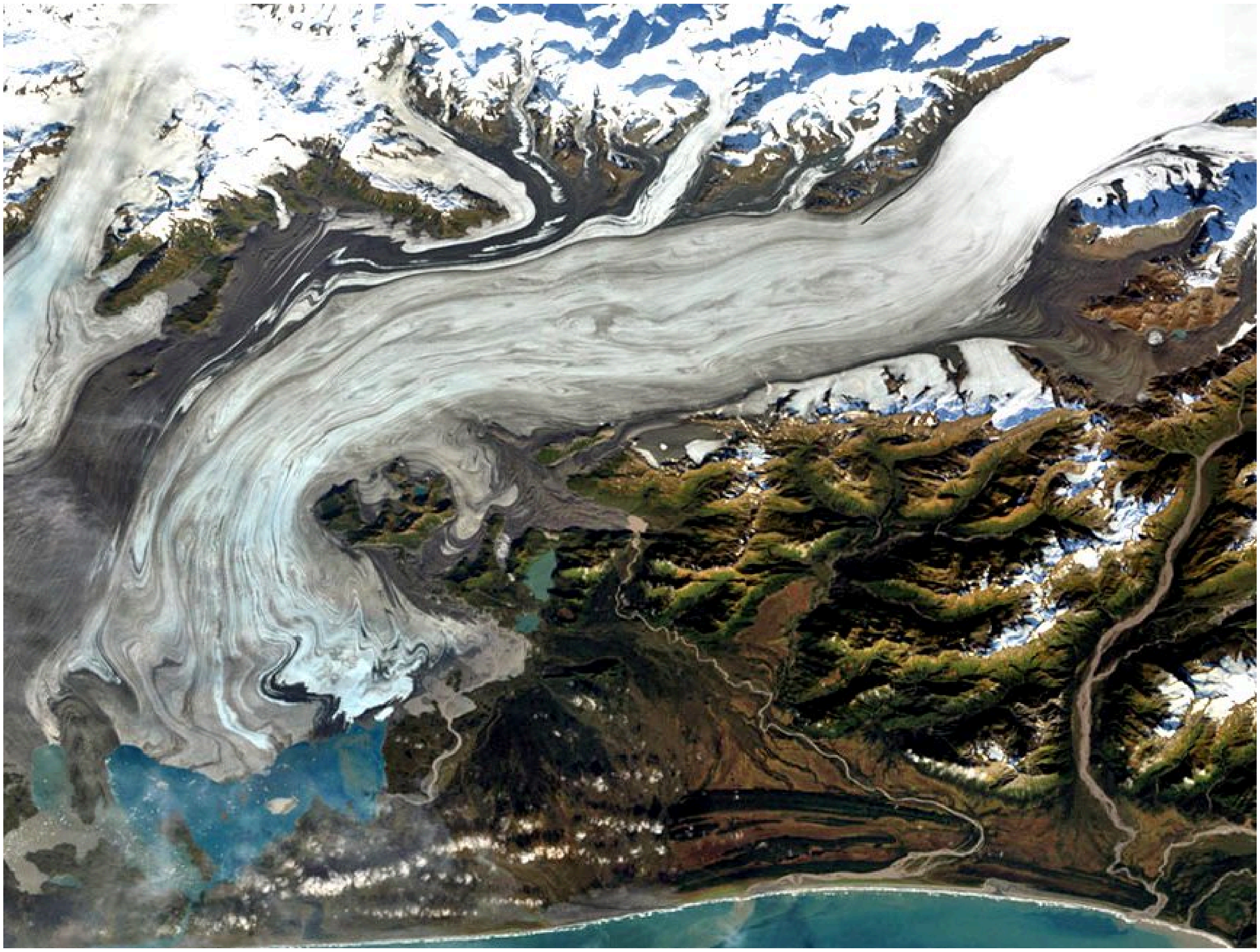

Figure 16.4.1 Part of the Bering Glacier in southeast Alaska, the largest glacier in North America. It is about 14 km across in the centre of this view.

