CHAPTER 7
GLACIERS

1. INTRODUCTION

1.1 Before the era of universal air travel, which commenced less than half a century ago, few of the world's population had seen a glacier. I suspect that majority of class members in this course have seen a glacier—if not close up, then out of a jetliner window. In the Canadian Rockies, you can drive to within almost a stone's throw of the terminus of the Athabasca Glacier, a classic active valley glacier. In many other parts of the world, valley glaciers are accessible to even casual day hikers. The great ice sheets of the world, in Antarctica and Greenland, remain much less accessible.

1.2 In the broad context of geologic history, the Earth is in an "icehouse" time, with recurrent major ice-sheet advances across the Northern Hemisphere continents. (There have been several other such icehouse periods in Earth history, separated by long intervals of ice-free times, called "greenhouse" periods, with no evidence of glaciation.) The Earth has only recently emerged from the latest episode of continental glaciation. Does it surprise you to learn that a mere twenty thousand years ago the Boston area was beneath a mile of glacier ice moving slowly southward toward its terminus south of what is now the south coast of New England?

1.3 The Earth is in many senses a glacial planet:

- 10% of the Earth is covered with glacier ice (about 15 million square kilometers).
- About 40% of the Northern Hemisphere in winter is covered with solid water at any given time (land and sea).
- 75% of the Earth's fresh water is in glaciers.
- Surficial deposits by glaciers cover a large percentage of the Earth's land surface.
- Glaciers have a profound effect on the Earth's climate (as well as being in turn controlled by climate).

1.4 And glaciers are not without practical importance:

- Role of ice sheets on climate
- Control on sea level

- Potential source of fresh water

1.5 Disciplines: in the US, people who deal with glaciers or their products are usually allied to *geology*; in other English-speaking countries, usually to *geography*. Disciplines relevant to this chapter:

glaciology
glacial hydrology
glacial geology
geomorphology
sedimentology
stratigraphy

1.6 What is a ***glacier***?: *a body of ice and recrystallized snow (plus refrozen meltwater), on land (or, if floating, then connected to land) and moving by deformation under its own weight.*

1.7 Why are there glaciers? A necessary and sufficient condition is an excess of snowfall over snowmelt at some locality for a long enough time to build up ice thick enough for it to flow under its own weight.

2. CLASSIFICATION OF GLACIERS

2.1 One good way of classifying glaciers is by the extent to which their shape and movement are affected by the underlying bedrock topography.

Glaciers not strongly constrained by underlying topography:

Ice sheet:

Superimposed on underlying topography; largely or entirely submerges that topography; ice flow reflects largely the size and shape of the glacier, less the shape of the ground. Ice sheets smaller than about 50,000 km^2 in area are called ***ice caps***.

Ice dome:

About symmetrically over land area involved; the top of the dome may be over bedrock highs or bedrock lows. Highest: 4300 m, Antarctica. The convex profile is a reflection mainly of flow mechanics, including the nature of the bottom roughness.

Outlet glacier:

A stream of ice that extends beyond an ice dome and drains it. Outlet glaciers are often very large: several hundred kilometers long and tens of kilometers wide. Commonly they are commonly associated both with large ice sheets and with smaller ice caps. Most of the Greenland ice sheet and about three-quarters of the Antarctic ice sheet is drained in this way.

Ice shelf:

A floating ice sheet that deforms under its own weight. The slab of ice is squeezed between the atmosphere and the ocean. The ice shelf has to be anchored at several points. In Antarctica, ice shelves constitute about 7% of ice area. The ice shelf may be fed partly or not at all by land glaciers. Accumulation: snow on the surface; land glaciers; bottom freezing. Ablation: melting (at the top or the bottom); calving.

Glaciers strongly constrained by topography:

***Ice field*:**

An approximately level area of ice, not an ice cap because not domelike; the flow reflects the underlying bedrock topography (ice fields are usually nestled in among mountains). Size: a few square kilometers to very large. Ice fields grade over into small ice caps. Requirement for existence of an ice field: high and overall gentle topography.

***Valley glacier*:**

A glacier flowing in a rock valley and surrounded by rock walls (therefore long and linear). Fed from an ice field or a cirque (see below). Usually 10–30 km long, but up to 100 km. The terminus can be on land or in the sea (by calving). A valley glacier can also disgorge from mountains and spread out as a flat mass in piedmont areas to form a what is called a ***piedmont glacier***. A good example of a piedmont glacier: the Malaspina Glacier, in Alaska, 40 km across. Valley glaciers are usually vigorous glaciers. And because they are typically present at low as well as high latitudes, they are among the most accessible of glaciers.

***Cirque glacier*:**

A small ice mass, fairly wide relative to its length, occupying a bedrock hollow or basin, usually on a mountain slope.

3. DISTRIBUTION OF GLACIERS

3.1 The total area of the Earth's surface covered by glaciers is 14.9 million km^2

- Antarctic ice sheet: 12.5 million km^2
- Greenland ice sheet: 1.7 million km^2
- All the rest: 700,000 km^2 (many ice caps, mostly less than about 10,000 km^2; many thousands of small glaciers, mostly valley glaciers)

3.1 Aside from the Greenland ice sheet, most of the larger glaciers in the Northern Hemisphere are mostly on Iceland and the Arctic Islands of Canada, because of the distribution of land and sea.

3.2 It's a lot more difficult to get the *volumes* of ice in glaciers than to get the *areas* covered by glaciers. Things have gotten better in this respect, though, because of use of echo sounding, similar to oceanographic depth sounding. But one still has to go out there to the glacier to do it.

Antarctica: 21.5 million km^3 (sea-level equivalent about 60 m)

Greenland: 2.4 million km^3 (sea-level equivalent about 5 m)

All other: 180,000 km^3 (largest ice cap: Vatnajökull, Iceland, 3100 km^3, and fairly accessible).

4. GLACIER ICE

4.1 The transition of new snow to glacier ice is similar to the deposition, diagenesis, and metamorphism of a sediment to form a metamorphic rock. (And in a real sense, glacier ice *is* a metamorphic rock.)

4.2 New dry snow has low bulk density and high porosity, and enormous internal surface area. Crushing, compaction, and exchange of water between flakes and air at low temperatures by sublimation and deposition, and at temperatures near melting by melting, refreezing, evaporation, and condensation, tend to round grains, producing the nearly spherical grains of granular snow. Except in the coldest environments, this takes a few days to a few weeks. In the process, the snow settles and gets more compact, and small grains disappear at the expense of larger ones. The recrystallized granular snow, still friable and porous, is called ***firn*** when it is more than a year old. The firn is then converted to glacier ice as compaction due to increasing weight of overlying younger snow continues. Air is forced out, decreasing the intergranular space. Recrystallization proceeds by solution at points of contact and deposition in the interspaces.

4.3 When the permeability to air has reached almost zero, the firn has become glacier ice. This change occurs in a relatively narrow density range around 0.82–0.84 g/cm^3, at depths of greater than 30 m. Crystal size has increased to about 1 cm. Firn is transformed into glacier ice in one year to a few hundred years, depending on temperature and rate of accumulation. Then, depending on both temperature and bottom slope, something like 50 m of ice is needed for the ice to flow plastically under its own weight (there's more detail on flow of glacier ice in a later section). The ultimate density of glacier ice approaches 0.90 g/cm^3, a bit short of the density of 0.917 g/cm^3 of pure ice because of the inevitable residual presence of little air bubbles.

5. THE BUDGET OF GLACIERS

5.1 Economy vs. Regimen

5.1.1 First off, the term ***accumulation*** applies to all of the ways *glacier ice mass is added to a glacier*, and the term ***ablation*** applies to all of the ways *glacier ice is removed from a glacier*.

5.1.2 The term ***economy*** refers to the relative magnitude of accumulation and ablation. A glacier with accumulation greater than ablation over some period of time (much longer than just a single year) is said to have a *positive economy*, and gains ice volume with time. A glacier with accumulation less than ablation is said to have a *negative economy*, and loses ice volume with time. A glacier with a positive economy not only thickens but also extends its terminus in the downglacier direction. A glacier with a negative economy becomes thinner, and the terminus retreats upglacier.

5.1.3 The term ***regimen*** refers to the absolute values or magnitudes of accumulation and ablation, irrespective of their balance. A glacier with large values of accumulation and ablation is is said to have an *active regimen*, whereas a glacier with small values of both accumulation and ablation is said to have an *inactive regimen*. So there are four different combinations of economy and regimen.

5.2 Glacier Bookkeeping

5.2.1 Ice accumulates on a glacier by a variety of processes: snow (plus sleet and other solid precipitation), rain that gets frozen onto or into glacier, rime, frost, and avalanches. Snow is by far the most important of these.

5.2.2 Modes of ablation are melting (plus runoff and evaporation), sublimation, deflation by wind, and calving. The important ones are *melting* and, for glaciers that terminate in water, *calving*.

5.2.3 Occupy an observation point that's fixed relative to the bedrock base or walls of the glacier, and consider any point on a the surface of the glacier. Look at the vertical thickness of water (i.e., the height equivalent to solid ice) added to or lost at a point.

5.2.4 In glacier bookkeeping, an important concept is the ***balance year***: *the time between two successive dates with minimum mass of glacier ice at a given location* on the glacier. Note: balance years are not necessarily 365 days, because the time of minimum ice mass depends upon the weather throughout the balance year, and the balance year can differ from point to point on the glacier.

5.2.5 Imagine plotting a graph showing the vertical ice-equivalent height added or subtracted as a function of time for a whole balance year (Figure 7-1). (Does the concept of "vertical ice-equivalent height" make sense to you? You have to convert accumulation and ablation into ice thickness.)

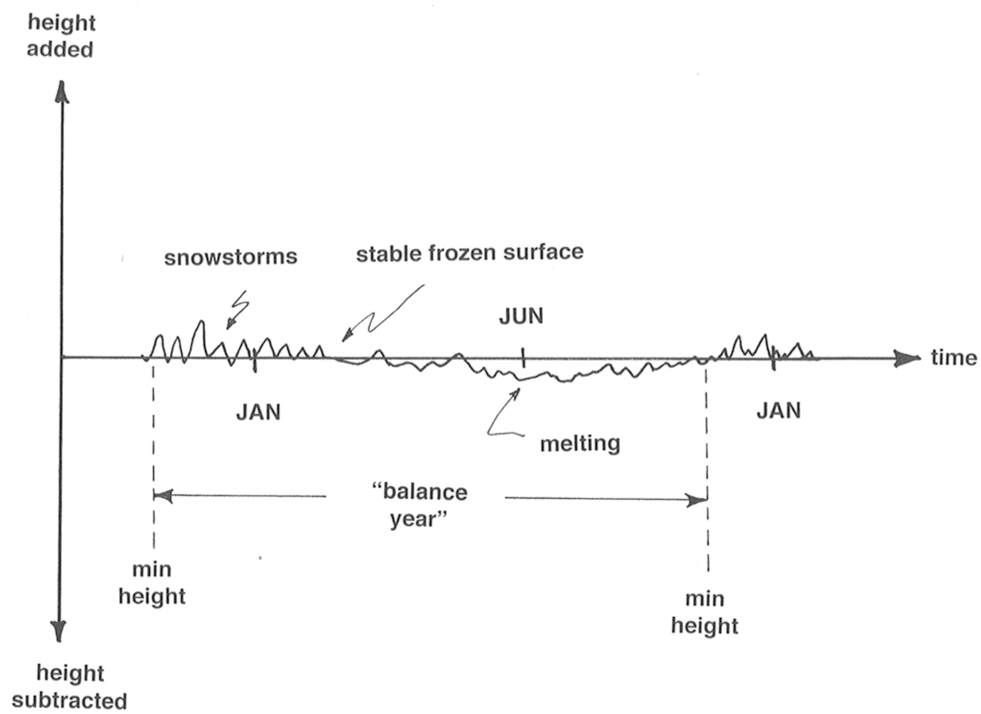


Figure 7-1. Keeping track of accumulation and ablation as a function of time on a glacier, for one balance year.

5.2.6 Now split the graph in Figure 1 into positive and negative parts (i.e., monotonically nonincreasing and monotonically nondecreasing), and then integrate (that is, keep track of a "running sum") the two parts separately to get an *accumulation curve* and an *ablation curve* (Figure 7-2). The difference between the accumulation curve and the ablation curve is called the ***mass-balance curve***. This curve is sort of the *net change as you go along* through the balance year. Don't worry that it's mostly above the horizontal axis; that's only because we chose to work on the basis of minimum-height times as the starting and ending points of the balance year. The difference in elevation of the mass-balance curve and the horizontal axis at the end of the balance year is the *net balance over the balance year*, positive at this particular place and in this particular year.

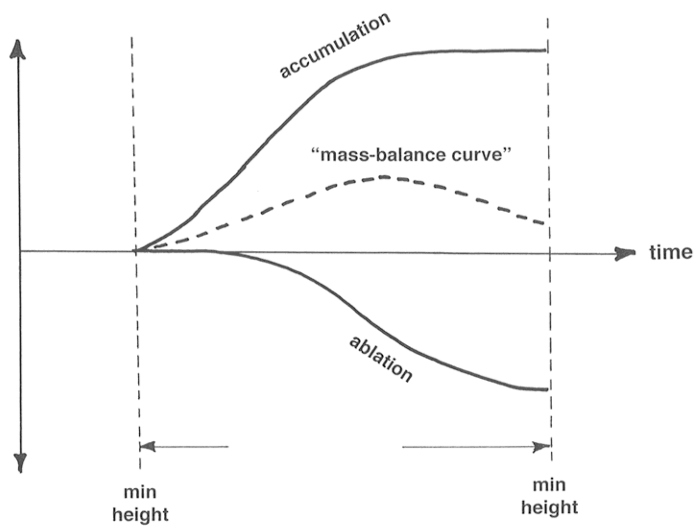


Figure 7-2. Vertical ice-equivalent height added to or subtracted from the surface of a glacier during a single balance year.

