CHAPTER 10: SURFACE PROCESSES – THE ICING ON THE CAKE

In this chapter you will learn about some of the processes that recently shaped Earth's surface through weathering and erosion. A big part of this in New England is glaciation and past climate change.

(Note: Terms in red and italics appear as entries in the companion glossary.)

Surface processes are important in not only controlling the shapes of modern landscapes but also ancient ones; consider for example how past erosion events created unconformities. The modern landscape, which is above sea level in most places, is an erosion surface that is currently forming, and it can represent an enormous gap in time between the present and the ages of rock formations exposed at Earth's surface. The rates at which surface processes influence the shape of landscapes is a balance between two factors: 1) the types of processes acting on the land surface, and 2) the structure and composition of material, also known as the framework, making up the rocks and surficial deposits of the landscape. In addition, time plays a role in determining how much weathering and erosion will occur. Landforms don't immediately take shape, and it may take thousands, or even, millions of years, i.e., lots of time, for a landscape to take on a form that is reflective of its framework and processes acting on it. There are situations in which certain sets of processes can rapidly erode the land surface and change its shape, but even these rarely occur in the span of a human lifetime. In simple terms, we can summarize the shape of a landscape as:

Landforms = Framework + Processes + Time.

10.1 WEATHERING, EROSION, AND GROUNDWATER

Many rock formations that get exposed at the land surface were originally formed at great depth. Since that time, there may have been a tremendous amount of erosion involved in exposing these rock units at the land surface. We often have little direct evidence, say in the form of remnant deposits or weathered material, that would tell us exactly how weathering and erosion took place at times in the distant past, but surface processes must have acted on ancient landscapes somewhat like they do today, slowly breaking down rocks and eroding the by-products of weathering. To some extent, we can use modern land surface processes to infer what happened, remembering, of course, that surface conditions might have been different, i.e., drier, wetter, colder, warmer, etc. than today. Climate has a huge influence on surface processes since it controls temperature and precipitation. We must also consider that the atmosphere may have changed, having higher or lower amounts of carbon dioxide or oxygen, especially since early Precambrian time.

Water is critical to weathering and erosion of the land surface. Water facilitates chemical decay and is also an important part of physical processes such as *freeze-thaw processes* and *runoff*. In addition to runoff and evaporation, some of the water that lands on Earth's surface as precipitation seeps into the ground. It enters the subsurface through the pores in the soil and along cracks and fractures in rocks as what is called *infiltration*. Infiltrating water may eventually reach the *water table*, the level in the subsurface below which all the fractures and pore spaces in rocks are filled with water. When infiltration reaches the water table, it enters the *groundwater* system by what is called *recharge*. Movement of the water continues in the groundwater system through rocks with

Figure 10.1 *– In the Fells, water seeping from the ground forms springs when it's warm and ice in the winter. A) Ice forming in the winter from seepage of water out of fractures on Boojum Rock. B) Ice forming along Mud Road where high permeability along a fault allows groundwater to seep out of the ground, forming a spring. It is shown here frozen in January 2016. C) Mud Road in March 2016 when it was muddy and wet.*

permeability, and it will eventually come out of the ground as groundwater *discharge*. We see evidence of this today where water seeps out of the ground from fractures, especially along highway road cuts, making the rock surface wet in the summer and leaving spectacular coatings of ice where it freezes in the winter (Fig. 10.1). Water also discharges to lakes and streams as a special type of discharge called *baseflow*. This is sometimes a hard principle for my students to learn: *Streams and wetlands in humid areas such as near Boston are generally places of groundwater discharge, NOT recharge.* Think about it! If we have a drought for 4-5 weeks in the summer, rivers and wetlands in New England generally remain wet. Where does the water come from if there is no runoff? Baseflow! If these water bodies were areas of recharge, there would be no water left in streams under drought conditions. The water would seep into the ground and rivers would get smaller down valley and eventually dry up.

Rainwater carries dissolved oxygen and carbon dioxide, which forms carbonic acid. Acids also form as a by-product of plant decay in soils. These solutions generate chemical reactions with rocks. Where a rock surface reacts with oxygen or acids, it changes the composition of the rock by stripping cations, first on mineral surfaces and then deeper in the rock over time. This weakening of the rock facilitates the propagation of fractures. Chemical reactions form an outer partly-decayed layer of the rock called a *weathering rind*, which is often noticeable by its lighter color (Fig. 10.2). Rainwater seeps into the subsurface where it reacts with rocks along fractures. Weathering occurs first along fractures and slowly advances its way inward away from fracture surfaces forming a pattern in the subsurface of isolated, rounded, non-weathered masses of rock called *corestones*,