Figure 16.4.1 illustrates some of the ways that sediments are transported and deposited. The Bering Glacier is the largest in North America, and although most of it is in Alaska, it flows from an icefield that extends into southwestern Yukon. The surface of the ice is partially, or in some cases completely covered with rocky debris that has fallen from surrounding steep rock faces. There are muddy rivers issuing from the glacier in several locations, depositing sediment on land, into Vitus Lake, and directly into the ocean. There are dirty icebergs shedding their sediment into the lake. And, not visible in this view, there are sediments being moved along beneath the ice.

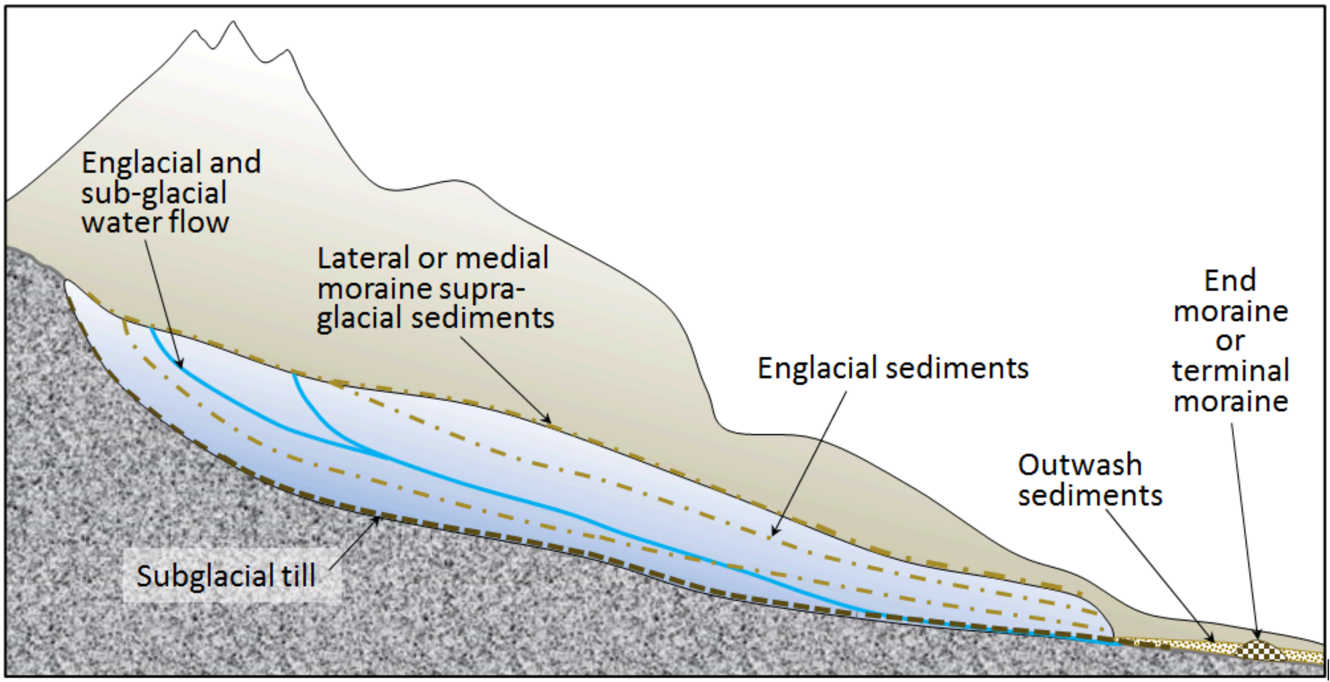

Figure 16.4.2 A depiction of the various types of sediments associated with glaciation. The glacier is shown in cross-section.

The formation and movement of sediments in glacial environments is shown diagrammatically in Figure 16.4.2. There are many types of glacial sediment generally classified by whether they are transported on, within, or beneath the glacial ice. The main types of sediment in a glacial environment are described below.

Supraglacial (on top of the ice) and **englacial** (within the ice) sediments that slide off the melting front of a stationary glacier can form a ridge of unsorted sediments called an **terminal moraine**. The end moraine that represents the farthest advance of the glacier is a **end moraine**. Supraglacial and englacial sediments can also be deposited when the ice melts. Sediments transported and deposited by glacial ice are known as **till**.