BACKGROUND: INTEGRATION AND INTEGRALS

1. This is not the place for a full-scale calculus lesson, but for those of you who have not been exposed to the beauties of differential and integral calculus, here's the concept of integration and integrals.

2. First, a few words about mathematical functions. You can think of a function as a magic box: you put a number into the box, and out comes a value that's associated with that number. You're probably more used to thinking of a function in strictly mathematical terms, in the form of an equation like $y = x^2$: plug in a value for x, and the function gives you the value of y.

3. In mathematics, the process of integration involves a summation of the values of some function, as the value of the function varies over its range. If the function is defined only over a range of discrete values of the input variable (1, 2, 3, ... , let's say), then the summation is simple and straightforward. All of you have done that. But how about when the function varies continuously over all values in the input variable, as, for example, the function $y = x^2$ above? There's a big conceptual leap involved here. The mathematical process by which the value of the function is summed continuously is called ***integration***, and the value of the sum thus obtained is called an ***integral***.

4. Obviously this is not the place either to develop the concept with mathematical rigor or to prescribe how to actually do integration. Here's a simple example to show the results. Figure 3 shows the results of integrating the function $y = x^2$ from $x = 0$ to $x = 2$. The curve in Figure 3 is that of the function $y = x^2$, and the shaded area under the curve between $x = 0$ and $x = 2$ represents the value of the integral, which turns out to be exactly 2/3. Your common sense tells you it has to be a bit less than 1, which is what it would be if the function were the straight line $y = x/2$.

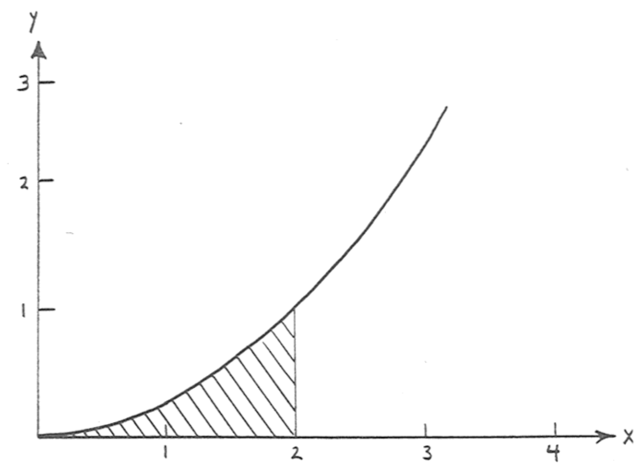


Figure 7-3. Integrating the function $y = x^2/4$ from $x = 0$ to $x = 2$.

5. If somebody challenged you to find such an integral, without actually knowing how to do it mathematically, your clever mind might hit on the strategy of dividing up the x axis into a great many little segments and drawing rectangles with the value of y at the their tops, then figuring out the areas of all of the litle rectangles, and adding up all of those areas (Figure 7-4). That would be a good approximation to the integral you want to find. If you're even sharper than that in such matters, you could imagine making the rectangles skinnier and skinnier, without limit. At that point you would be very close to the heart of the mathematical concept of integration!

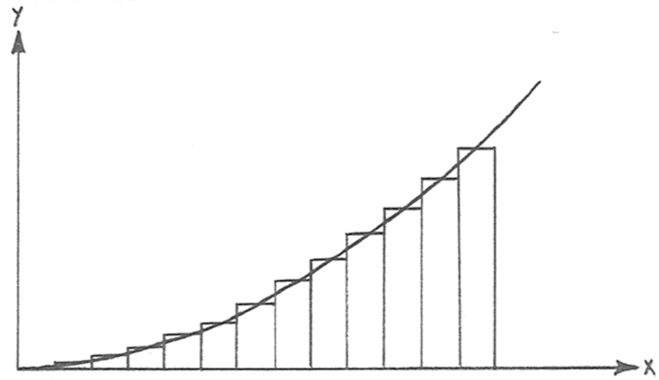


Figure 7-4. How you can evaluate the integral by approximating it with a lot of little rectangles that follow the curve.

5.2.7 The curves in Figures 7-1 and 7-2 differ from year to year at a point. Also, obviously they differ greatly from point to point: at points high on the glacier there's a strongly positive net balance, maybe with no ablation all year, whereas low on the glacier there's likely to be a strongly negative balance, with not much accumulation (presumably there's always some) but lots of ablation.

5.2.8 Figure 7-5 is a plot of ice-equivalent height gained by accumulation and lost by ablation as a function of elevation on a given glacier. In the right-hand part of the graph, high on the glacier, accumulation is greater than ablation, whereas in the left-hand part of the graph, low on the glacier, ablation is greater than accumulation. At a certain elevation the two curves intersect; this elevation is called the ***equilibrium line***. Upglacier there's still some of last year's solid water (firn and glacier ice) left when new accumulation begins at the start of the new budget year; downglacier, all of last year's snow melted before new accumulation starts. (Don't worry about the areas under the curves; they don't have to be equal, because they depend on the distribution of glacier area with elevation.)

5.2.9 Another thing you could do is integrate the accumulation and ablation curves in Figure 2 over the entire surface of the glacier to get the total accumulation and ablation. The result would look something like Figure 7-6. A little thought should convince you that glacier mass is unchanged through a given balance year if and only if Area 1 equals Area 2. If $A_1 < A_2$, the balance for that year is negative, and the glacier loses mass; if $A_1 > A_2$, the balance for that year is positive, and the glacier gains mass.

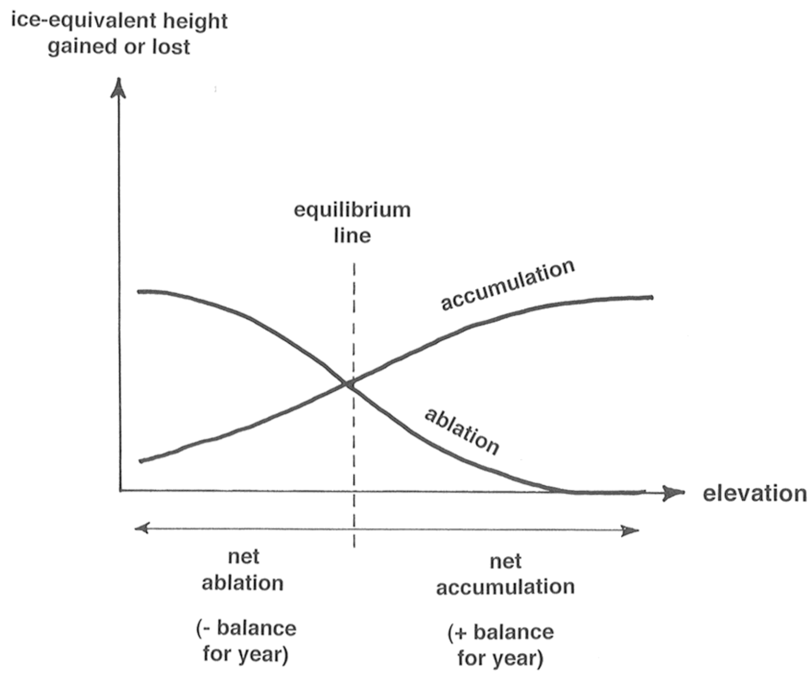


Figure 7-5. Integrating the positive and negative parts of Figure 1 separately to obtain an accumulation curve and an ablation curve.

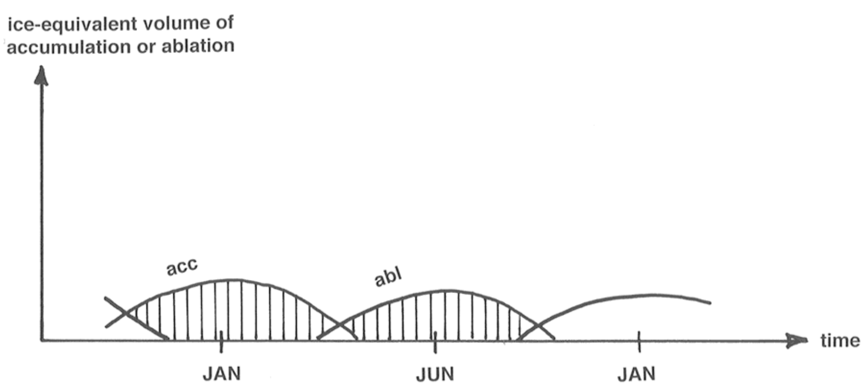


Figure 7-6. Ice-equivalent height gained by accumulation and lost by ablation as a function of elevation on a glacier.

5.2.10 It is very difficult to obtain accumulation and ablation curves at points on a glacier. It's easiest to measure (i.e., estimate) the net balance at many points in both the accumulation area and the ablation area and then integrate over the area of the glacier to find the net gain or loss of glacier mass for the given balance year. How? Dig pits down to last year's snow in the accumulation area and then measure ablation at fixed stakes in the ablation area. (There's some distortion involved, because of glacier movement.) Keeping track of accumulation and ablation as a function of time, as in Figure 7-4, is very time-consuming.

ADVANCED TOPIC: THE SITUATION AROUND THE EQUILIBRIUM LINE

Features and relationships near the glacier surface around the equilibrium line are more complicated than I've indicated so far. To see this, imagine a survey from head to foot of the glacier at the end of summer melting season, to look at near-surface materials (Figure 7-7). Here are some comments on the various zones labeled in Figure 7-7.

dry snow zone: no melting even in summer.

percolation zone: some surface melting in summer. Water percolates down into snow at 0°C and refreezes, thereby warming the surrounding snow. Two characteristic forms of ice are deposited: ***ice layers*** *or* ***ice lenses***, formed when water spreads out at some relatively impermeable horizon, and ***ice pipes*** or ***ice glands*****,** formed by freezing of vertical water channels. The depth of saturation increases downglacier to the saturation line—the point where by the end of summer all of the snow deposited since the end of the previous summer has been saturated.

soaked zone: by the end of the summer, all the current year's snow has been saturated and raised to 0°C. Some meltwater also percolates into deeper layers.

superimposed-ice zone: in the lower part of the soaked zone, at lower elevations, there's so much meltwater that ice layers and patches among the firn merge to form a continuous mass of ice, called ***superimposed ice***. Summer melting exposes this ice. The exposed part that remains at the end of the melting season is called the ***superimposed-ice zone***. The ***firn line*** (the boundary between the current year's snow and the newer ice) is easy to find.

ablation zone: the zone below which all of the firn and superimposed ice from the current year has been melted away, to expose older ice.

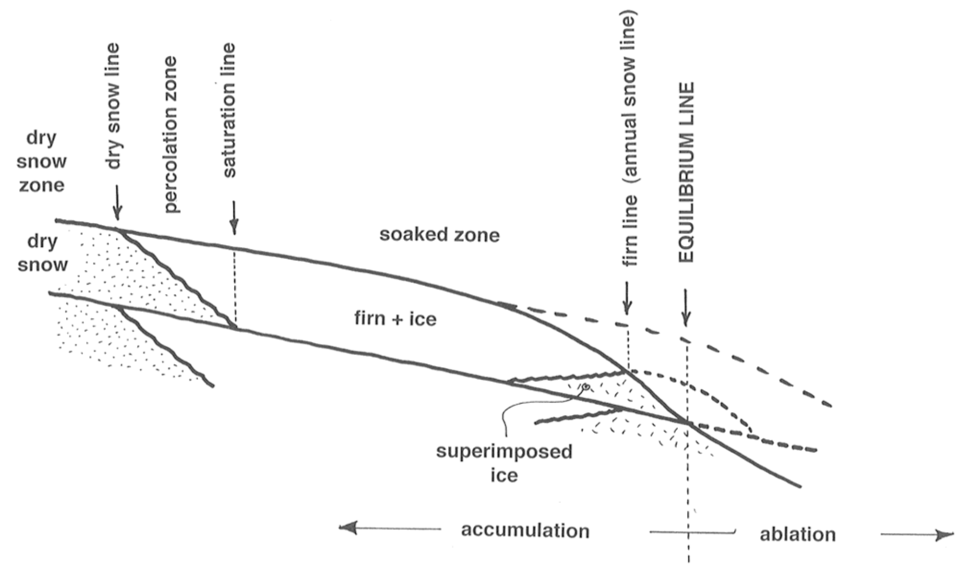


Figure 7-7. Schematic streamwise vertical cross section through a glacier in the vicinity of the equilibrium line.

This is all highly idealized. Most glaciers don't show all the zones, or at least not every season. The dry snow zone may be absent; melting eliminates all the superimposed ice if the net balance is negative; and the relative importance of evaporation and percolation varies. In a series of negative years, one can often see several "firn lines" or "firn edges" if development of superimposed ice is not important (Figure 7-8).