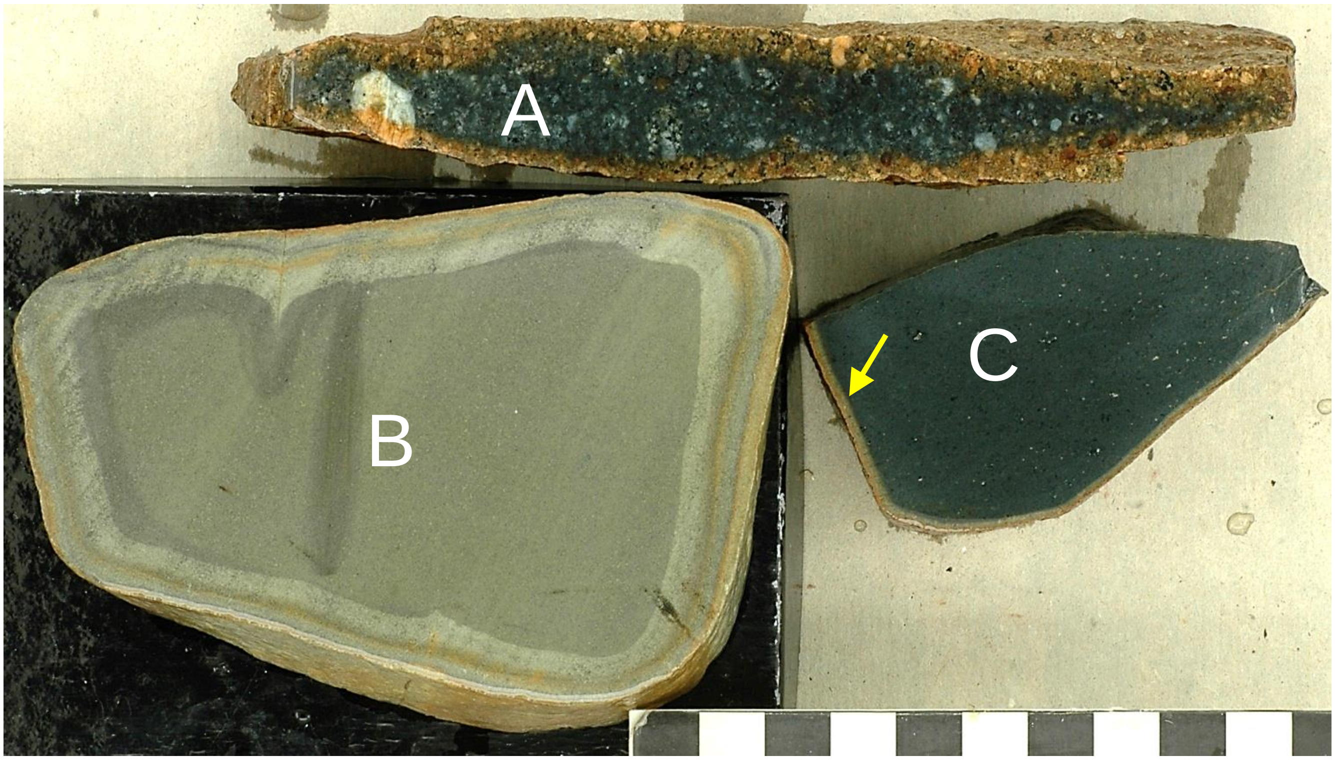

Figure 10.2 *– Weathering rinds formed on the outsides of rocks as a result of chemical alteration (weathering) of rock surfaces. They appear as outer layers that are lighter in color because of removal of cations or rustier due to oxidation. Weathering rinds get thicker over time as weathering penetrates deeper into rocks. A) Orange weathering rind from porphyritic felsite on Mt. Lincoln in the White Mountains of New Hampshire. B) Light gray weathering rind with concentric layers in a calcite-cemented, fine sandstone cobble from a beach in Boston Harbor. The weathering rind has had its calcite removed. C) Thin weathering rind with outer oxide layer (arrow) on basalt from Jenny Jump Mountain in northwestern New Jersey. Scale in cm.*

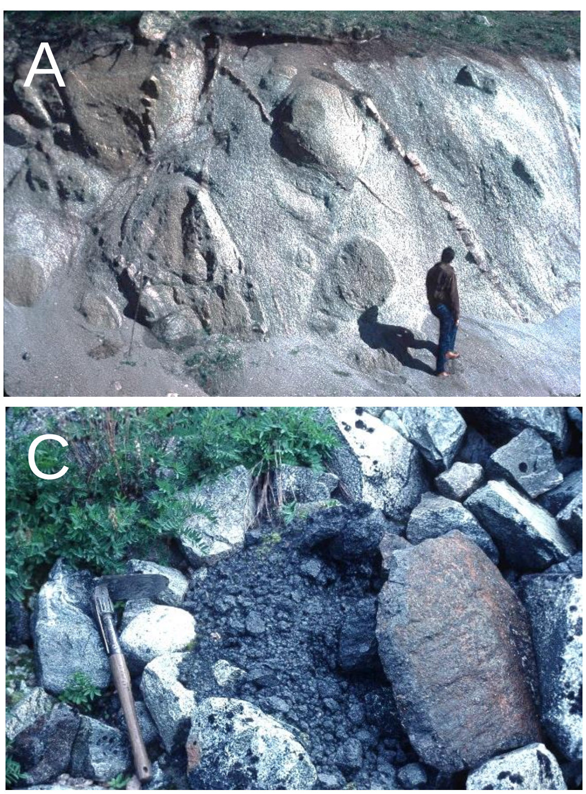

Figure 10.3 *–Grus formation spreads along fractures that isolate rounded areas of solid rock not yet effected by weathering. When these hard areas are exposed by erosion, they appear as rounded corestones. A) Granite in Boulder, Colorado disintegrating to grus with corestones. Note also the fine-grained dike that does not form grus. B) Road cut in granite at Virginia Dale, Colorado with grus and corestones. Erosion of grus has concentrated corestones at the land surface. C) Gabbro boulder disintegrating into grus in central Alaska. Iron oxide has cemented the boulder surface (lower right).*

that are separated by weathered material (Fig. 10.3). In coarse-grained igneous rocks, this process can occur somewhat rapidly. Weathering can occur along the boundaries between mineral grains, which separates them, forming a granular material called *grus*. Erosion will preferentially remove the grus, leaving behind rounded corestones as boulders. Grus and corestones are often easily removed by glacial erosion and the corestones end up being glacially transported (Fig. 10.4).

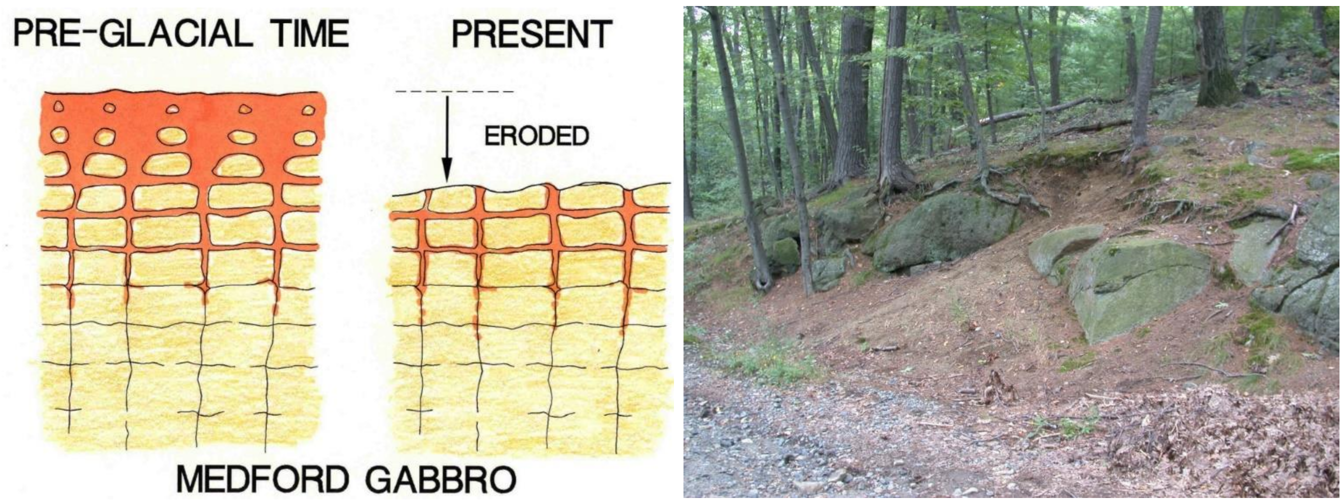

Figure 10.4 *– (left) In preglacial time and between glacial stages, when there was a warm climate like today, the Medford Dike weathered to grus (orange) at the land surface and deep into the subsurface along fractures. Glacial erosion during ice ages removed much of this material and exposed areas of the rock formation that were deeply weathered prior to glaciation. (right) Exposure of the Medford Dike along Tower Road on Pine Hill showing grus between large corestones. Grus has washed out of the fractures and from the soil above and is piled up at the base of the exposure.*

In many places, where weathered material is not removed soon after it forms, a layer of weathered debris builds up at the land surface. Geologists define this weathered layer as a *soil*. This definition of soil is different than the definition used by engineers, who use the term for all loose sediment, surficial deposits, and their weathered zones above the solid bedrock surface. Soils are frequently recognized by their orange color that is produced by the oxidation of iron.