Subglacial sediment (e.g., **lodgement till**) is material that has been eroded from the underlying rock by the ice, and is moved by the ice. It has a wide range of grain sizes (in other words it is poorly sorted), including a relatively high proportion of silt and clay. The larger clasts (pebbles to boulders in size) tend to become partly rounded by abrasion. Lodgement till forms as a sheet of well-compacted sediment beneath a glacier, and ranges from several centimetres to many metres in thickness. Lodgement till is normally unbedded. An example is shown in Figure 16.4.3a.

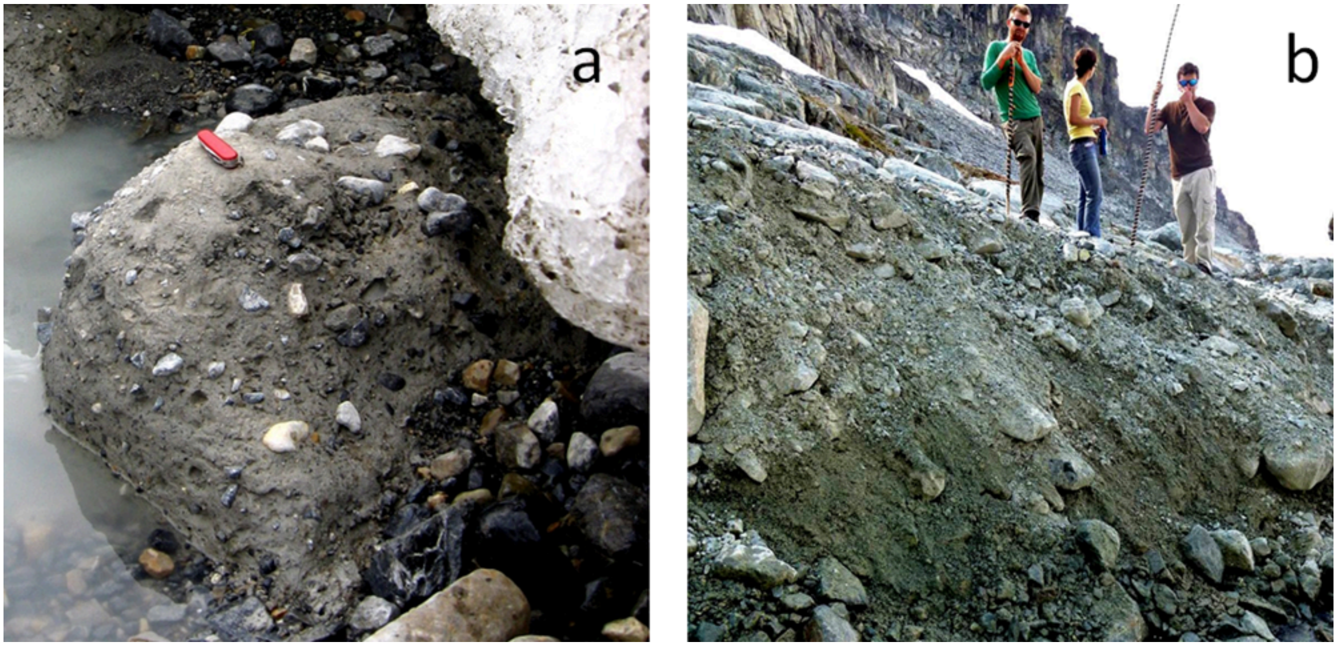

Figure 16.4.3 Examples of glacial till: a: lodgement till from the front of the Athabasca Glacier, Alberta; b: ablation till at the Horstman Glacier, Blackcomb Mountain, B.C.

Supraglacial sediments are primarily derived from freeze-thaw eroded material that has fallen onto the ice from rocky slopes above. These sediments form **lateral moraines** (Figure 16.0.1) and, where two glaciers meet, **medial moraines**. (Medial moraines are visible on the Aletsch Glacier in Figure 16.3.4.) Most of this material is deposited on the ground when the ice melts, and is therefore called **ablation till**, a mixture of fine and coarse angular rock fragments, with much less sand, silt, and clay than lodgement till. An example is shown in Figure 16.4.3b. When supraglacial sediments become incorporated into the body of the glacier, they are known as englacial sediments (Figure 16.4.2).