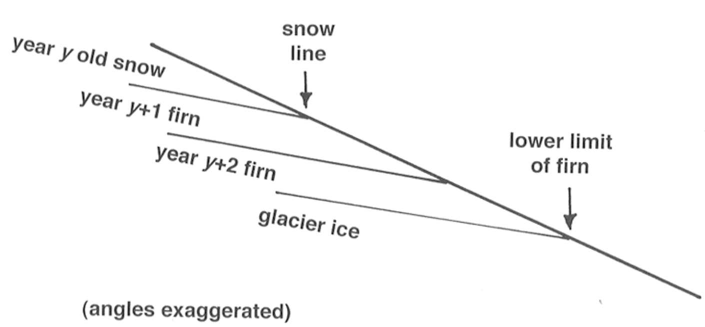


Figure 7-8. Detail of firn lines developed during a series of years, in vertical streamwise cross section.

5.2.11 A concept closely related to the equilibrium line is the ***annual snowline***: the lower limit of current year's snow on the glacier surface. What are the controls on the snowline? Mainly *winter precipitation* and *summer temperature*. The annual snowline is the local manifestation of what's called the ***regional snowline***: a band with a width up to a couple of hundred meters in elevation in which the local snowline lies on regional scale (This band is patchy locally but very consistent regionally.) Figure 9 shows the average position of the regional snowline as a function of latitude, as well as the average precipitation.

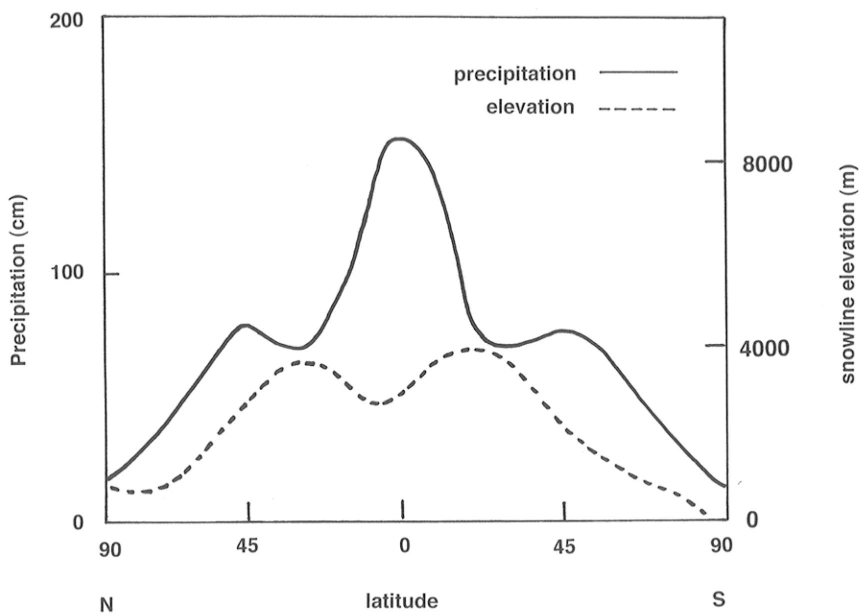


Figure 7-9. Combined plot of precipitation and snowline elevation as a function of latitude.

6. MOVEMENT OF GLACIERS

6.1 General Aspects of Glacier Movement

6.1.1 An important consequence of the balance considerations in the previous section is the glacier tends to thicken in its upper part and thin in its lower part, thus increasing its surface slope. The glacier flows under its own weight, passing ice across the equilibrium line to maintain an equilibrium slope. Figure 7-10 is a highly schematic view of how the glacier does this. Figure 7-11, a streamwise vertical cross section through a representative valley glacier, is a less idealized way of looking at this.

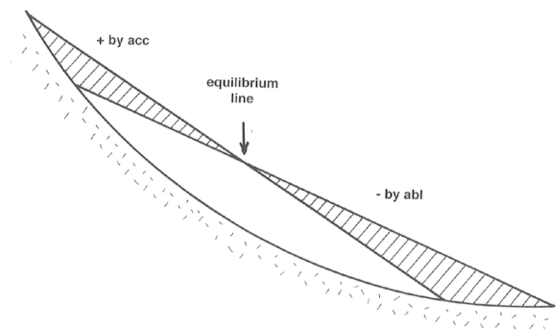


Figure 7-10. Schematic view of gain and loss of ice in a sloping glacier.

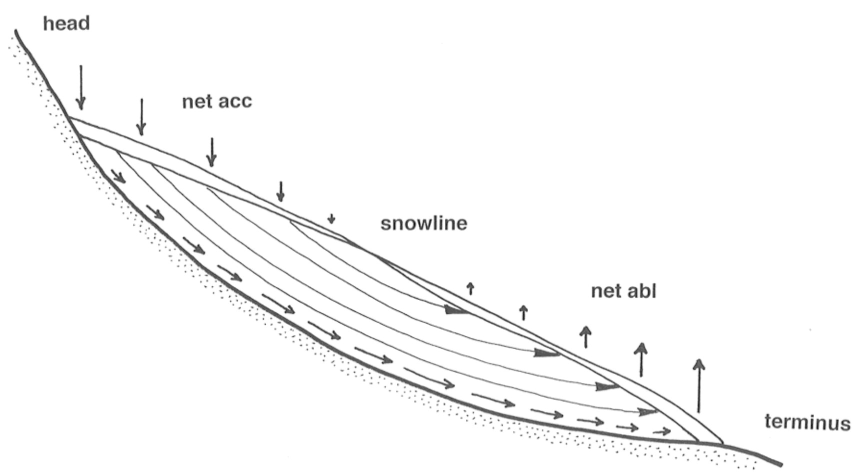


Figure 7-11. Streamwise vertical cross section through a representative valley glacier, showing accumulation, ablation, and flow lines.

6.1.2 The discharge (volume rate of flow) of a glacier is greatest at the snowline. Why? Because if the profile is to stay the same, the annual discharge through any cross section equals the integral of accumulation minus ablation over the whole area of the glacier above that cross section, and it's at the equilibrium line that this integral is greatest, because upglacier of the equilibrium line, where the accumulation curve and the ablation curve intersect, the accumulation curve is everywhere above the ablation curve; see Figure 5. Does glacier flow fastest there too? Maybe or maybe not; it depends on how the thickness of the glacier varies in longitudinal profile.

6.1.3 Figure 7-12, similar to Figure 7-10, shows a vertical cross section through an ice sheet. In a typical ice sheet the equilibrium line is near the margins of the glacier, because accumulation is small over a broad area but ablation is concentrated near the margins. Note the increase in near-bottom velocity toward the margin and the very slow movement of ice deep in the center of the glacier. If you want to find the oldest ice, that's where to look.

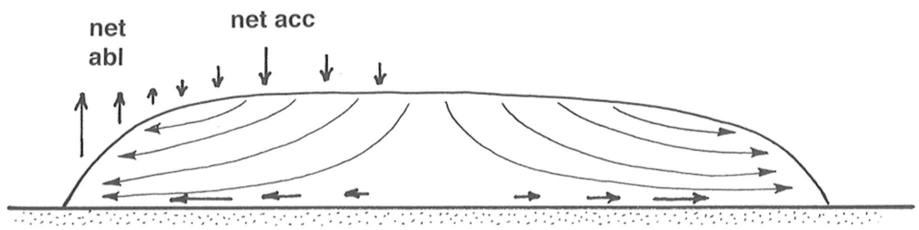


Figure 7-12. Vertical cross section through a representative ice sheet, showing accumulation, ablation, and flow lines.

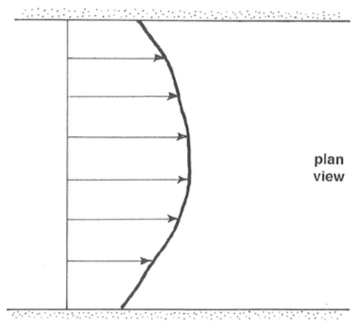
6.2 Velocity of Glaciers

6.2.1 A glacier doesn't move fast enough for the motion to be perceptible directly to the eye, although in some cases speeds are up to several tens of meters per day, so you could almost see it moving, if you had just the right observation post where you could get a good view of the contact between the base the glacier and the underlying bedrock. But usually speeds are of the order of meters per week or meters per month.

6.2.2 Surface velocities on a valley glacier can be measured easily by planting a straight row of stakes across the glacier, surveying their positions relative to fixed points on the bedrock walls, and then coming back at a later time to resurvey the positions to see how far the glacier has moved downvalley. (This kind of thing is not nearly as easy for ice sheets and ice caps, though.) The first systematic work of this kind was in the French Alps in the 1840s.

6.2.3 If you looked at the resurveyed positions of the row of stakes, what kinds of profiles would you see? Figure 7-13, a map view, shows what's generally found: an approximately parabolic or U-shaped velocity distribution. Just as with the free-surface flows we looked at in Chapters 1 and 5, this variation in velocity across the profile is a manifestation of internal shearing deformation.

6.2.4 Figure 7-14 shows that you can resolve the total movement into two components: ***internal deformation*,** and ***basal slip***, or basal sliding. Basal slip varies in importance from *zero* in cold, slow glaciers to *large* in warm, fast glaciers.



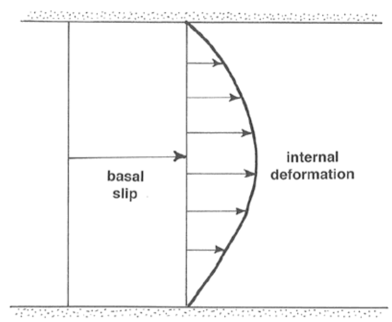
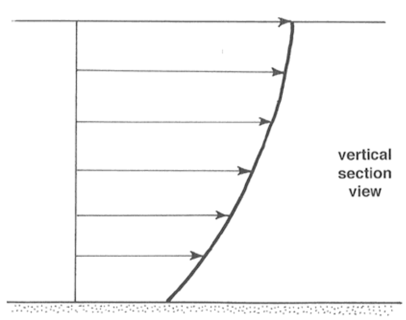


Figure 7-13 (left). Map view of a valley glacier, showing velocity distribution.

Figure 7-14 (right). Map view of a valley glacier, showing resolution of movement into basal slip and internal deformation.



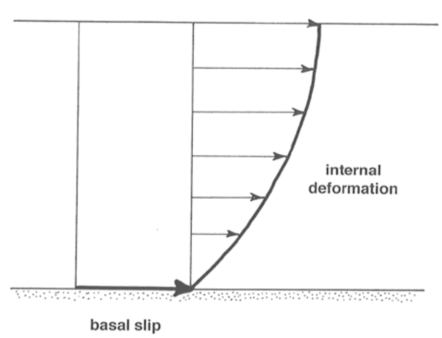


Figure 7-15 (left). Vertical streamwise cross section through a valley glacier, showing velocity distribution.

Figure 7-16 (right). Vertical streamwise cross section through a valley glacier, showing resolution of movement into basal slip and internal deformation.

6.2.5 Figures 7-15 and 7-16 show that the picture of velocity distribution is similar in vertical section. Again the motion can be resolved into a component of basal slip and a component of internal deformation.

6.2.6 Figure 7-17 is an example of an unusually complete result on the velocity of ice at the glacier surface, relative to bedrock, as a function of streamwise position on a valley glacier.

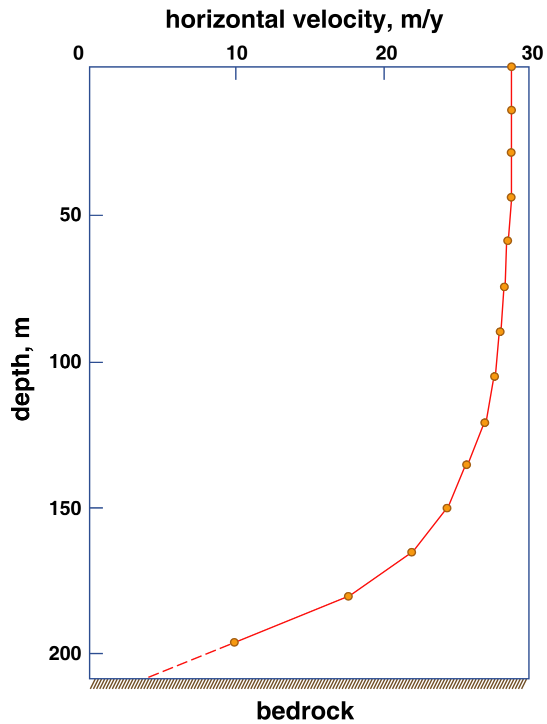


Figure by MIT OCW.

Figure 7-17. The vertical velocity profile in the Athabasca Glacier, Canada. (From Savage and Paterson, 1963.)

6.2.7 One other aspect of glacier velocity is important to understand. Because of accumulation in the upper part of the glacier, what is at the glacier surface at a given time is found at some depth below the glacier surface at a later time. The velocity at which some marker object moves downward from the surface is called the ***submergence velocity***. The corresponding velocity of upward movement of the marker toward the glacier surface in the zone of ablation is called the ***emergence velocity***. In glaciers in temperate regions, a typical value of submergence velocity in the accumulation zone is 1 m/yr, and a typical value of emergence velocity in the ablation zone is 3–4 m/yr.

ADVANCED TOPIC: HOW THE SURFACE VELOCITY OF A GLACIER IS MEASURED

1. How is the surface velocity of a glacier actually measured? Basically, by standard surveying techniques—but there's more to it than meets the eye. In the following discussion, refer to Figure 7-18. Suppose that you implant a vertical stake in the surface of the glacier somewhere in the ablation area. The slope angle of the ice surface in this part of the glacier is α, and the thickness of the glacier, the minimum distance from bedrock to the glacier surface, is *H*.