Figure 10.5 - *Soil, the weathered surface of a glacial sand and gravel deposit in the Pequest Valley near Washington, New Jersey. The sand and gravel has a high percentage of limestone pebbles. At the top is a brown layer that is a mixture of organic-rich topsoil and oxidized material where a plow has turned over the upper part of the soil. Beneath that is a yellow to orange layer with a high iron oxide and clay content created by weathering processes. Weathering has removed calcite from this oxidized layer. In the base of the pit is the original sand and gravel deposit where it has not been significantly weathered. Pebbles are clearly visible, and calcite has not been removed by weathering. Shovel for scale.*

10.2 GLACIATION

10.2.1 Glacials and Interglacials

Since the early history of Earth, and continuing today, Earth's climate has experienced extreme cold and warm events. There have been many influences on climate such as the positions of continents relative to the poles, the circulation patterns of ocean currents, and the abundance of greenhouse gases in the atmosphere, especially carbon dioxide and methane. There have also been cold and warm oscillations in response to regular cycles in Earth's orbit that control the amount of solar radiation received at Earth's surface, or *insolation*. Beginning about 40 million years ago, Earth's climate shifted to a cooler average condition, which led to the development of the Antarctic Ice Sheet at the South Pole. Beginning at about 2.5 million years ago, large *ice sheets* began to form in the northern hemisphere. During oscillations to cooler conditions, ice sheets expanded, and during warm intervals they receded. The cool intervals that facilitated the expansion of ice sheets are called *ice ages*, *glaciations,* or *glacials*, and the warm intervals are *interglaciations*, or *interglacials* (Fig. 10.6). During glaciations, large ice sheets formed over North America and northern Europe. In the northeastern U.S., ice sheets expanded as far south as central New Jersey and Pennsylvania, and beyond the southern coast of New England. During warm intervals like today, ice sheets survived in some northern hemisphere arctic areas such as Greenland, Iceland, Svalbard, and islands in northern Canada. Between glacials and interglacials in mid-latitude areas like New England, there was a shift in mean annual temperature of about 20°C.

We have direct evidence in the northeastern U.S. from glacial deposits that glaciers reached this area during what are known as the Wisconsinan (the last ice age; ~90,000 to 11,700 yr ago) and Illinoian (200,000 to 140,000 yr ago) glaciations. In central and eastern North America, major glaciations are named after Midwestern states, where deposits are well represented. There is evidence in Pennsylvania and New Jersey for at least two earlier glaciations of uncertain age, but we are not sure how many additional times prior to the Illinoian Glaciation that ice sheets invaded New England. The record of earlier glaciations is scarce and incomplete because of subsequent weathering and erosion destroying the evidence, which was partly due to glaciation during more recent ice ages. During the last ice age, there was enough water tied up in ice sheets to drop global sea level about 130 m. During ice ages, the land surface along the edge of the ice sheet was populated by mastodons, (see Hyde Park Mastodon and Arlington Mastodon), which went extinct at the end of the Wisconsinan ice age (~12,000 yr ago). Mastodon skeletal remains are not uncommon in the Hudson Valley from when the last ice sheet was receding.

We are currently in a warm interglacial, like the one that separated the Illinoian and Wisconsinan Glaciations. The last major interglacial in North America is known as the Sangamon Interglacial. Interglacials are named after counties in the Midwestern U.S., where soils formed during these warm intervals were first recognized. Fun fact: Sangamon County, Illinois was also the home of Abraham Lincoln. The Sangamon Interglacial was significantly warmer than our current interglacial. Ice sheets were smaller than today, which raised global sea level 6 m above where it stands today. It was also warm enough for alligators to live as far north as Cincinnati, Ohio in the Mississippi and Ohio River valleys. The Southern Quahog clam (*Mercenaria campechiensis*) lived in Boston Harbor, whereas today it only lives as far north as New Jersey. Only the Northern Quahog (*Mercenaria mercenaria*) lives here today, which is famous for the bluish-purple coloration of the interior of its shells.