Massive amounts of water flow on the surface, within, and at the base of a glacier, even in cold areas and even when the glacier is advancing. Depending on its velocity, this water is able to move sediments of various sizes and most of that material is washed out of the lower end of the glacier and deposited as outwash sediments. These sediments accumulate in a wide range of environments in the **proglacial** region (the area in front of a glacier), most in fluvial environments, but some in lakes and the ocean. **Glaciofluvial sediments** are similar to sediments deposited in normal fluvial environments, and are dominated by silt, sand, and gravel. The grains tend to be moderately well rounded, and the sediments have similar sedimentary structures (e.g., bedding, cross-bedding, clast imbrication) to those formed by non-glacial streams (Figure 16.4.4a and 16.4.4b).

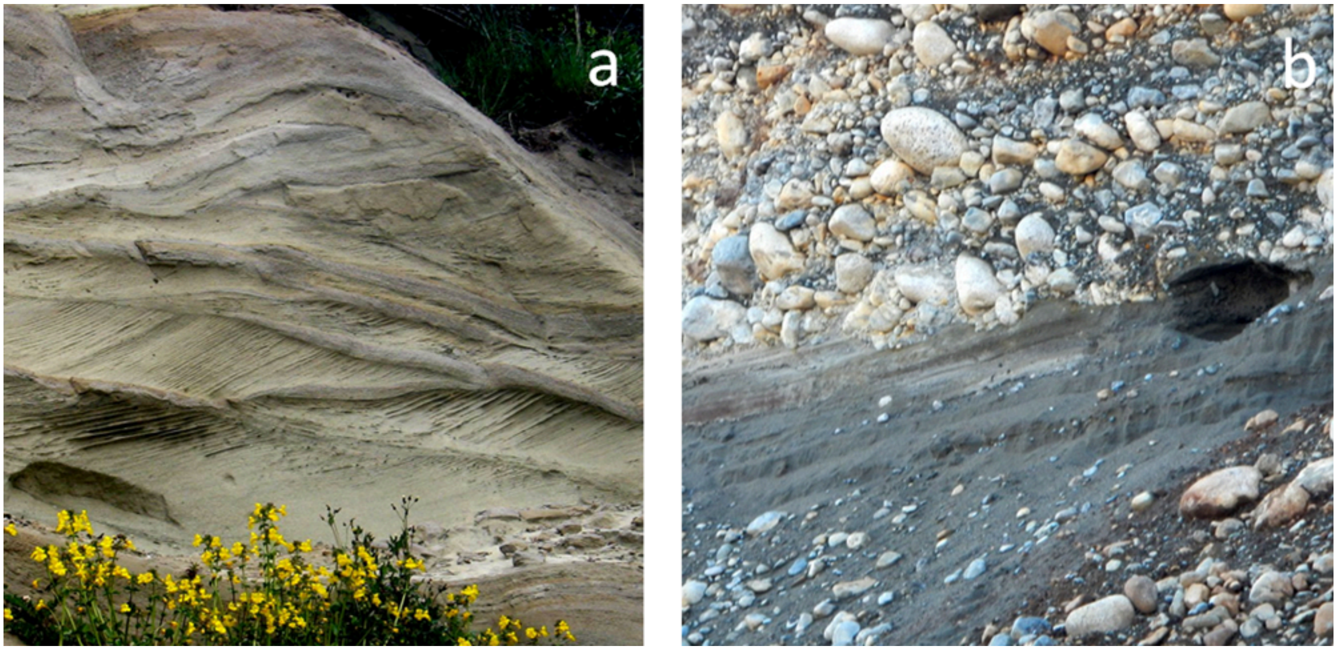

Figure 16.4.4 Examples of glaciofluvial sediments: a: glaciofluvial sand of the Quadra Sand Formation at Comox, B.C.; b: glaciofluvial gravel and sand, Nanaimo, B.C.

A large proglacial plain of sediment is called a **sandur** (a.k.a. an **outwash plain**)**,** and within that area, glaciofluvial deposits can be tens of metres thick (Figure 16.4.5). In situations where a glacier is receding, a block of ice might become separated from the main ice sheet and then get buried in glaciofluvial sediments. When the ice block eventually melts, a depression forms, known as a **kettle**, and if this fills with water, it is known as a **kettle lake** (Figure 16.4.6).