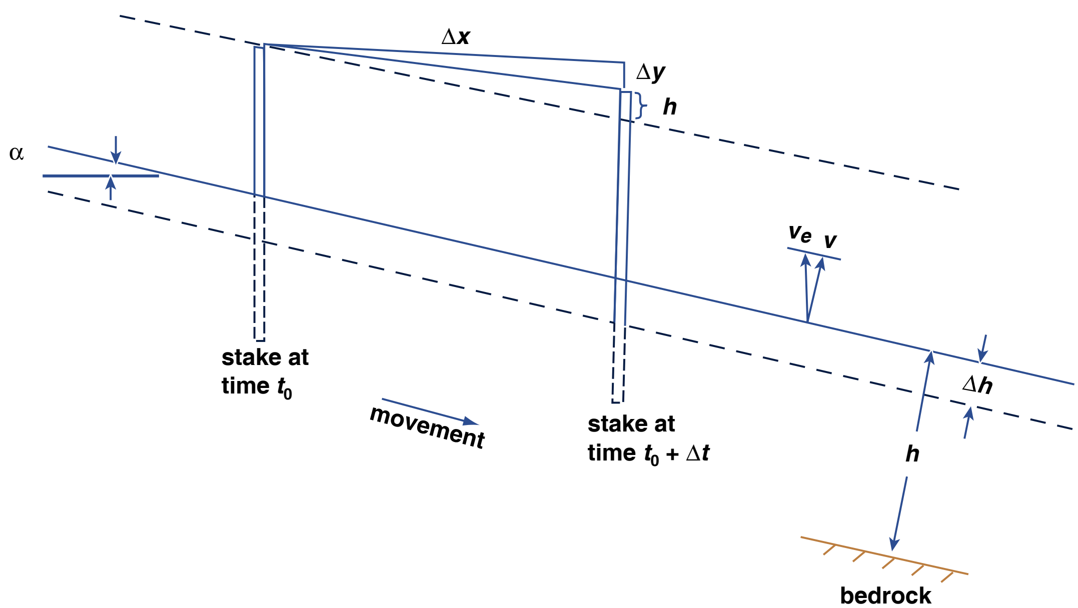


Figure by MIT OCW.

Figure 7-18. Definition sketch for analysis of the surface velocity of a glacier.

2. Wait a time Δt, while the stake moves some distance downglacier. The horizontal component of movement of the stake, relative to the bedrock beneath the glacier, between time t_o and time $t_o + \Delta t$ is Δx, and the vertical component, again relative to bedrock, is Δy. Clearly the horizontal component of time-average velocity *U* of the glacier, relative to bedrock, is $\Delta x/\Delta t$. Likewise, the vertical component of velocity, again relative to bedrock, is $\Delta y/\Delta t$. And the resultant velocity of the glacier relative to bedrock is the vector sum of these two velocity components.

3. Note in Figure 7-18 that the ice surface is lowered (or raised) by a distance ΔH normal to the ice surface during time Δt. Think about the net rate of change in ice thickness, $\Delta H/\Delta t$. Part of $\Delta H/\Delta t$ is accounted for by downward melting of the ice surface, but part is caused by streamwise compression or extension of the whole glacier at this cross section, independently of ablation or accumulation at that point.

4. Let's deal first with the downward melting of the ice surface. This leads to another interesting vertical velocity: the vertical velocity of the ice relative to the local plane of the glacier ice surface. This is the rate at which the ice surface would rise or fall vertically if there were no ablation or accumulation. It's this vertical velocity that was called the emergence velocity (in the ablation zone) or the submergence velocity (in the accumulation zone) in Paragraph 6.2.7. It's so called because if the glacier is at equilibrium with a balanced economy the emergence velocity is how fast things embedded in the ice approach the surface, and the submergence velocity is how fast things put on the ice surface "move away" from the surface to become embedded in the glacier.

5. Shown at the top of the stake at time $t_0 + \Delta t$ is the distance h, the change in elevation of the top of the stake relative to the plane of the ice surface (*not* relative to the bedrock beneath the glacier). By use of some trigonometry, the emergence velocity V_e can be written

$$V_e = \frac{h}{\Delta t} = \frac{\Delta x \tan \alpha - \Delta y}{\Delta t} \tag{1}$$

6. The component v_e of V_e perpendicular to the ice surface is just $V_e \sin \alpha$. This represents the part of the net rate of change of ice thickness H caused by melting; positive v_e corresponds to negative $\Delta H/\Delta t$. The other part of ΔH, remember, is caused by extension or compression; call it v_d. The sign of $\Delta H/\Delta t$ is determined by the relative values of v_e and v_d: in the ablation zone, if v_d is negative then $\Delta H/\Delta t$ has to be negative, because positive v_e contributes to negative $\Delta H/\Delta t$. Only if v_d is positive and greater in absolute value than v_e does the ice surface rise relative to bedrock despite downmelting of the ice surface.

6.2.8 So far we've talked only about surface velocities. How does one find the velocity of a glacier at depth? It's a big job. Drill a borehole from the surface to the base of the glacier with a thermal drill, at a speed of something like a meter per hour. This is fairly easy in glacier ice that's at its melting temperature, but it's very difficult in glacier ice that's below its melting temperature. After you finish the hole you have to install a casing, both to keep out meltwater and to keep the hole from closing up by deformation!

6.2.8 Then put an inclinometer down the hole and read the inclination of the hole as a function of depth; that allows you to plot the vertical profile of the hole (relative to the glacier surface) as a function of time and therefrom find velocities (relative to the glacier surface). This relative-velocity profile is converted to an absolute-velocity profile by combining it with the already-measured absolute surface velocity. But what you cannot determine by this method is the vertical component of velocity at depth. Finally, if your hole reaches bedrock, you can indirectly measure the basal slip. And even if the hole ends a short distance from the base of the glacier, you can get a good idea of the basal slip just by extrapolation. Figure 7-19 shows an example of a vertical velocity profile measured in this way.

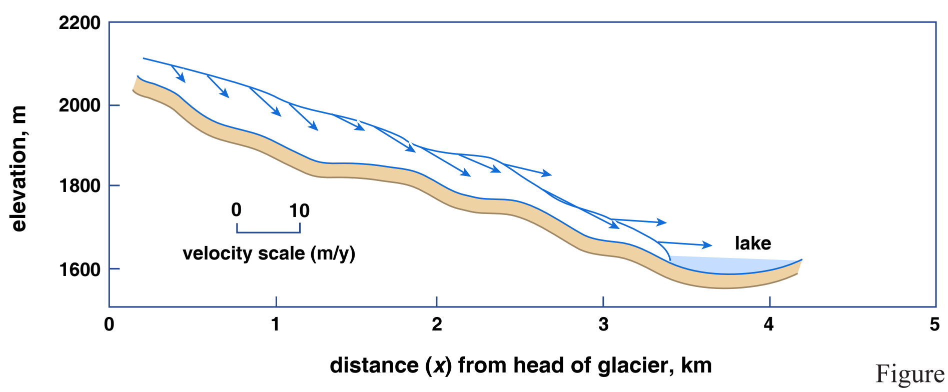


Figure by MIT OCW.

Figure 7-19. Surface velocity vectors from South Cascade Glacier, Washington, U.S., (From Meier and Tangborn, 1965.)

7. DEFORMATION OF ICE

7.1 As in many crystals, the way ice crystals are deformed or sheared when a stress is applied is by propagation of dislocations through the crystal. A ***dislocation*** is *a line defect in a crystal that disrupts the otherwise ideal and regular arrangement of atoms or molecules.*

7.2 Figure 7-20 shows a simple, idealized example of a dislocation. If we sheared this crystal, the dislocation could migrate simply by breakage of bonds and formation of a new bonds. The net effect is to move the "defected" plane to the right relative to the lower part of the crystal. The dislocation shown in Figure 21 is a line defect that extends indefinitely in one dimension; there are other kinds of dislocations as well.

7.3 Ice deforms mainly by propagation of dislocations along the *a* axis (that's what the direction perpendicular to the hexagons in the ice structure are called; see the section on ice structure in Chapter 1), so slip is along the basal plane (that is, the plane perpendicular to the *a* axis). Here's a loose but not misleading analogy: think of ice as a pack of cards oriented along the basal planes—they're easy to deform by simple shear along these planes. It's been shown experimentally that there's no preferred direction of gliding within the basal plane itself, and it can be demonstrated theoretically that to within a few degrees there shouldn't be any. There can be gliding by dislocations in nonbasal planes, but that's much harder—stresses 10–20 times as great are needed—and apparently it's unimportant.

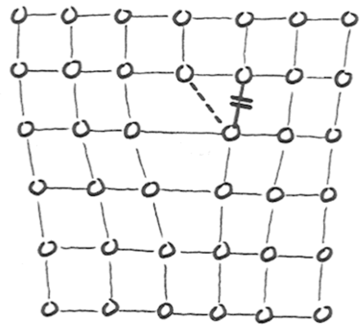


Figure 7-20. Idealized example of a dislocation.

7.4 Another way of looking at the nature of deformation of ice is to compare its behavior with other materials in a graph of deformation rate vs. applied shearing stress (Figure 7-21). I mentioned in Chapter 1 that certain fluids, air and water included, show a linear relationship between the applied shearing stress and the rate of shearing deformation. Such fluids are called ***Newtonian fluids***. Fluids that show some other kind of relationship between stress and rate of strain are called ***non-Newtonian fluids***. Ice is one of those. Materials like ice are harder to deform with increasing stress, so that the curve of deformation rate against shear stress is convex upward.

ADVANCED TOPIC: THE FLOW LAW FOR ICE

1. The relationship between applied stress and rate of deformation for a continuum is called a ***flow law***. What is the flow law for ice? The flow law for ice (both single-crystal and polycrystalline) is of the form

$$\text{rate of shearing deformation} = A(\text{shearing stress})^n \qquad (2)$$

where A is a coefficient and n is an exponent. (In the case of water or air, the exponent n is just 1.)

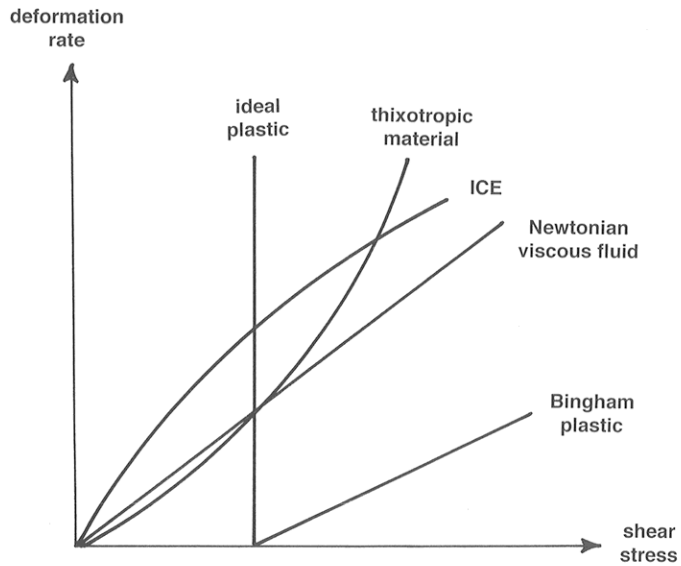


Figure 7-21. Graph of deformation rate against applied shear stress, showing behavior of various kinds of materials.

2. In simple shear (Figure 7-22),

du

(3)

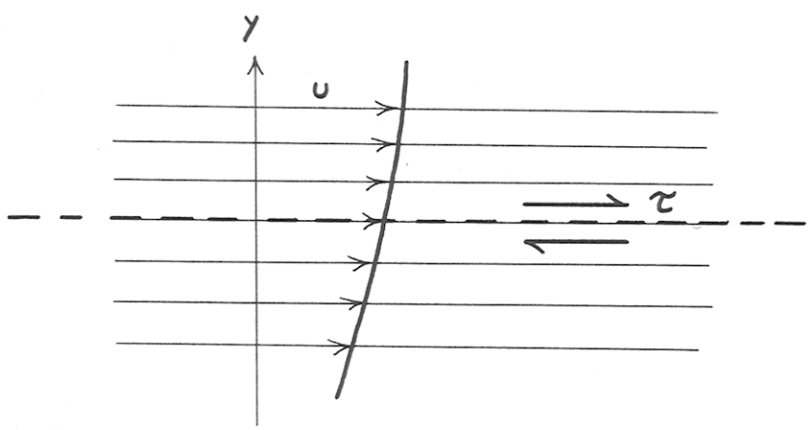


Figure 7-22. Definition sketch for analysis of simple shear in glacier ice.

The derivative *du*/*dy*, the rate of change of velocity with distance perpendicular to the planes of shearing, is what I referred to above as the rate of shearing deformation. (The coefficient 2 gets there by virtue of the way the rate of deformation is defined; don't worry about it.) In Equation 3 the coefficient *A* is like 1/viscosity. It depends strongly on temperature. As you might expect, it's much smaller (that is, the viscosity is much greater), by two orders of magnitude, for polycrystalline ice than for single-crystal ice oriented with the basal plane parallel to the shear planes. This is because deformation of ice is by internal gliding along the basal planes, and in polycrystalline ice, most crystals are not oriented favorably for this. The exponent *n* is hard to measure accurately, even in the laboratory! The value usually cited is 3 for polycrystalline ice—but there's nothing exact about this value. And n doesn't depend importantly on either temperature or pressure.

ADVANCED TOPIC: THE DOWNSLOPE FLOW OF GLACIERS

1. One thing we can do by way of theory on glacier movement is to consider the equilibrium of a given reach or streamwise segment of the glacier by writing a force-balance equation. This parallels the derivation of the fundamental resistance equation for rivers in Chapter 5.