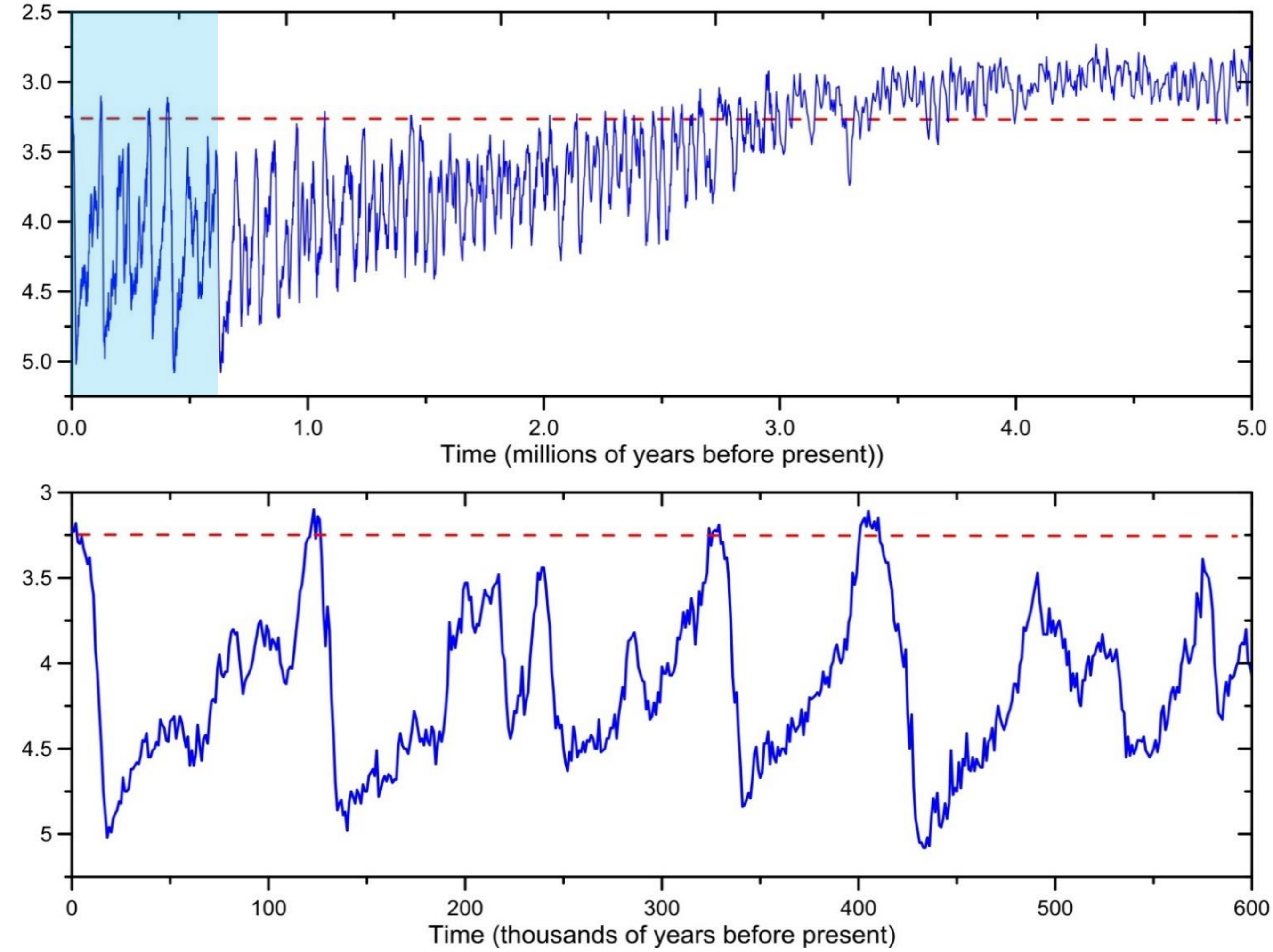

Figure 10.6 *– The pattern of oscillating climate over the last 5 million years (top), which resulted in glacial and interglacial stages in the last 2.5 million years, as depicted by oscillations in oxygen isotope ratios in microfossils from marine sediment cores (data from Lisiecki and Raymo, 2005). Oxygen isotope ratios (vertical axis of graph) vary with global ice volume, with lower numbers corresponding to lower global ice volume. In the lower graph is an expanded plot for the last 600 thousand years for the area shaded in blue on the upper graph. The red dashed lines represent conditions like today. Peaks near the tops of the graphs (lower values) represent warm intervals with low global ice volume. Peaks near the bottom (higher values) represent cold intervals with higher global ice volume. The difference in ice volume from an interglacial like today (red line) to a glaciation like the last one or the one at 430 thousand years ago would stack enough ice on the continents to drop sea level about 130 m. At about 2.5-3.0 million years ago, ice ages began in the northern hemisphere with the growth of the Greenland Ice Sheet and the beginning of the periodic growth of other ice sheets, especially in North America and Europe. It is clear from the graphs that there have been many glaciations in the last 2.5 million years. The last ice age ended about 12,000 years ago with an abrupt change to a warmer climate that continues today. We are currently in an interglacial, much like previous interglacials, except that human burning of fossil fuels and other activities are delivering excess greenhouse gases, especially* CO_2*, to the atmosphere. Fossil fuel burning has elevated carbon dioxide to its highest level since millions of years ago and we should expect global temperatures to rise to levels higher than in any previous interglacial as* CO_2 *and methane levels continue to increase in the atmosphere due to human activity.*

10.2.2 Glacial Erosion

The last two glaciations in the Boston area, especially the most recent Wisconsinan Glaciation, have left very conspicuous evidence in the form of glacial erosion features. Glacial erosion occurs as moving glacial ice removes rock from the surface. This is primarily by *abrasion* of the land surface as a glacier slides across a rock surface. Abrasion occurs when debris, either trapped in the *basal ice* of the glacier or between the ice and the rock surface below, is dragged across the land surface. This produces scratches, known as *striations*, and large grooves where large rocks were dragged across the surface (Fig. 10.7). Movement of the glacier is recorded by the directional trend of striations on the bedrock surface. Therefore, striations serve as very useful indicators for reconstructing the movement direction of an ancient glacier. However, an ice striation is a lineation that only shows a trend, and it is not possible to tell toward which end of the striation the glacier was moving. After all, a scratch is a scratch, with no absolute direction or azimuth. However, some rock surfaces have hard mineral grains or areas that do not abrade as fast as surrounding areas composed of softer materials. These hard areas are often quartz veins, large quartz grains in igneous rocks, or pebbles in a conglomerate that are harder than the finer material surrounding them. When rocks like this are abraded, the hard areas are left standing higher than the more easily eroded adjacent material. This leaves a hard knob and a tail extending behind it in the direction of ice flow, where rock is more protected from abrasion. This is known as a *rattail* (Fig. 10.8; see also Fig. 1.8 in Chapter 1), and it tells us the direction, or *azimuth*, not just the *trend*, of glacier movement. The hard knob is the rat and the protected rock extending behind it is the tail. The rat is always up glacier relative to the tail and was the first thing to meet oncoming ice.

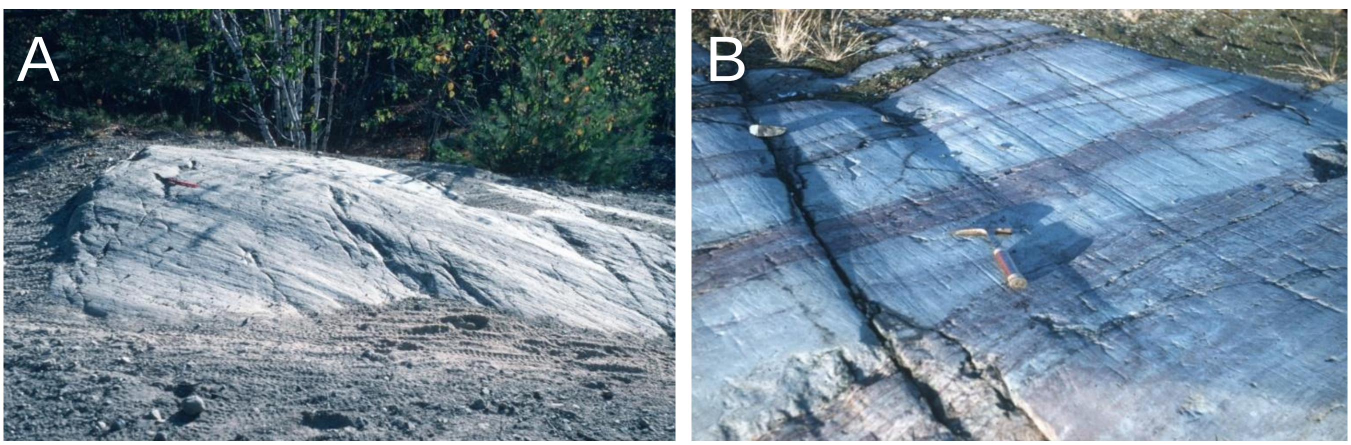

Figure 10.7 *– Glacial abrasion, caused by the grinding of rock debris at the base of a glacier as it slides across a rock surface, creates striated and grooved pavements. A) Highly abraded and streamlined granite surface in the Fells. Ice flow was from right to left. B) Striated sandstone surface in Attleboro, Massachusetts. South is toward the bottom of the image.*

Some rock surfaces are completely smoothed off and covered by striations and grooves as a *glacial pavement*. Boulders, cobbles, and pebbles transported or overrun by the glacier also bear the work of abrasion in the form of striations and flat faces called *facets*, created when they are held in the same position during abrasion (Fig. 10.9). The grinding action of the glacier will pulverize rock debris to a very fine silt size, known as *rock flour*. If debris at the base of a glacier is silt-sized sediment, glacial abrasion will take the form of a *glacial polish*. Glaciers can also remove large blocks of rock by lifting them away in a process called *plucking* or *quarrying*. Occasionally, the glacier will

Figure 10.8 (above) *– Glaciers will abrade underlying rock surfaces as they slide across them, causing erosion. Where the underlying rock is a little harder, and more resistant to abrasion, than surrounding areas, a knob (rat) will form that then protects the area down flow (tail). Together these features are known as rattails. The arrow points to a resistant quartz knob and shows the inferred ice flow direction on an outcrop of metamorphic rock in Kennebunk, Maine. See also Fig. 1.8 in Chapter 1.*