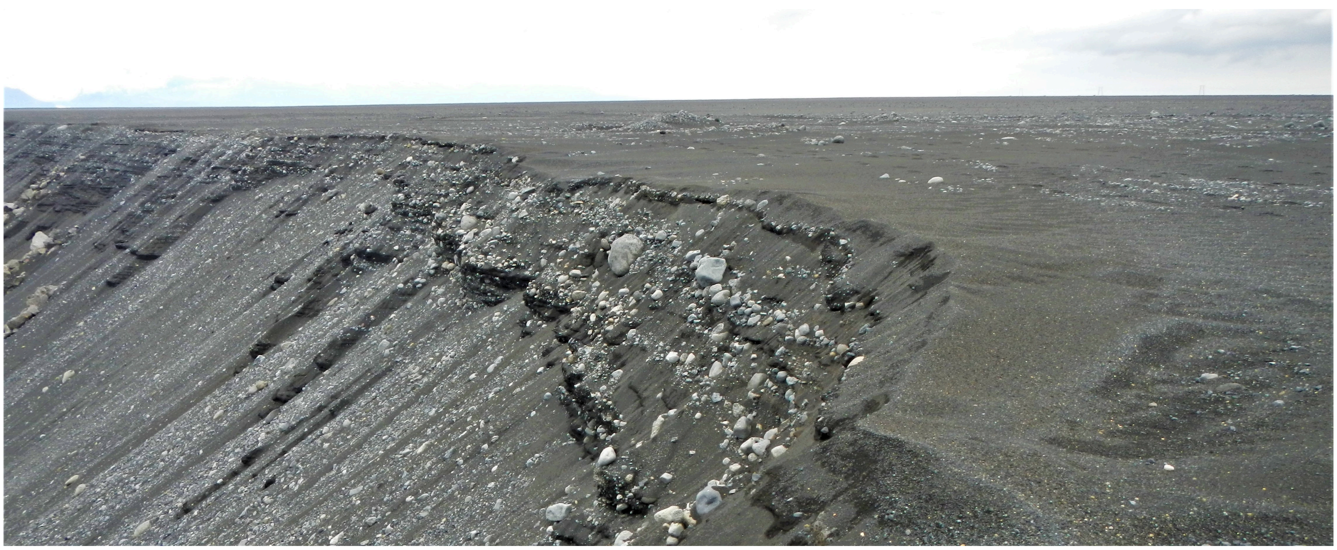

Figure 16.4.5 Part of a sandur in front of the Vatnajokull Glacier in Iceland,. The Sandur extends for many tens of km. Here is has been partially eroded by a stream (not visible).

Figure 16.4.6 A kettle lake amid vineyards and orchards in the Osoyoos area of B.C.

Figure 16.4.7 Part of an esker that formed beneath the Laurentide Ice Sheet in northern Canada.

A subglacial stream will create its own channel within the ice, and sediments that are being transported and deposited by the stream will build up within that channel. When the ice recedes, the sediment will remain to form a long sinuous ridge known as an **esker**. Eskers are most common in areas of continental glaciation. They can be several metres high, tens of metres wide, and tens of kilometres long (Figure 16.4.7).

Outwash streams commonly flow into proglacial lakes where **glaciolacustrine sediments** are deposited. These are dominated by silt- and clay-sized particles and are typically laminated on the millimetre scale. In some cases, **varves** develop; varves are series of beds with distinctive summer and winter layers: relatively coarse in the summer when melt discharge is high, and finer in the winter, when discharge is very low. Icebergs are common on proglacial lakes, and most of them contain englacial sediments of various sizes. As the bergs melt, the released clasts sink to the bottom and are incorporated into the glaciolacustrine layers as **drop stones** (Figure 16.4.8a).

The processes that occur in proglacial lakes can also take place where a glacier terminates at the ocean. The sediments deposited there are called **glaciomarine sediments** (Figure 16.4.8b).