2. Think about a somewhat idealized glacier of uniform thickness and infinitely wide lateral extent flowing down a planar bedrock surface (Figure 7-23). There must be a balance between the downslope component of gravity (the driving force) and the friction between the glacier and its bed (the resisting force), just as in steady uniform flow of water in an open channel. Accelerations of the glacier are very small and can be ignored for this purpose, although they are not unimportant in other respects.

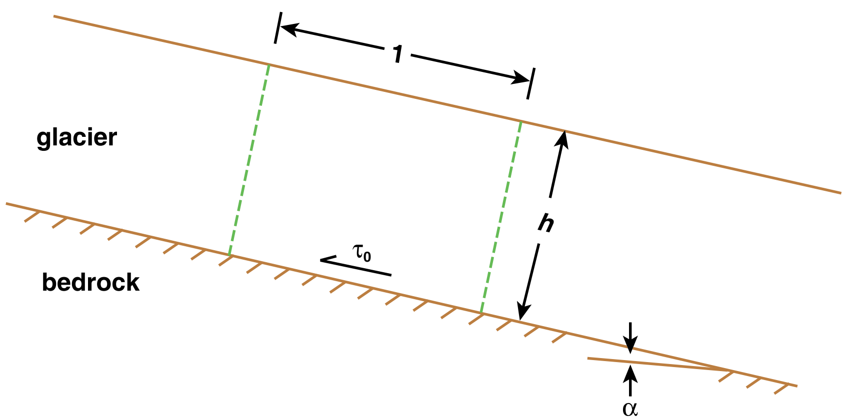


Figure by MIT OCW.

Figure 7-23. Definition sketch for force balance on a mass of ice in a valley glacier, to derive a resistance equation for glacier flow.

3. Writing the force balance for a block of the glacier with unit width and unit length, as shown in Figure 7-23 (just as for steady uniform channel flow of water; see Chapter 5 on rivers),

$$\tau_o = \rho g h \sin \alpha \qquad (4)$$

where τ_o is the shear stress, ρ is the density of ice, g is the acceleration of gravity, h is the thickness of ice, and α is the slope angle.

4. Here are two ways Equation 4 is important:

- It provides the best way of estimating the boundary shear stress of a glacier: typically 0.5–1.5 bars (1 bar is approximately 1 atmosphere). Note: this is far smaller than the hydrostatic pressure at the base of the glacier.
- This doesn't tell us anything about velocities in the glacier. But it can, if we combine it with the flow law for ice. See the following development if you're interested in the details.

5. First modify the resistance equation, Equation 4, to get τ as a function of position above the bed (Figure 7-24):

$$\tau = \rho g(h - y)\sin\alpha \tag{5}$$

Now we have two equations for τ:

$$\tau^n = \frac{1}{2A}\frac{du}{dy} \tag{6}$$

$$\tau^n = [\rho g \sin\alpha(h - y)]^n \tag{7}$$

Eliminating τ from Equations 6 and 7,

$$\frac{du}{dy} = 2A[\rho g(h - y)\sin\alpha]^n \tag{8}$$

This simple differential equation is easy to solve:

$$u = 2A(\rho g \sin\alpha) n \int_0^h (h-y)^n dy \quad + c \tag{9}$$

To evaluate the constant of integration c, use the boundary condition that $u = u_s$, the surface velocity, at $y = h$. You find that $c = u_s$. So

$$u_s - u = \frac{1}{n+1} \quad 2A(\rho g \sin\alpha)^n (h-y)^{\,n+1} \tag{10}$$

If $u = u_b$ and $y = 0$, where u_b is the basal-slip part of the glacier velocity, Equation 10 becomes

$$u_s - u_b = \frac{1}{n+1} \quad 2A(\rho g \sin\alpha)^n h^{n+1} \tag{11}$$

Using the value of 3 for n, Equations 10 and 11 become

$$u_s - u = \frac{A}{2} (\rho g \sin\alpha)^{\;3} (h-y)^4 \tag{12}$$

$$u_s - u_b = \frac{A}{2} (\rho g \sin\alpha)^{\;3} h^4 \tag{13}$$

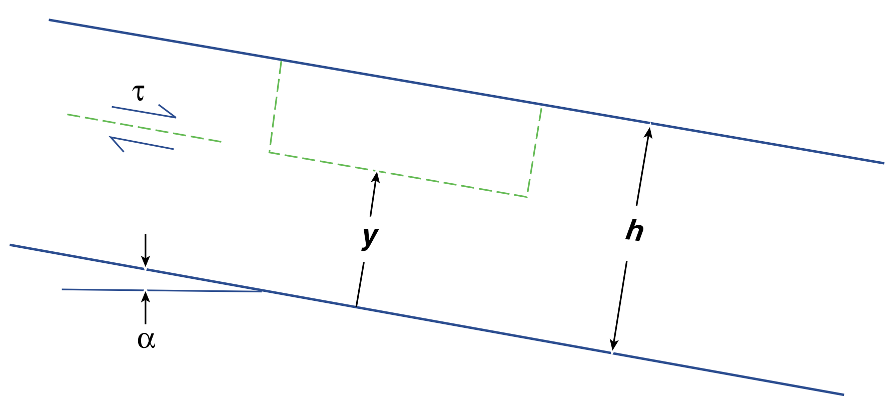


Figure 7-24. Definition sketch for force balance on a mass of ice in a valley glacier, to derive a velocity distribution along a line normal to the base of the glacier.

6. Note in Equations 12 and 13 that, other things being equal, glacier velocity varies as the third power of the slope and the fourth power of the thickness.

7. All of this is for steady uniform flow, but because glaciers have very small accelerations, and to a first approximation they are sheets, it's not bad. But temperature has a very important effect, through variation of the value of *A*.

8. How doesa theoretical result like this compare with measured velocities? Not bad. So far, such comparisons have been made only for valley glaciers. (To do so, one has to extend the model slightly to deal with a non-infinitely-wide channel, but that's straightforward).

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8. THE THERMAL STRUCTURE OF GLACIERS

8.1 Processes of Energy Exchange at Glacier Boundaries

8.1.1 A glacier can gain or lose heat in several ways. It's clear that these processes operate both at the surface of the glacier and at the base of the glacier. What are these processes, and what is their relative importance? Table 1 lists the ways heat can be gained or lost by a glacier at its surface.

8.1.2 First of all, keep in mind that the combined effect of all of these processes depends on whether the glacier is at its melting point or below its melting point (Figure 7-25). Obviously, if the temperature of the glacier is below the melting point, then the effect of addition of heat is to raise the temperature and the effect of *extraction* of heat is to *lower* the temperature. But if the temperature of the glacier is at the melting point, then addition of heat serves to melt glacier ice. Of course, extraction of heat when the glacier is at the melting point lowers the temperature below the melting point.

mechanism	condition
How is heat gained by glaciers?	
solar radiation (short-wave)	sun shines (even diffuse light)
long-wave radiation, space/atmosphere/ice	+ balance, complicated
conduction from air	dT/dh + in air; $T_{ice} < T_{air}$ at h = 0
condensation (or sublimation)	$d/dh(P_{H_2O})$ + in air; T_{ice} < dewpoint of air at $h = 0$
falling of rain	$T_{ice} < T_{rain}$
How is heat lost by glaciers?	
long-wave radiation, space/atmosphere/ice	- balance, complicated
conduction to air	dT/dh - in air; $T_{ice} > T_{air}$ at $h = 0$
evaporation of water to air	$d/dh(P_{H_2O})$ - in air; T_{ice} > dewpoint at $h = 0$

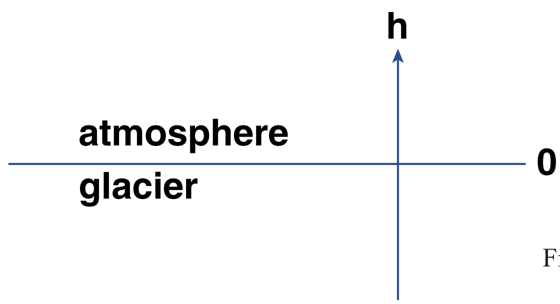


Figure by MIT OCW.

Table 7-1. Heat-exchange processes associated with a glacier.

8.1.3 An important indirect factor in all of this is heat conduction within the ice. This is the way heat is moved to or from the glacier surface, and it controls rates of radiation or conduction of heat at the glacier surface, because surface radiation and conduction is a function of the surface temperature of the ice.

$$\text{If } T < 0, \text{ change } T$$

$$\text{If } T = 0, \begin{cases} \Delta Q\ -, & \text{decrease } T \\ \Delta Q\ +, & \text{melt ice} \end{cases}$$

Figure 7-25. Effect of heat-exchange processes on the temperature of a glacier.

8.1.4 Here are some miscellaneous descriptive points about Table 7-1, keyed by number:

(1) The ***solar constant*** is *the rate at which the Sun delivers heat to the Earth*. It's conventionally taken to be the value that would be measured just outside the Earth's atmosphere when the distance between the Sun and the Earth is at its mean annual distance. The value of the solar constant is very close to 2 cal/cm^2-min.

(1) The ***albedo*** of the Earth is *the percentage of incoming solar radiation that is reflected directly back to space*, on the average. The albedo of a glacier varies considerably: snow surfaces have an albedo of 0.7–0.9, but glacier ice has an albedo of only 0.2–0.4.

(2) The surface energy exchange is greatly different under clear skies and under cloudy skies. Under clear skies, radiation directly to space is involved; typically the glacier loses heat, unless the air is very warm. Under cloudy skies, the direction of net flux of energy depends mostly on the relative temperature of clouds and ice.

(3) This would be minor without the effect of the wind. When the wind blows, there's turbulent diffusion, which you know from the section on fluid dynamics in Chapter 1 to be much greater than molecular diffusion; then conduction to or from the overlying air can be very important.

(4) This is a minor effect.

(5) This isn't important if the ice is at its melting point; 10 cm/day equals one day of long-wave radiation. But if the rain freezes when it falls, then this can be an important effect.

8.1.5 Energy exchange at the base of the glacier is a simpler matter, and the magnitudes of energy flux are not only much smaller but also far less variable. Geothermal heat is enough to melt about 5 mm of ice per year. This may not sound like much, but in a glacier whose basal ice is at the melting point it plays a significant role in glacier movement, by way of the lubricating effect of the thin film of water that's continuously produced and then slowly drained away. There's also heat from internal friction. This tends to be produced mainly in the lower part of the glacier, both because of basal slip and also because shearing is strongest in the lowermost part of the glacier. The rate of heat generation by internal deformation varies from considerably less to considerably more than the heat flux from the bedrock below, but it's about the same order of magnitude.

8.2 Thermal Characteristics of Glaciers

8.2.1 The temperature characteristics of glaciers are important because deformation of ice, and therefore flow of glaciers, is strongly dependent on the temperature of the ice. At first thought it may seem that all glaciers are cold. But an important distinction can be made between

> **cold ice:** ice below its pressure melting point (there can be no liquid water), and
>
> **warm ice:** ice at the pressure melting point (there's at least a little liquid water, under or between grains).

8.2.2 Recall from Chapter 1, in the section on water, that the melting point of ice falls slightly with increasing pressure. That translates to about a 2°C decrease beneath a thick ice sheet. Therefore there's actually a downward temperature gradient in a glacier, although it's small.

8.2.3 You can't necessarily classify an entire glacier as warm or cold, because commonly the upper part or the geographical interior of the glacier has cold ice and the lower part or the lower-latitude fringes has warm ice. But in a given geographical region of the glacier, if the glacier is such that in winter all of the ice is below the melting point, and only the surficial part is raised to the melting point in summer, the glacier is said to be a ***cold glacier***. On the other hand, if all of the ice is raised to the melting point in summer, and only a part is cooled below the melting point in winter, the glacier is said to be a ***warm glacier***.

8.2.4 One of the most important consequences of the thermal structure of glaciers has to do with basal phenomena. Therefore we can talk about ***warm-based glaciers*** and ***cold-based glaciers***. It's generally believed that cold-based glaciers show little or no basal slip; the ice is frozen fast to the bedrock, and all movement is by internal deformation. On the other hand, a warm glacier has a thin layer of water at its base, facilitating basal sliding.

8.2.5 Cold ice is formed in two different ways:

(1) *Accumulation is at a temperature so low that there's no surface melting during the summer*. This is the case over most of the Antarctic ice sheet. The temperature of the firn and ice below the level of seasonal temperature change is approximately the same as the mean annual air temperature at the site. But temperature increases downward, because of geothermal heat. See Figure 7-26. Two opposing tendencies determine the course of the curve in Figure 7-26: (1) the value of geothermal heat flux from below, and (2) the rate of firn accumulation, which tends to "carry cold downward" into the ice sheet.

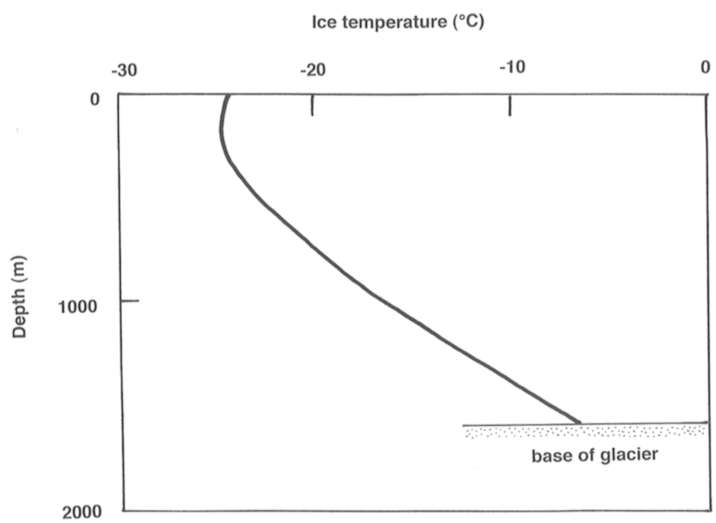


Figure 7-26. Graph of ice temperature vs. depth within a thick cold glacier.