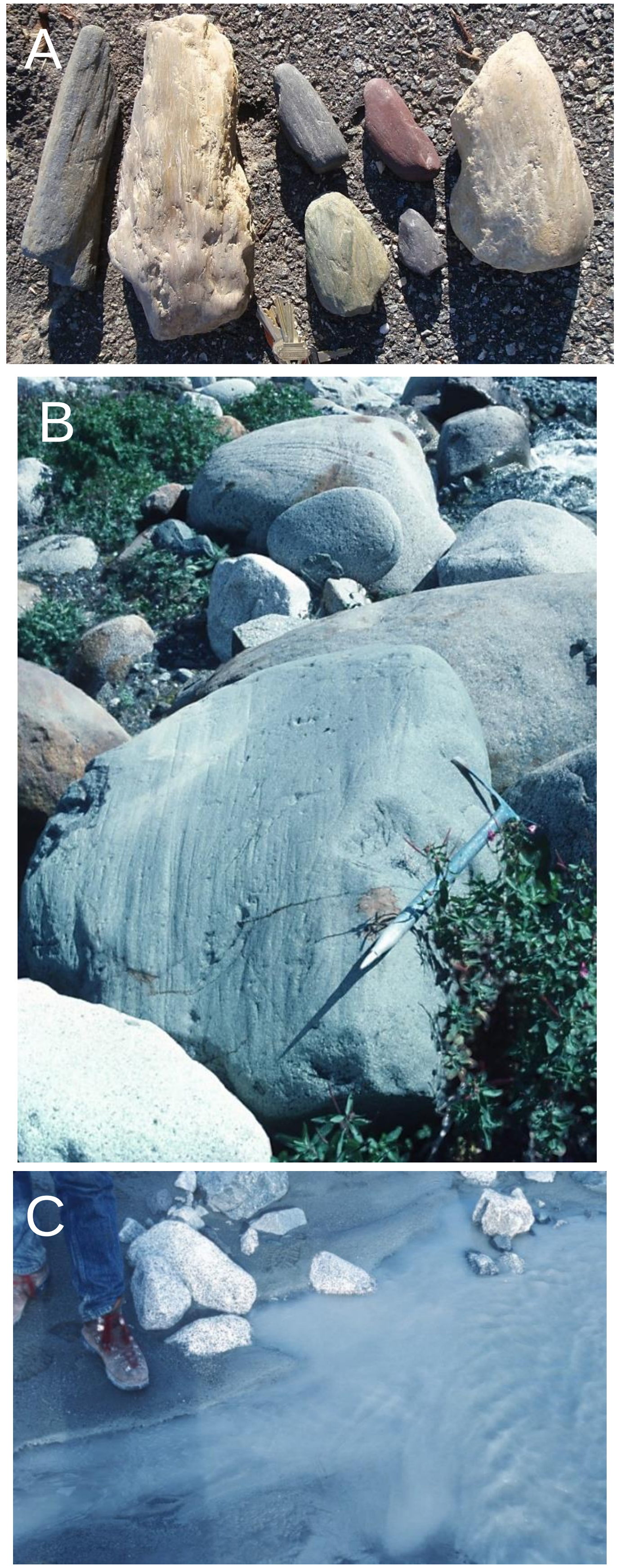

Figure 10.9 (right, top to bottom) *– When rocks are transported at the base of a sliding glacier, they get abraded, forming striated and faceted stones. The grinding action also forms fine silt known as rock flour that is easily observed in streams that drain glaciers. A) Striated stones. Note that the striations are parallel to the long axis dimensions of the stones. B) Large striated and faceted (flat-faced) boulder near the Maclaren Glacier in central Alaska. C) Rock flour in glacial meltwater at the Gulkana Glacier in central Alaska gives the water a milky appearance.*

move a boulder far enough that it is resting on a different rock unit than its source. Boulders or stones of this type are known as *glacial erratics*. To be an erratic, a rock must be different from the bedrock beneath it.

10.2.3 Glacial Deposits

Eroded glacial rock debris, composed of clay-sized particles to boulders, is transported by glacier movement and flowing meltwater. These materials are later deposited to form *glacial deposits* (Fig. 10.10). Beneath a glacier, the forward movement of the glacier drags debris toward the glacier's edges. In some places, the glacier smears the land surface with this debris, or the debris is released from melting ice, to form a glacial deposit known as *till* (Figs. 10.10 and 10.11A). (Note: it is redundant to say "glacial till," since till is defined as a deposit of sediment laid down directly by glacial ice. Since till, by definition, cannot be formed by non-glacial processes, there is no other type. It is always glacial.) In some places, including the Boston area, till is plastered down to form streamlined hills known as *drumlins* (Figs. 10.10 and 10.11B). Drumlins are formed beneath glacial ice at least 500 m thick that is flowing or sliding. Drumlins, like striations, can be used as glacial flow indicators because they are elongate, or streamlined, parallel to the flow direction of the glacier that deposited them.

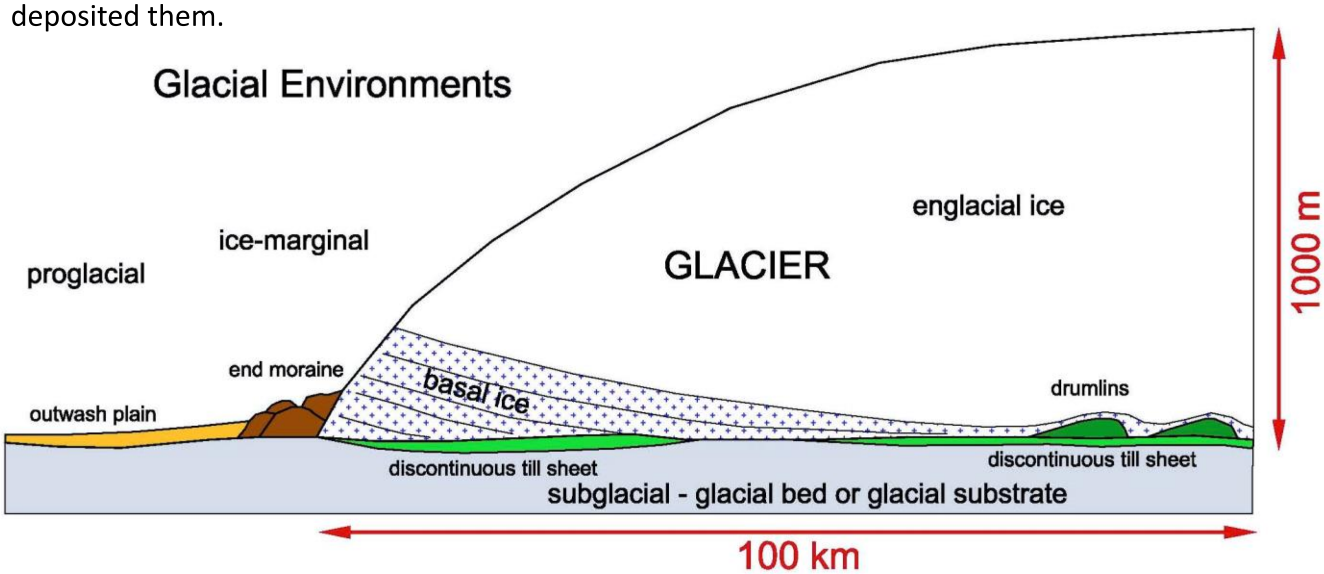

Figure 10.10 *– Glacial deposits of various types shown where they are formed relative to a glacier in a cross-sectional view. Deposits are formed beneath the glacier (subglacial), at the glacier margin (ice-marginal), and in front of the glacier (proglacial). Basal ice is bubble-free, debris-rich ice formed by subglacial water freezing to the base of the glacier. Above the basal ice is englacial ice, which is bubbly with very little debris and formed by the compaction and recrystallization of snow that originally fell on the surface of the glacier. This cross-sectional view is highly exaggerated in the vertical dimension, and if drawn with similar vertical and horizontal scales would be about 50 times thinner than shown here.*

The flowing glacier acts like a conveyor belt: debris being carried to the margin of the glacier melts out of the ice there, or other debris is dragged or pushed by the ice, such that it accumulates as a deposit at the margin. Under the right conditions, when the glacier margin is stable, with its flow rate being equal to its melting rate, or when the margin of the glacier advances and acts like a bulldozer, debris will pile up as a hill at the glacier front known as an *end moraine (*Figs 10.10 and 10.12). End moraines are useful for reconstructing the outline of glacier margins at times in the past because they preserve the glacier's shape. In New England, end moraines tend to be irregular mounds of glacial debris with high concentrations of large boulders.