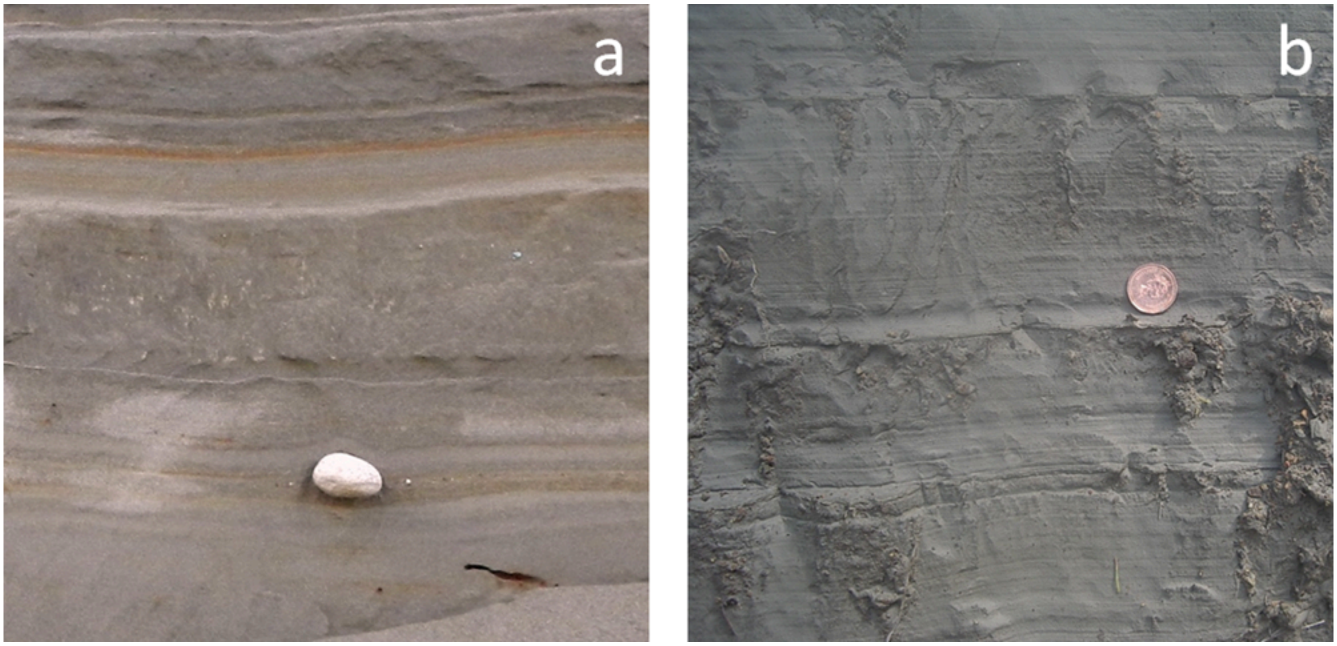

Figure 16.4.8 Examples of glacial sediments formed in quiet water: a: glaciolacustrine sediment with a drop stone, Nanaimo, B.C.; and b: a laminated glaciomarine sediment, Englishman River, B.C. Although not visible in this photo, the glaciomarine sediment has marine shell fossils.

Exercise 16.4 Identify glacial depositional environments

This photo shows the Bering Glacier in Alaska (same as Figure 16.4.1).

Glacial sediments of many different types are being deposited throughout this area. Identify where you would expect to find the following types of deposits:

1. Glaciofluvial sand
2. Lodgement till
3. Glaciolacustrine clay with drop stones
4. Ablation till
5. Glaciomarine silt and clay

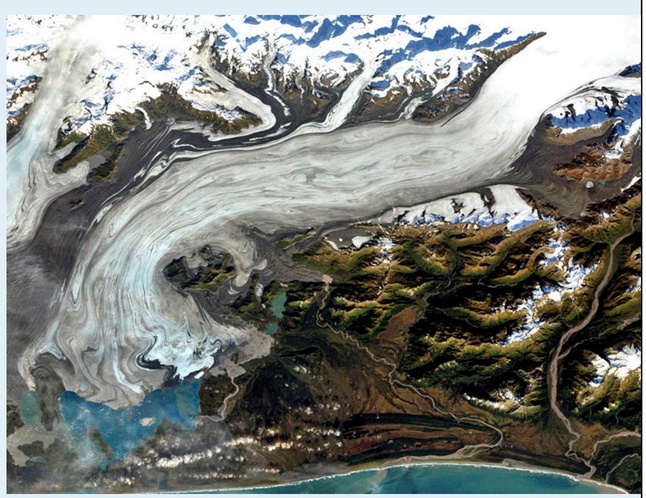

Figure 16.4.9

See Appendix 3 for Exercise 16.4 answers.