(2) *Cooling the surface layer by winter cold.* This effect extends down as much as 20 m. This happens at the surfaces of all glaciers in winter.

8.2.6 How is warm ice formed? By heating to raise the ice to its melting point (anywhere). This happens at the surface of the glacier, by one or more of the heat-transfer processes listed above, to form a surficial warm layer, and then meltwater percolates down and warms the ice by refreezing. This is an important effect: when 1 g of meltwater freezes, enough latent heat is released to raise 160 g of ice 1°C. Note that this warms the ice but doesn't melt it. Note also that latent heat is the only source for warming at depth, because there's almost no temperature gradient and therefore almost no conduction.

8.2.7 Warm ice is produced at the base of the glacier by basal heat sources. This is favored by the following circumstances:

- thick ice
- high surface temperature
- low accumulation rate
- high ice velocity

Over large areas of both the Antarctic and Greenland ice sheets, the basal ice is at its pressure melting point!

8.2.8 An important point is that when even a thin layer of warm ice is produced at the base of the glacier, the temperature gradient is about zero, so all the basal heat (frictional and geothermal) is used for melting, because there's no conduction. This provides a continuous supply of meltwater at the base of the glacier.

9. GLACIAL MELTWATER

9.1 General

9.1.1 ***Glacial meltwater*** is *the liquid water produced by ablation of glaciers*. Meltwater is in most glaciers by far the most important product of ablation; it's much more important than evaporation. Of course, in glaciers that terminate in the ocean, calving is more important.

9.1.2 The importance of glacial meltwater is twofold:

- It's intimately involved with the movement of glacier ice, by way of its influence on both creep and basal slip.
- Meltwater can carry enormous quantities of glacial sediment and deposit that material nearby or far away from the glacier.

9.1.3 Whereas the activity of the glacier ice itself is greatest near the equilibrium line, the activity of meltwater increases to a maximum at the terminus of the glacier.

9.1.4 Glacial meltwater is abundant on the surfaces of all temperate glaciers below the snowline. Even during the melting season, surface meltwater isn't common above the snowline, because the water readily sinks into the melting snow. Downglacier of the snowline, however, surface flows of meltwater during the melting season are common. Meltwater is present on polar glaciers only locally and temporarily, because it soon refreezes.

9.1.5 Usually the meltwater streams on the surface of a glacier plunge down into the body of the glacier before they reach the terminus; remember that liquid water is more dense than ice, so the meltwater tries to find a way down into the glacier. It's a little like pouring water into a pan of semiconnected ice cubes. You know what eventually happens: the ice cubes end up floating in the water. In a real sense, the meltwater of a glacier has a tendency to work its way underneath the glacier so as to cause the glacier to float in its own meltwater. The only thing that keeps a real glacier from becoming like the semiconnected ice cubes floating in your pan is that the balance between production and drainage of meltwater favors drainage over production.

9.1.6 Meltwater is classified on the basis of where it is in the glacier, as ***supraglacial meltwater*** (on top of the glacier), ***englacial meltwater*** (inside of the glacier), and ***subglacial meltwater*** (beneath the glacier). In the same way, the sources of meltwater can be viewed as surficial, internal, and basal. On most glaciers the surface sources are much greater than the internal or basal sources, by one or even two orders of magnitude. Surface sources are strongly seasonal, but internal and basal sources are largely unaffected by the seasons.

9.1.7 Surface sources:

- The main source is ablation during the summer melting season. This drops off sharply upglacier—just the reverse of normal fluvial watersheds.
- Rainfall on the ablation area, mainly in the warm season, is another source. In a narrow technical sense this isn't meltwater, but it's indistinguishable from true meltwater and is always considered in the same way.

9.1.8 Internal and basal sources:

- If the ice at the base of the glacier is at its pressure melting point, geothermal heat melts ice there rather than being conducted upward. Depending on the local value of heat flow, this accounts for one or two centimeters of ice thickness per year. Films of meltwater at the bases of glaciers have been observed directly.
- *Frictional heating by both internal deformation and basal sliding* cause melting if the ice is at its pressure melting point. This generally accounts for something like 0.5 to 5 cm of ice per year.

- Some meltwater is produced by melting by the heat generated by *the friction of the meltwater flow itself*. The quantities are not important, but this effect seems to be important in creation and maintenance of englacial and subglacial drainageways.
- *Groundwater flow* out of the regolith and bedrock beneath the glacier is locally important.

9.2 Drainage Routes

9.2.1 We observe surface runoff in the form of supraglacial channels, but we also see these surface meltwater streams plunging into crevasses and vertical tubular holes called ***moulins***. And we see important meltwater streams emerging from beneath the glacier at the terminus. So there must be important englacial and subglacial meltwater routes as well as supraglacial routes.

9.2.2 It's easy to make casual studies of surface meltwater drainage, but there have been few detailed studies. And it's difficult to study subglacial and especially englacial streams, for obvious reasons. But there are some fairly reliable theoretical approaches that are broadly consistent with what's known about englacial and subglacial drainage and its depositional consequences.

9.2.3 Surface streams form wherever more meltwater is produced than can be absorbed locally into the glacier or held as pore water in firn or snow at the glacier surface. The stream channels range in size from tiny rills to large channels several meters wide and deep. The streams may or may not be incised into valleys on the glacier surface; that depends on the relative rates of channel downmelting and interchannel ablation.

9.2.4 Meltwater streams on glacier-ice surfaces have a strong tendency to form meanders, much like streams flowing on land. Meandering on ice is no better understood than meandering on sediment, but presumably there's some kind of instability that involves preferential melting at certain points and less melting, or even refreezing, at other points. In many other cases, however, the courses of the channels are determined by lines of structural weakness in the ice.

9.2.5 Supraglacial streams also tend to form *dendritic stream networks*, as do stream systems on land. But there are several important differences:

- The network is *dense and rill-like*; it's hard to cut major trunk streams when the ice is moving and deforming.
- The drainage pattern shows *a strongly subparallel pattern*, because of the relatively steep slope of the glacier.
- The *drainage density decreases upglacier*, because meltwater production decreases upglacier.

- The channel pattern is *highly changeable*: it looks different each year depending on the development of new englacial drainageways. Remember that the whole glacier surface moves downslope, but the zones of overall extension and compression of the glacier caused by subglacial bedrock topography, which tend to control englacial drainage routes, stay in one place. Moulins develop, are used for several seasons, and then are abandoned as new ones from upglacier take their place.

9.3 Dynamics of Meltwater Flow Beneath the Surface

9.3.1 Here are some basic observations on englacial and subglacial meltwater flow:

• In the uppermost zone of the glacier, water flows approximately vertically down crevasses and moulins in free-fall flow. These vertical drainageways have a definite water level in them that can be measured fairly easily. (Try not to fall into one, though.) This water level changes with time, on scales of hours to days, depending on air temperature, sunshine, and rainfall.

• Typically, the changes in water level from moulin to moulin are correlated throughout the entire glacier, suggesting the existence of a connected water table within the glacier.

• This water table can vary from right at the glacier surface, at times of maximum meltwater production in the early part of the melting season, when there aren't a lot of easily exploitable passageways, to very deep (probably all the way to the base of the glacier) at the end of the melting season, when the passageways are fully developed but the supply of meltwater has dropped off.

• There have been few observations of the shapes of englacial drainageways below the moulins. These drainageways are thought to be nonvertical, with a large horizontal component to their orientation. One piece of evidence: circular englacial tunnels seen in the faces of freshly calved icebergs around the Greenland ice sheet. (A circular tunnel is the equilibrium shape if the water is flowing in a closed conduit in the ice, because approximate isotropy of the polycrystalline ice is usually a good assumption if the ice is far from the base of the glacier, where shearing by internal deformation is strongest.)

• *Travel times*: dye is injected at points of submergence and monitored where subglacial streams emerge. Speeds of movement are something like 1–2 km per hour, but there's wide variability. This means total travel times of one to several days on ordinary valley glaciers. But speeds must increase downstream within a given drainage system: the flow in the large ice tunnels at the downstream end of the network is probably very high, many meters per second.

The spectacular fountains formed where such subglacial tunnels discharge underwater in glacier-margin lakes is good evidence of this.

• What's known about flow in large subglacial tunnels? Often the flow emerging from the tunnels has a free surface, but this free surface usually slopes upward relative to the roof in the upstream direction, so it's reasonable to suppose that only a short distance upglacier the flow is closed-conduit flow, with no free surface. And the streams that terminate beneath glacial lakes are clearly flowing full all the way to the end. At times of low discharge near the end of the melting season, however, most or all of the conduits have free surfaces in them.

• Any glacial tunnel below a few tens of meters, where ice can flow plastically, tends to close up completely by inflow of ice, if it's not maintained open by some other means. There are two ways the tunnel can be held open: by water pressure equal to the hydrostatic pressure in the ice itself, or by melting of the walls of the tunnel by the flowing water. Figure 7-27 shows a graph of the half-life time for closing of a vertical ice tunnel as a function of overburden pressure for ice at its pressure melting point.

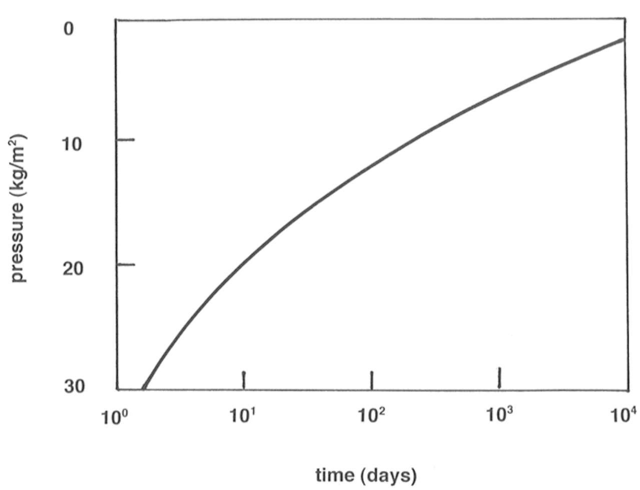
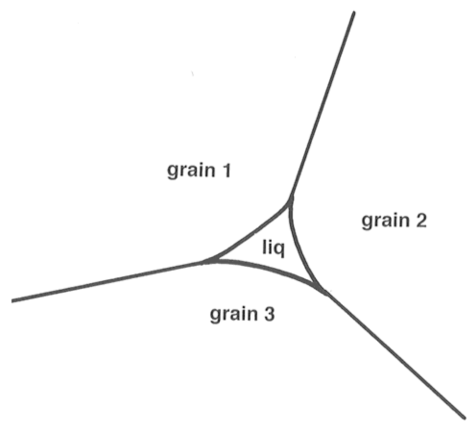


Figure 7-27. Half-life time for closing of a vertical ice tunnel as a function of overburden pressure for ice at its pressure melting point.

• Polycrystalline glacier ice at its pressure melting point is known both observationally and theoretically to be permeable to flow of water. There are veins or tiny passageways at three-grain linear boundaries (Figure 7-28), which meet in fours at four-grain point junctions. These junctions have generally tetrahedral shape. This can be observed in careful microscopic work with ice, and it also can be justified in terms of surface-energy arguments. The angle of junction between two ice grains and liquid water is about 20° (Figure 7-29). Water can therefore always percolate through a warm glacier, whatever the state of the large passageways.



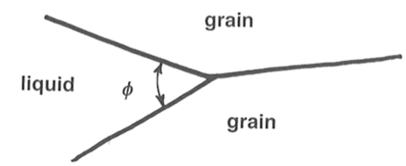


Figure 7-28 (left). Sketch of water passageways at three-grain linear boundaries in glacier ice.

Figure 7-29 (right). The angle of junction between two ice grains and liquid water is about 20°.

9.3.2 The basic idea about the hydraulics of meltwater flow within a glacier is that *there's a feedback between water pressure and ice pressure that controls the size of the flow tunnels*. To understand the nature of these adjustments, look at a simplified tunnel (Figure 7-30) that extends vertically from the surface to the base of the glacier and then horizontally to the terminus. The water table—the free surface of the water in the vertical part of the tunnel—is shown near the surface of the glacier.

9.3.3 If the tunnel is too big, it's able to carry a greater discharge than is supplied from the glacier surface, and the water level in the vertical tunnel falls. That decreases the water pressure farther along in the tunnel to a value less than the ice pressure around the walls of the tunnel, so the tunnel closes up, thus constricting the flow through the tunnel and causing the water level to rise until the water pressure builds up to be equal to the ice pressure, stabilizing the diameter of the tunnel.

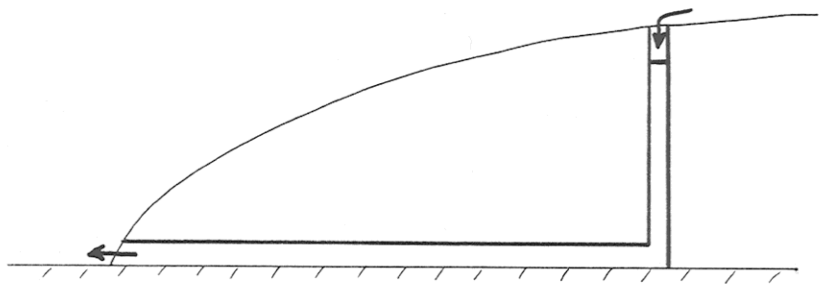


Figure 7-30. A simplified water-flow tunnel through a glacier.