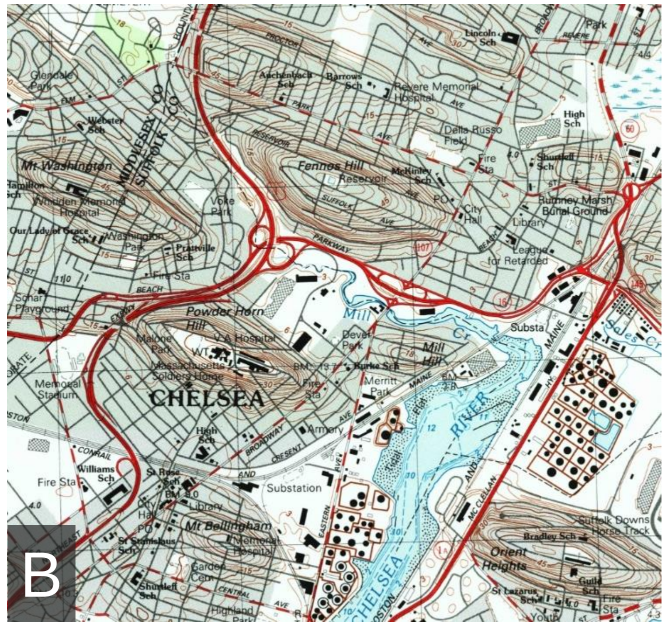

***Figure 10.11 (above and upper right)** – Subglacial deposits formed by the glacier dragging sediment across the land surface are generally classified as till. A) Till that is a poorly sorted mixture of clay to boulders with many striated stones. This till is in Portland, Pennsylvania in an area where the bedrock is limestone. B) Till may be molded beneath a glacier to form hills known as drumlins, which are elongate parallel to ice flow. It takes a thickness of 100's of meters of glacier ice flowing across the land surface to form drumlins. The elongate hills in Chelsea shown on this part of the Boston North Quadrangle are drumlins that are about 50 m higher than the surrounding landscape and depict an ice flow direction about S70°E. These drumlins survived being overrun by the last glacier in this area that moved at S20-30°E and are left over from an earlier glaciation.*

***Figure 10.12 (right)** – End moraines are piles of glacial sediment that accumulate at the margin of a moving glacier. End moraines are either bulldozed or plowed by the advancing front of a glacier, or they are formed when a glacier drags sediment to the glacier margin and drops it off. This end moraine was formed in 1968 by a small readvance of the margin of the Maclaren Glacier in Alaska. By 1984, when this image was taken, the glacier had receded to the right out of the field of view. In the center of the image, the Maclaren River, which flows from right to left, has cut through the end moraine.*

Glacial deposits and features can also be the result of deposition by *glacial meltwater* produced by glaciers. Meltwater is produced mostly by surface melting of a glacier during the summer. At times when a glacier is receding, not only does snowfall from the previous winter melt, but a large volume of older ice also melts, which produces a meltwater volume many times greater than that produced by runoff from annual precipitation. Meltwater can drain off the surface of the glacier, but almost all of it goes down *crevasses* and cracks in the glacier surface. Water that drains into a crack or crevasse in a glacier surface will pile up and build pressure to the point where it exceeds the strength of the ice and can widen the crevasse and open passageways to the base of the glacier. Water arriving at the base of the glacier then flows along the bed of the glacier toward the glacier margin in sheets or channels that generally get larger and better defined near the margin. Basal water erodes sediment at the base of the glacier and transports it to the glacier margin. In some cases, the water flows in tunnels that are carved and melted upward into the ice. These tunnels can become filled with water-transported sediment. When the glacier melts away, the tunnel fillings are left behind on the land surface as winding ridges of water-deposited sand and gravel called *eskers* (Fig. 10.13). When water exits the glacial tunnel system at the glacier margin, it enters river systems or water bodies such as lakes or the ocean. On steep land surfaces, meltwater runs in channels that are widened and deepened (eroded) by the large volume of seasonal water moving away from the glacier. We call these features *meltwater channels*. Today, they appear as large dry channels that have no water source. In larger valleys with gentle gradients, sediment will accumulate in glacial streams and build upward (aggrade), filling the valley with meltwater sediment called *outwash* (Figs. 10.10 and 10.14). When glacial tunnels exit into a lake or the ocean, sediment accumulates at the mouth of the tunnel and produces *ice-contact deltas*.

Figure 10.13 *– An esker is sediment deposited by meltwater in a tunnel at the base of a glacier. It is generally made of sorted sand and gravel. When the glacier melts away, this deposit is left behind on the land surface as a winding ridge of sand and gravel. A) An esker in the Connecticut Valley of western Massachusetts. (Image courtesy of Richard Little of Greenfield Community College.) B) Cross section view of an esker that has been excavated for sand and gravel and is now a lumber yard in the Connecticut Valley of northern Vermont.*

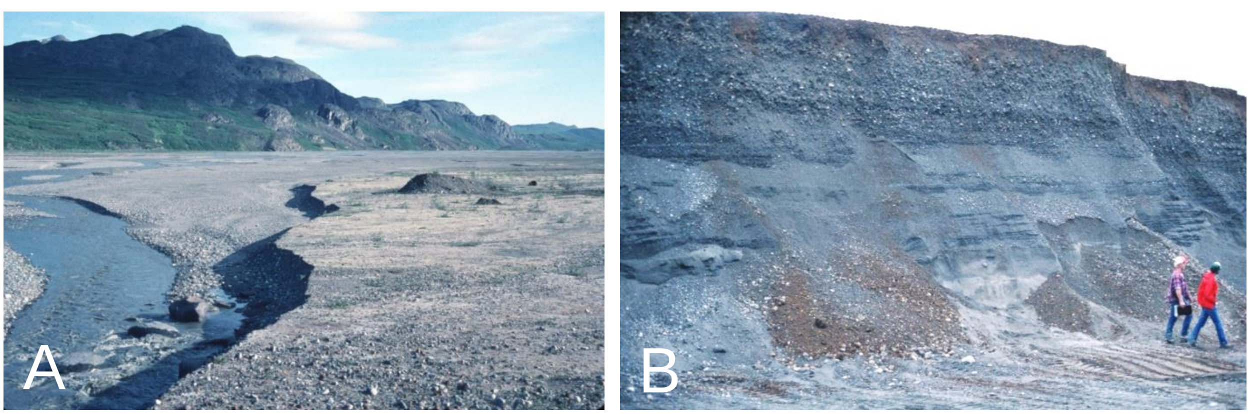

Figure 10.14 *– Outwash is sediment carried out from beneath a glacier by meltwater into a river. A) Outwash plain formed by meltwater from the Maclaren Glacier in Alaska. View away from the glacier. B) Outwash from the last glaciation in a sand and gravel pit near Ogdensburg, New Jersey. This exposure has long, flat beds of gravel. C) Sandy outwash deposit from last glaciation at Poland, New York. This deposit has lens–like bodies of cross-bedded sand called cut and fill structures, which were formed by river channel erosion and bar deposition.*

10.2.4 Recession of an Ice Sheet

Landscapes that were once covered by an ice sheet, such as in the Boston area, must undergo a dramatic transition from being covered by glacial ice more than 1000 m thick to a situation where the ice recedes from the land surface, leaving behind features related to both subglacial erosion and deposition. Subglacial deposits are later superposed by ice-marginal and proglacial deposits, leaving an array of landforms from different glacier environments in one place. In some places, this transition occurs rapidly, while in other areas it may be slower.