Media Attributions

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- Figures 16.4.2, 16.4.3, 16.4.4, 16.4.5, 16.4.6, 16.4.8: © Steven Earle. CC BY.
- Figure 16.4.7: “Esker on a morainal plain” © Canadian Soil Information Service (CANSIS). Approved for non-commercial reproduction.
- Figure 16.4.9: “Bering Glacier” by NASA Earth Observatory. Public domain.

Summary

The topics covered in this chapter can be summarized as follows:

Section	Summary
16.1 Glacial Periods in Earth's History	There have been many glaciations in Earth's distant past, the oldest known starting around 2,400 Ma. The late Proterozoic "Snowball Earth" glaciations were thought to be sufficiently intense to affect the entire planet. The current glacial period is known as the Pleistocene Glaciation, and while it was much more intense 20,000 years ago than it is now, we are still in the middle of it. The periodicity of the Pleistocene glaciations is related to subtle changes in Earth's orbital characteristics, which are exaggerated by a variety of positive feedback processes.
16.2 How Glaciers Work	The two main types of glaciers are continental glaciers, which cover large parts of continents, and alpine glaciers, which occupy mountainous regions. Ice accumulates at higher elevations—above the equilibrium line—where the snow that falls in winter does not all melt in summer. In continental glaciers, ice flows outward from where it is thickest. In alpine glaciers, ice flows downslope. At depth in the glacier ice, flow is by internal deformation, but glaciers that have liquid water at their base can also flow by basal sliding. Crevasses form in the rigid surface ice in places where the lower plastic ice is changing shape.
16.3 Glacial Erosion	Glaciers are important agents of erosion. Continental glaciers tend to erode the land surface into flat plains, while alpine glaciers create a wide variety of different forms. The key feature of alpine glacial erosion is the U-shaped valley. Arêtes are sharp ridges that form between two valleys, and horns form where a mountain is glacially eroded on at least three sides. Because tributary glaciers do not erode as deeply as main-valley glaciers, hanging valleys exist where the two meet. On a smaller scale, both types of glaciers form drumlins, roches moutonées, and glacial grooves or striae.
16.4 Glacial Deposition	Glacial deposits are quite varied, as materials are transported and deposited in a variety of different ways in a glacial environment. Sediments that are moved and deposited directly by ice are known as till. Glaciofluvial sediments are deposited by glacial streams, either forming eskers or large proglacial plains known as sandurs. Glaciolacustrine and glaciomarine sediments originate within glaciers and are deposited in lakes and the ocean respectively.

Questions for Review

See Appendix 2 for answers to Review Questions.

1. Why are the Cryogenian glaciations called Snowball Earth?
2. Earth cooled dramatically from the end of the Paleocene until the Holocene. Describe some of the geological events that contributed to that cooling.
3. When and where was the first glaciation of the Cenozoic?

4. Describe the extent of the Laurentide Ice Sheet during the height of the last Pleistocene glacial period.
5. In an alpine glacier, the ice flows down the slope of the underlying valley. Continental glaciers do not have a sloped surface to flow down. What feature of a continental glacier facilitates its flow?
6. What does the equilibrium line represent in a glacier? Explain.
7. Which of the following is more important to the growth of a glacier: very cold winters or relatively cool summers? Explain.
8. Describe the relative rates of ice flow within the following parts of a glacier: (a) the bottom versus the top and (b) the edges versus the middle. Explain.
9. What condition is necessary for basal sliding to take place?
10. Why do glaciers carve U-shaped valleys, and how does a hanging valley form?
11. A horn is typically surrounded by cirques. What is the minimum number of cirques you would expect to find around a horn?
12. A drumlin and a roche moutonée are both streamlined glacial erosion features. How do they differ in shape?
13. Four examples of glacial sediments are shown here. Describe the important characteristics (e.g., sorting, layering, grain-size range, grain shape, sedimentary structures) of each one and give each a name (choose from glaciofluvial, glaciolacustrine, lodgement till, ablation till, and glaciomarine).

Figure A

14. What are drop stones, and under what circumstances are they likely to form?
15. What types of glacial sediments are likely to be sufficiently permeable to make good aquifers?

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