9.3.4 Likewise, if tunnel is too small for a given meltwater discharge, the water table in the vertical part of the tunnel rises, thus increasing the water pressure to a value greater than the ice pressure at the tunnel wall, so the tunnel opens up by radial outflow of the ice wall, leading to increased discharge and a fall in the water table. The tunnel diameter thus again becomes stabilized at a value for which the water pressure is equal to the ice pressure on the tunnel walls.

9.3.5 In accordance with the foregoing argument, it's usually assumed that in a steady state (constant discharge, certain water level, certain tunnel diameter) the water pressure is equal to the ice pressure at every cross section in the tunnel. But there has to be one significant correction to this. Meltwater discharge through the tunnel tends to melt the tunnel walls, by two effects: (1) the heat generated by friction in the flow, and (2) heat carried from the surface by meltwater that's slightly above freezing. So at equilibrium the ice walls of the tunnel have to flow inward toward the center of the tunnel at a finite rate to balance this rate of wall melting. The adjustment described above is modified in such a way that the water pressure is a little greater than the ice pressure.

9.3.6 A good case can be made for another important consequence of this wall-melting effect: larger passageways grow at the expense of smaller passageways. The reason? (1) more heat relative to wall area is generated by viscous friction in the larger passageways than in the smaller passageways; and (2) more heat relative to wall area is carried by above-freezing water from the surface in the larger passageways than in the smaller passageways.

9.3.7 The consequence is that because of this differential growth of larger passages, the three-dimensional network of passageways in the glacier tends with time to become dendritic, with tributaries joining into ever-larger trunk passageways. This is broadly consistent with the few observations of the pattern of internal meltwater passageways in glaciers.

ADVANCED TOPIC: THE HYDRAULICS OF MELTWATER FLOW WITHIN GLACIERS

1. The vertical and then horizontal passageway in Figure 7-30 that was used to illustrate the foregoing points is clearly unrealistic. *What's the direction of meltwater flow in the ice?* That question leads in turn to what makes water flow inside a glacier in the first place. The answer to that latter question is: spatial gradients in the difference between the actual water pressure and the hydrostatic pressure (i.e., the pressure that would be measured at the given point if the water were not moving).

2. Figure 7-31 shows this effect in a simplified passageway. The hydrostatic pressure is constant all along the horizontal segment of the passageway, but the water flows from the higher tank to the lower tank because of the gradient in water pressure caused by the difference in water level between the two tanks. If that doesn't convince you, just consider that the state of flow in the horizontal segment would be exactly the same if you increase the water level in both tanks by the same vertical distance, thereby changing the hydrostatic pressure but not the gradient in pressure.

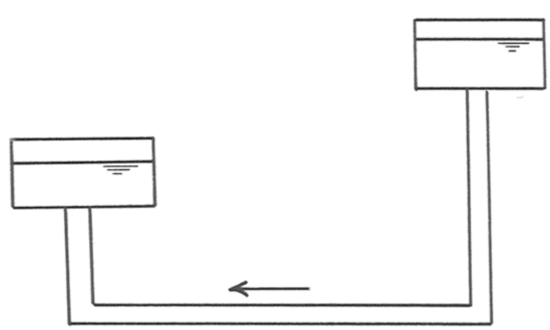


Figure 7-31. Flow of water in a horizontal passageway due to a gradient in water pressure.

3. Now look at the situation in a real glacier. Think about the water pressure and ice pressure at a point P within the glacier (Figure 7-32). *H* is the elevation of the ice surface above an arbitrary datum, and z is the elevation of Point P above that same datum. The ice pressure p_i at Point P is

$$p_i = \rho_i g(H - z) \tag{14}$$

and the water pressure p_w at the same point is approximately

$$p_w = p_i \tag{15}$$

by the line of reasoning in Paragraphs 9.3.2 through 9.3.4. (Forget about the small effect of melting of wall ice.)

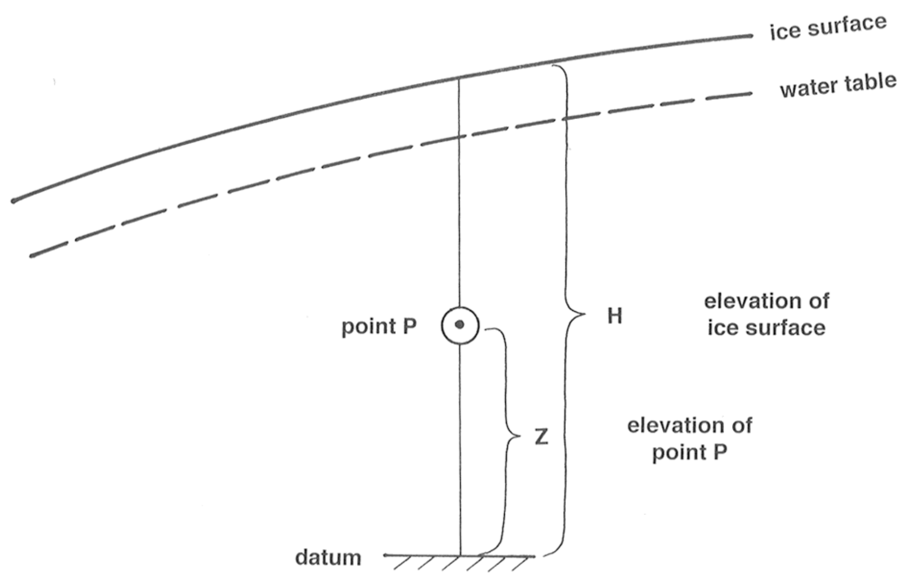


Figure 7-32. Definition sketch for analysis of the effect of water pressure and ice pressure at a point in a glacier.

4. Water tends to move through the network of passages in the direction of the gradient of a potential Φ that can be represented as a family of smoothly curving surfaces with the property that the direction of most rapid decrease in water pressure is everywhere normal to the surfaces. This is just a generalization of the idea that in a one-dimensional situation like a straight circular pipe the water moves in the direction of decreasing pressure. (It's just like the gravitational potential function that describes the direction of fall of bodies at all points near the surface of the Earth, which is the direction of most rapid increase or decrease of potential energy as you move the body up or down in the gravity field of the Earth.)

5. The potential can be represented by a equation like

$$\Phi = \Phi_o + p_w + \rho_w g z \tag{16}$$

where the first term on the right is just an arbitrary additive constant, the second term is the actual water pressure, and the third term is the hydrostatic pressure that would be produced by a column of motionless water above the given point.

6. Substituting Equations 14 and 15 into Equation 16 to get rid of p_w and p_i, and ignoring the arbitrary constant Φ_o,

$$\begin{aligned} \Phi &= p_i + \rho_w g z \\ &= \rho_i g(H - z) + \rho_w g z \\ &= \rho_i g H + (\rho_w - \rho_i) g z \end{aligned} \tag{17}$$

(Remember that the direction of drainage will be normal to the equipotential surfaces.)

7. The dip angle α of these equipotential surfaces can be found with the aid of Figure 7-33:

$$\tan\,\alpha = \frac{\dfrac{\partial \Phi}{\partial x}}{\dfrac{\partial \Phi}{\partial z}} \tag{18}$$

or, solving for the angle α,

$$\alpha = \arctan \frac{\dfrac{\partial \Phi}{\phi x}}{\dfrac{\partial \Phi}{\partial z}}$$

$$= \arctan \frac{\frac{\partial}{\partial x}\left[\rho_i g H + (\rho_w - \rho_i) g z\right]}{\frac{\partial}{\partial z}\left[\rho_i g H + (\rho_w - \rho_i) g z\right]}$$

$$= \arctan \frac{\rho_i g \frac{\partial H}{\partial x}}{(\rho_w - \rho_i) g}$$

$$= \arctan \frac{\rho_i \frac{\partial H}{\partial x}}{(\rho_w - \rho_i)} \qquad (19)$$

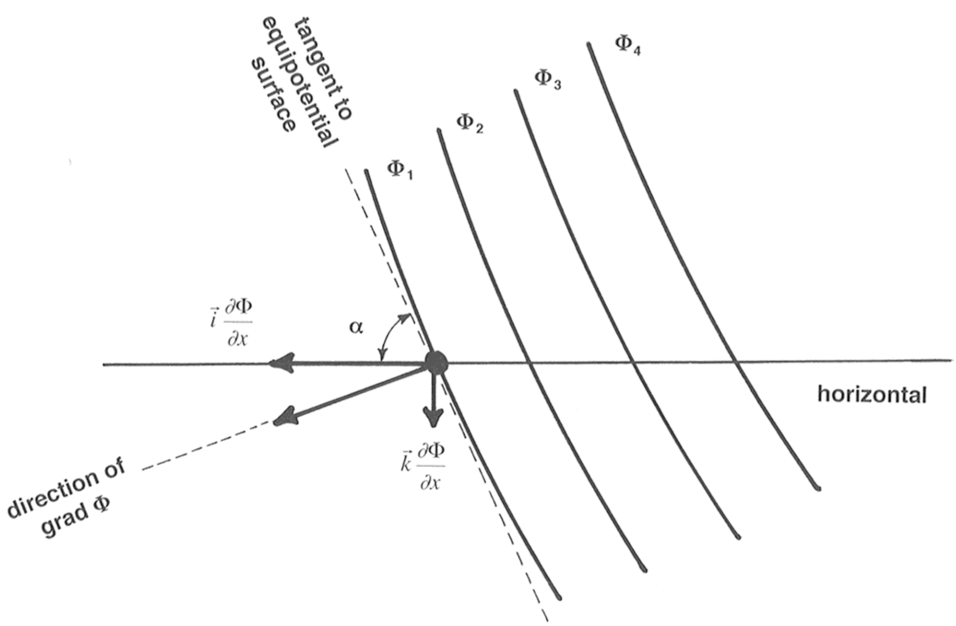


Figure 7-33. Flow of water in a glacier in response to the gradient of a potential.

Given that $\rho_i = 0.9\rho_w$, the result in Equation 19 tells us that englacial tunnels slope downglacier about 11 times as steep as the glacier surface!

8. Then what happens when the passageways reach the base of the glacier? Subglacial tunnels are constrained to follow the locus of steepest descent of the component of the potential function Φ parallel locally to the glacier bed. (Just think in terms of the curves lying on the glacier bed that are formed by the intersections of the equipotential surface with the bed, and then taking directions on the bed that are normal to those intersection curves.) On a horizontal glacier bed, this is in the same direction as the surface slope. But if the glacier bed isn't level, the tunnels can cross bedrock divides. There can even be subglacial lakes, where there are "hollows" in the equipotential surfaces (i.e., where the directions normal to the equipotential surfaces dip locally at a gentler angle than the bed of the glacier).

10. GLACIAL EROSION

10.1 Introduction

10.1.1 Glaciers are very effective in eroding, transporting, and depositing bedrock. How do we know that? There are three major lines of evidence:

- *We can see material in transport by modern glaciers.* In the most general way, where could such material have come from? Preexisting loose material (regolith; core stones; joint-bounded blocks); solid bedrock eroded by the glacier; and material that fell onto the glacier from bedrock weathering above. And we see glaciers making deposits.
- *We observe deposits that by excellent evidence were deposited by glaciers.*
- *We observe landforms that are the result of glacial erosion.*

10.1.2 The subject of glacial erosion is a difficult one. We know it happens, but it's hard to observe how it happens. Very few tunnels have been driven to the base of a glacier to watch erosion, and those haven't been representative anyway, in terms of depths and times involved. Also, very few experiments have been made. So there's a lot of deduction and speculation, and this can be very dangerous. And this is not a very graphic or photogenic topic.

10.1.3 It's generally agreed that there are two kinds of erosional activity of glaciers: ***abrasion*** and ***plucking*** (also called ***quarrying***). These affect bedrock on different scales (although there are intergradations). For each, I'll discuss evidence and possible mechanisms.

10.2 Glacial Abrasion

10.2.1 ***Tools*** (*rock and mineral particles, large and small, held in the base of the moving ice*) can abrade the underlying rock surface. Basically, this involves wearing away particle by particle.

10.2.2 What's the evidence that this happens? Mainly glacial striations and rock flour.

10.2.3 ***Glacial striations***, or ***glacial striae***, are *subparallel striations or grooves cut on the bedrock base of the glacier by tools frozen into the basal ice*. They are a very common (although by no means ubiquitous) feature of glaciated areas. They are commonly are found on rounded undulating surfaces of glacially abraded bedrock. There's a wide and continuous range in size, from microscopic, at the small end, to meters deep, meters wide, and hundreds of meters long. (The biggest grooves were probably made not by single tools but by groups of tools.) Striations are also on the larger tools themselves. Often the bedrock surface shows two or more intersecting directions, indicating either that the direction of ice flow changed or that the tools were rotated relative to direction of movement. The finest and most delicate striations are cut on soft fine-grained but nonfractured rocks like carbonates. They are coarser in medium-grained to coarse-grained rocks like granite or sandstone. Striations are fairly readily weathered, so they are best seen soon after they're made or after they're freshly uncovered of overlying sediment. Striations are known from ancient glaciations as well as Pleistocene glaciations.