For students, a sometimes-confusing aspect of the process of glacier recession is what happens to ice flow, or the forward movement of the glacier. Glaciers always flow, or move, from places where the surface of the glacier has a high elevation to areas with low surface elevation at the glacier's margin. When the glacier recedes, it *NEVER* turns around and flows back to where it came from because this would be flow from a low surface elevation to a high glacier surface elevation. So, what exactly happens when glaciers disappear? What triggers the formation and advance of a glacier in the first place is a cooling climate that allows glacier ice to survive at lower elevations. In this case, accumulation of snow on a glacier exceeds the melting of ice near its margin, and the glacier expands upward and outward, pushing ice away from its center. When a glacier recedes, melting exceeds the forward flow rate of the glacier, and the margin of the glacier is often said to retreat. The glacier always continues to move (flow or slide) toward its margin. It does not reverse its flow direction. Therefore, I like to say a glacier ***recedes***, rather than retreats. The margin of the glacier may appear to retreat, but the ice always moves forward toward the glacier's margin.

10.3 POSTGLACIAL LANDSCAPE PROCESSES

With warming climate at the end of the last glaciation, when the ice sheet in New England began to melt more rapidly, the climate was still colder than today. It was too cold to support the dense forest vegetation that covers the New England landscape today. Under the early postglacial climate regime with its sparse vegetation, loose glacial sediment cover, and cold temperature (*periglacial*) processes, erosion was very fast and many hill slope glacial deposits were eroded into adjacent valleys. Sparse vegetation meant that runoff from rain or snow melt would be able to more easily cut rills and gullies on slopes (Fig. 10.15). Also, cold conditions likely facilitated the expansion of fractures by the deep freezing of water. This separated already-fractured rocks and loosened rock and soil debris that could then make its way down slope. Much of this erosion may have been by *mass movement*, which is the down slope movement of material under the influence of gravity without a fluid transporting it. Mass movement deposits are generally known as *colluvium*. One form of colluvium called *talus* forms from angular rock debris that falls to the bottom of cliff faces and rests at a steep angle.

Postglacial climate conditions in New England may have been cold enough to trigger permafrost, in which some part of the subsurface remained frozen throughout the year. Permafrost soils drain poorly when surface ice thaws in the spring and there is still ice below. This causes the upper soil to become saturated, and under these conditions on a slope, it is relatively easy for mass movement to occur. Even if permafrost did not occur in eastern Massachusetts shortly after the ice sheet receded, deep freezing each winter would have caused frost heaving, and when the soil thawed in the summer, down slope movement of the weakened soil would have been rapid. The cold climate also allowed the wind to erode sediment on a landscape protected by only sparse vegetation. Silt and fine sand deposited by the wind in some areas are now mixed with other mineral material and decaying vegetation in the upper parts of soils.

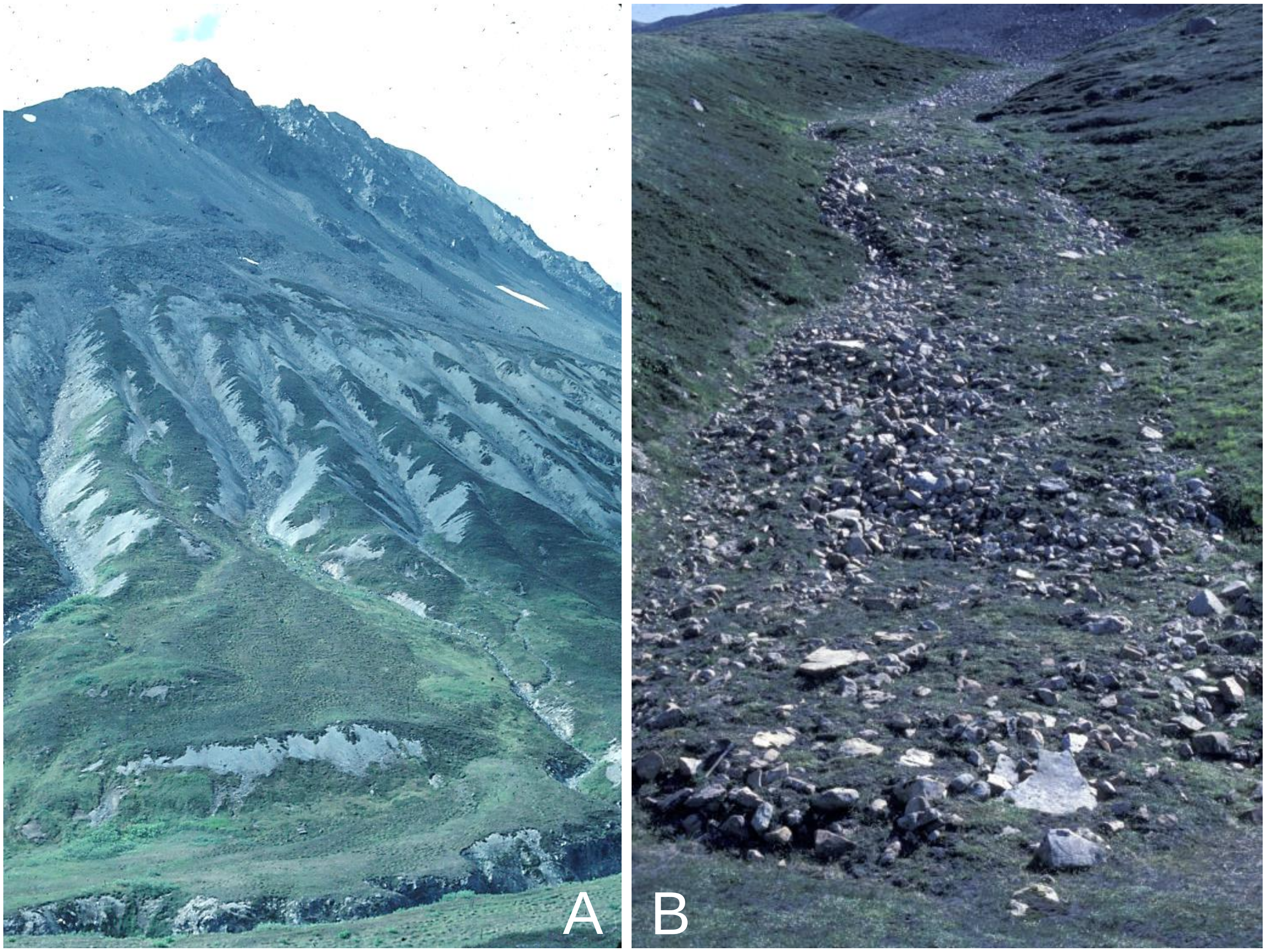

Figure 10.15 *– A) Gullies formed in till of the last ice age on a steep slope in the Maclaren River Valley of central Alaska. After glacial ice receded thousands of years ago, and up to the present, this hill slope has been sparsely vegetated, which has allowed surface runoff and frost processes to cut gullies in the till. B) When till freezes and then thaws it can fail and generate muddy debris flows, which deliver sediment to the base of a slope, in this case adjacent to the Canwell Glacier in central Alaska. The debris flow has been funneled down a gully.*

In addition to landscape erosion triggered by cold climatic conditions, other features and processes related to running water also operated since the last glacier receded. These include erosion by runoff and the development of streams. With sparse vegetation on the land surface, runoff during heavy rainstorms could cause small-scale flash flooding. Deposits of coarse stream sediment may pile up at the lower end of a steep valley. This forms an *alluvial fan*, which is a gently sloping apron of stream sediment. Deposition of this sediment is triggered by changes in the shape of a valley, where it goes from a narrow, restricted channel to a broad, open plain, as is the case where small hillside streams enter a main valley.

The development of *wetlands* is another postglacial phenomenon. As soon as the last glacier melted, there were likely ponds in some areas. Wetlands also formed on the poorly-drained, flat, and low gradient floors of valleys and in meltwater channels that no longer carried water after the ice receded. Ponds and wetlands didn't have the dense vegetation surrounding them that occurs naturally today. Today, many of the early postglacial wetlands have flat, vegetated surfaces where ponds have become filled with organic sediment over the last 10,000 years. In some cases, valleys may have been impounded by beavers, which were abundant prior to the arrival of European colonists. The filling of ponds with plant debris, such as dead aquatic vegetation and leaf litter that was carried into the ponds, was greatly facilitated by vegetation on the surrounding land surface. With plant cover on the land, surface nutrients were washed into the ponds, and aquatic plants and other organisms became abundant. Both high productivity in the ponds and leaves carried into the ponds by runoff and the wind helped to fill the ponds with organic sediment.

Following glaciation, the land surface became inhabited by plants that occurred as a succession of plant communities that adapted to changes in climate (Peteet and others, 1994; Newby and others, 2000; Oswald and others, 2018). The first vegetation following the retreat of glacier ice was sparse vegetation composed of tiny herbs, shrubs, and grasses like we see today in arctic *tundra* environments. Some areas of southern New England had permafrost, an area where some part of the ground in the subsurface remains frozen through the summer (Fig. 10.16). Following glaciation, *lichens* (Fig. 10.17) also began to grow on rock surfaces, and they continue to be abundant today. Lichens assist with the weathering of the rocks to which they cling. Today, lichens can be frustratingly thick and widespread on rock surfaces, and they can make it difficult for a geologist to identify a rock type or observe structures in a rock.

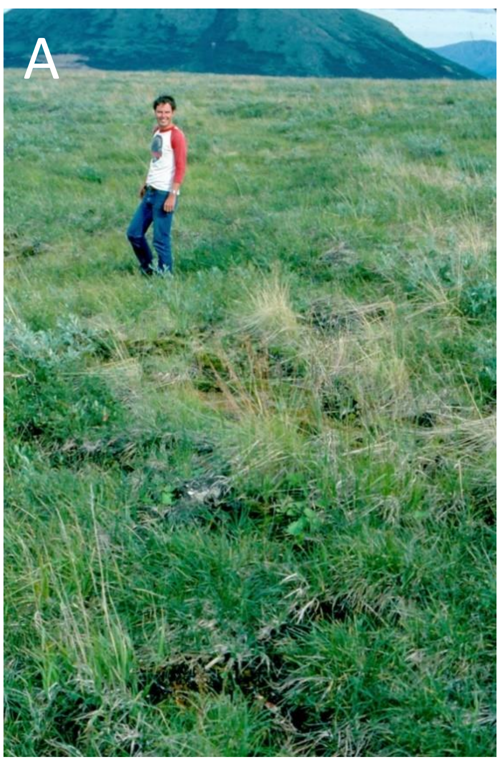

Figure 10.16 *–Tundra vegetation in central Alaska much like the grass, herb, and shrub landscape that covered southern New England immediately following recession of the last glacier. A) Tussocks of grass growing over permafrost in central Alaska along the Richardson Highway. B) Tundra in an area of thawing permafrost where melt ponds (thermokarst lakes) have developed. Note that trees are scarce in areas of permafrost.*

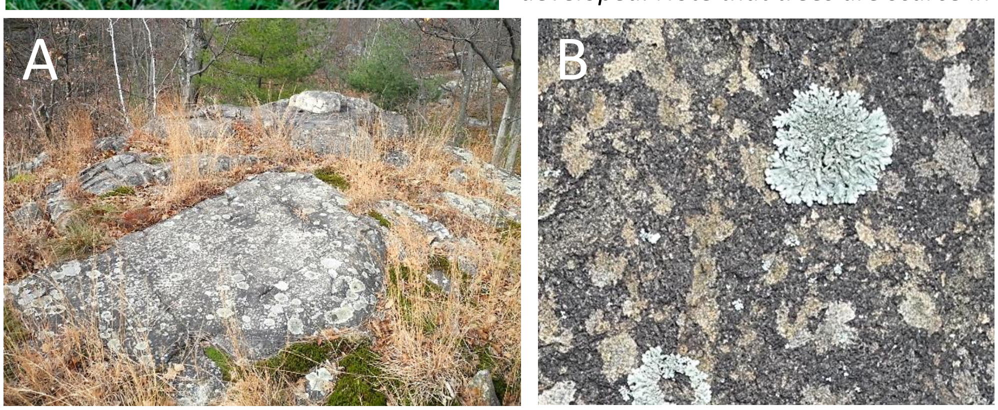

Figure 10.17 *– Circular lichens on rock surfaces in the Fells (A&B). The circular lichen in image B is about 4 cm. Lichens are the symbiotic growth of algae and fungus and are common on rock surfaces and tree trunks.*

In southern New England, tundra vegetation lasted until about 14,000-15,000 yr BP and then gave way to the first forests, which were dominated by spruce trees (Peteet and others, 1994; Newby and others, 2000; Oswald and others, 2018; Fig. 10.18). The spruce-dominated *coniferous* forest lasted until about 12,000 yr BP, when it started to give way to forests with pine trees. About 11,000-9000 yr BP, the first *deciduous*, or hardwood, forest developed, which in the Boston area was dominated by oak, birch, beech, and maple trees (Fig. 10.19). The hardwood forest dominated the land surface up until the time that the first European settlers arrived.

Figure 10.18 *– A spruce-dominated forest interspersed with permafrost areas of grass, wetlands, herbs, and shrubs along the Richardson Highway in central Alaska. The Alaska Pipeline runs across this small basin near Delta Junction, Alaska. This type of vegetation probably began to populate southern New England at 15,000-14,000 yr BP following a period of tundra vegetation. With the dramatic warming of climate after 12,000 years ago in New England, this vegetation regime was replaced by pine, and then later deciduous trees.*

The arrival of Europeans marked a dramatic change in the landscape (Cronon, 1983; Foster, 1995; Richburg and Patterson, 2005). Cutting of the forests by early settlers changed the dominant vegetation in eastern New England and most other areas of the northeastern U.S. Settlers needed wood for fuel and building material, and they needed to clear space for farming and grazing animals. Soon after their arrival, the colonists clear-cut much of the forest surrounding Boston. This led to runoff erosion of soils that were no longer protected by dense forest vegetation. Accelerated upland and slope erosion triggered the filling of stream valleys with flood plain sediment and deposition of small alluvial fans. Clear-cutting of the forest to make grazing land for livestock had the effect of allowing the invasion of many types of grasses and weeds and pioneering tree species. Along with frequent fires, this has left exposed bare rock on hilltops in the Fells. I would love to say that I regret that this happened, but it has made a detailed investigation of the bedrock in the Fells possible.

***Figure 10.19** – The forest that covers most of the Fells today is dominated by deciduous trees. Mixed with the deciduous trees are pines and hemlocks (conifers). This is different than the natural deciduous forest that covered New England 1000's of years ago. Instead, it is a forest that has grown back in many areas from clear-cutting and frequent fires. Forest adjacent to Bellevue Pond near Pine Hill.*

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