10.2.4 If abrasion happens, there must be a fine-grained product. What's the nature of this product, and what happens to it? Most of what's produced by abrasion is *mineral fragments*, evidently mostly less than 100 micrometers, and predominantly fresh. This material, expressively called ***rock flour***, is largely carried out of the glacier by meltwater; glacial streams have very high suspended-sediment concentrations of grams to tens of grams per liter, and even up to a few hundreds of grams per liter, which turn the streams a characteristic dilute-milk whitish color.

10.2.5 A case can be made for the necessity of continuous removal of the fine-grained abrasion products in order for abrasion to continue; otherwise the ice–rock interface would become clogged by this stuff, like overused sandpaper. There are two ways for this to happen: it can be carried obliquely upward by flow of ice, under special circumstances of compressive flow, or more importantly, it can be washed out by meltwater. Remember that in a warm-based glacier, even if no meltwater is supplied from above, there will be a thin layer of flowing water, due to geothermal heat and friction, along the base from melting.

10.2.6 Direct observations on production of rock flour have been few and simple. One thing you might try to do is make marks or holes on the bedrock near the terminus and hope for readvance and then re-retreat within your lifetime. This points up the obvious difficulty in making systematic observations. Another more subtle difficulty is the possibility of weathering of the bedrock before readvance. The observation most often cited is some recent work that involved tunneling under a glacier and planting two rock slabs, one of marble and the other of basalt, at the base of an active glacier, retrieved after 9.5 m of passage of tool-studded ice over them. Both slabs became striated; the marble lost 3 mm of thickness, and the basalt lost 1 mm.

10.2.7 How about *the mechanism or mechanisms of abrasion*? Probably many of you have seen striations, but how are they made? By the ice dragging tools across the bedrock, you would say; yes, but under what conditions does this happen?

10.2.8 The first and most important condition for abrasion is that *the glacier ice has to be moving at the base*. That seems to mean that abrasion is important only under warm-based ice, for which there is basal slip. Cold-based ice is "glued" (frozen fast) to bedrock, and so doesn't drag tools across the bedrock. Only if there are very large tools that stick up into faster-moving ice and are rotated and pushed down against bedrock can there be abrasion in this case. Abrasion is generally conceded to be minimal in cold-based glaciers, especially when they're relatively clean.

10.2.9 But even in warm-based glaciers there's a considerable problem in keeping the tools in contact with the bedrock. Consider an isolated tool at the base of the glacier. As it rides along bedrock and exerts a normal force on the bedrock, the bedrock exerts an equal and opposite normal force on the tool (Figure 7-34). If there's basal slip in the first place, then the ice at the base is at its pressure melting point, and this means that the tool will retract slowly into the ice, and therefore stops abrading the bedrock!

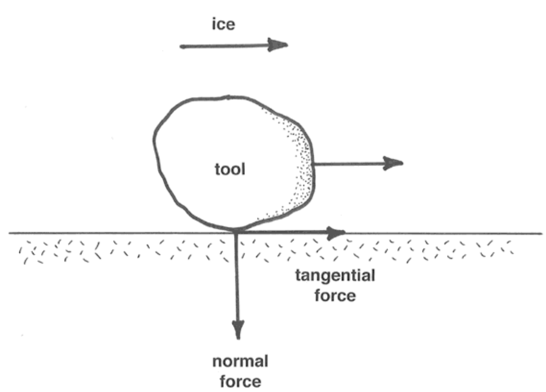


Figure 7-34. Forces and motions associated with the motion of a tool in contact with the sole of a glacier.

10.2.10 But there are invariably many tools, presumably of all sizes, packed in near the base of the glacier. Higher tools tend to hold lower tools to the base. Remember that the ice is also exerting tractive forces on the tools to keep them moving forward against the tangential resistance of the bedrock sole. Imagine a large tool, being pushed forward by the ice, and in turn pushing a smaller tool downward onto the bedrock (Figure 7-35): this produces a very large abrasive force on a small area for a long time, because the large tool retracts only very slowly, given its large area and relatively small ice–rock contact force.

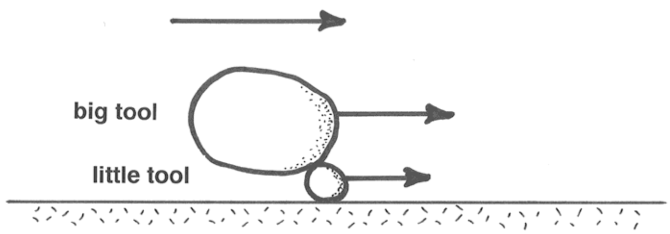


Figure 7-35. A big tool pressing upon a little tool in contact with the sole of a glacier.

10.2.11 But why wouldn't a condition eventually be reached in which all tools have attained a state of retraction? It seems as though this has to happen, eventually. So for continuing abrasion a mechanism is needed for replenishment of tools at the base of the glacier.

10.2.12 Something else to remember is that the tools themselves are abraded approximately as fast as the bedrock, if the two are of about the same resistance to abrasion. (If the tool is softer than the bedrock, the bedrock tends to abrade the tools, whereas if the bedrock is softer than the took, then the tool tends to abrade the bedrock.)

10.2.13 So tools are progressively used up, and new ones have to be supplied from somewhere. Probably the most efficacious abrasion happens in a situation where fragments are incorporated into the ice base by quarrying somewhere and then carried downglacier to abrade softer rocks.

10.2.14 The whole field of abrasion micromechanics is (as you can see) in a very speculative state. It's ripe for further observation, experiment, and theoretical work. (But there are difficulties with all three).

10.2.15 Striations tell you the orientation of the ice movement at the time they were cut. Unfortunately they don't give you the direction as well as the orientation of the movement: some are asymmetrical one way, others asymmetrical the other way. An obvious thing to do is map the orientations of striations over large areas to get an idea of the patterns of glacier movement. This has been done many times (Figure 7-36 is a good example), and the results are valuable.

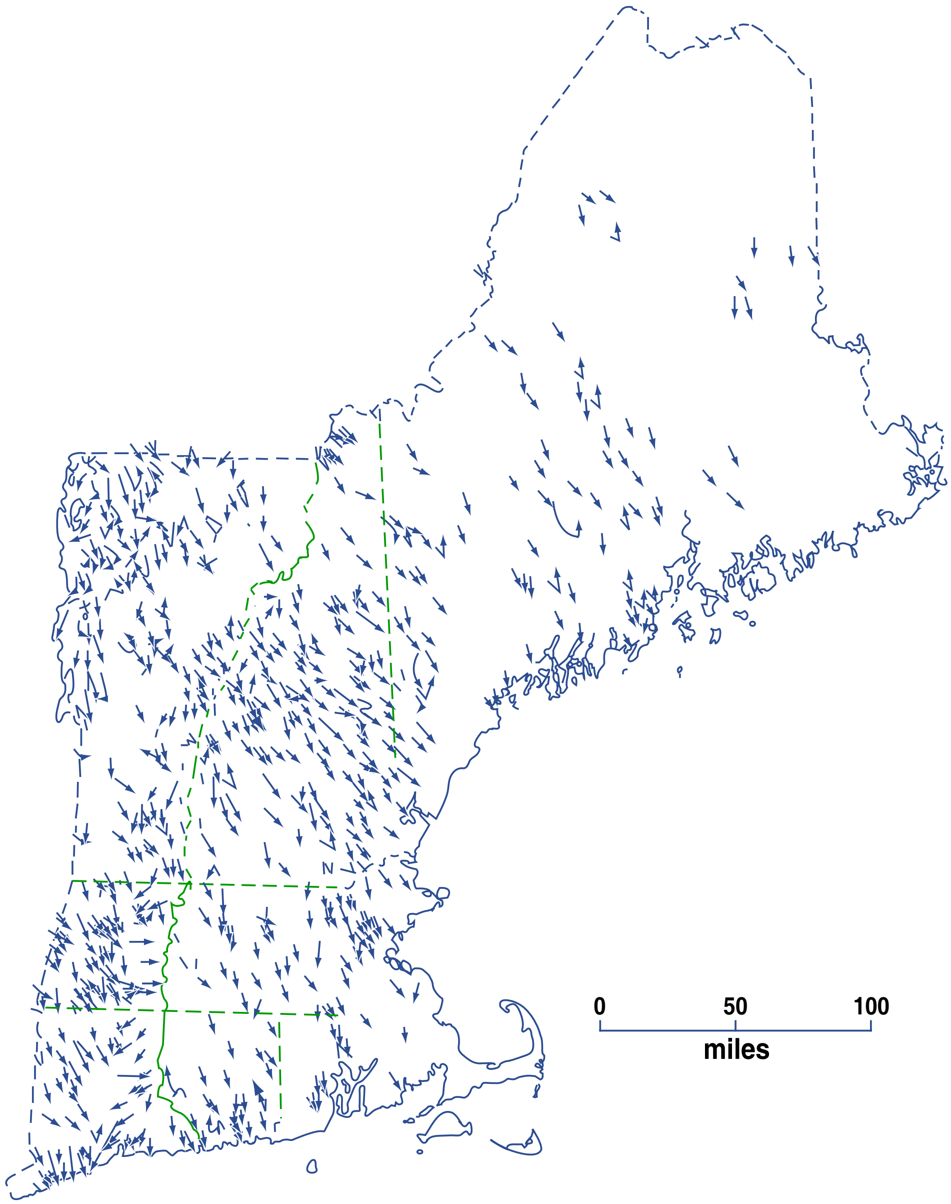


Figure by MIT OCW.

Figure 7-36. Map of glacial striae in New England, compiled from many sources. (From Flint, 1971.)

10.2.16 But there are some serious problems and limitations:

• Commonly there are very strong local variations in orientation because of local topography. So you either have to make very detailed maps or do some averaging. Striations indicate that flow can be locally at 90° to the main flow in small troughs oriented across the ice movement, or even "eddies" in cavities.

• Directions of ice movement can change with time as the geometry of ice sheets and ice caps changes as they wax and wane. In particular, there can be almost complete reversal if an outlying ice cap develops on a locally higher area during general retreat of an ice sheet (Figure 7-37). This seems to have happened in North America just south of the St. Lawrence depression.

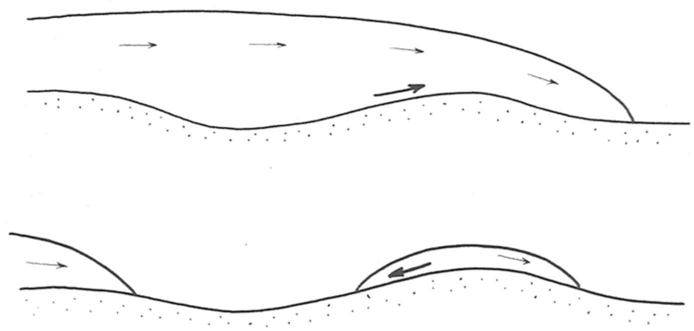


Figure 7-37. Reversal of ice-flow directions as an ice sheet retreats past a high land area.

10.2.17 A related point to remember is that striations record only the very latest abrasion, and this could be radically different in orientation from earlier, and perhaps more important, abrasion. It's probably typical that the last ice movement is unrepresentative of the main ice movement.

10.2.18 Finally, another important point is that striations on bedrock surfaces have been observed to be produced by several other mechanisms:

- drifting ice (icebergs or ice floes)
- debris flows
- snowslides and avalanches

And unfortunately you can't tell which mechanism just by looking at striations. Other evidence is usually available, though.

10.2.19 Another minor abrasional feature often discussed in the literature is ***friction cracks***. These are of several characteristic kinds, all apparently produced by the normal and tangential forces exerted cyclically by tools on bedrock at regularly spaced points in the downstream direction. These cracks occur in trains analogous to the chatter marks produced in certain machining operations, and they are probably a manifestation of stick-slip friction.

10.2.20 Some of these friction cracks show removal of material from the bedrock surface; if they are concave upglacier they are called ***crescentic gouges***, and if they are convex upglacier they are called ***lunate fractures***. Others, called ***crescentic fractures***, show no removal of material from the bedrock surface; these are almost always (or even always?) convex upglacier.

10.2.21 Controls on the geometry and spacing of friction cracks are unclear. Some experiments on scoring of optical glass with steel balls have been made to simulate, crudely, the conditions under which friction cracks are made. From these experiments it's known in a general way that crescentic fractures are produced when there's no rolling, and crescentic gouges are produced when there's rolling. In any case, it's clear that friction cracks are not reliable indicators of the direction of ice movement.

10.3 Glacial Quarrying

10.3.1 ***Glacial quarrying***, ***glacial plucking***, and ***joint-block removal*** are approximately equivalent terms for a process involving *incorporation of relatively large discrete fracture-bounded blocks into the moving ice at the base of the glacier*. Typically this process is viewed as operating on fragments that are decimeters to several meters across.

10.3.2 Glacial quarrying must be a very important mode of glacial erosion, because how else can you get all the large fragments observed to be carried in glaciers and in glacial deposits (assuming that there is a definite limit to the material initially available before the glacier covered the area)?

10.3.3 The evidence for glacial quarrying is perhaps best summarized by discussing what is usually considered to be the most characteristic form of glacial erosion: ***stoss-and-lee topography***.

10.3.4 Just as common as simply abraded surfaces is a landform that involves hillocks or knobs of bedrock on scales of meters to tens of meters to even a few hundreds of meters in plan view. These hills have a strong tendency to be asymmetrical, with gentle and streamline-molded upstream sides and steep downstream sides. The upstream sides are smoothly rounded in the large and striated in the small. The downstream sides are rough, blocky, and craggy.

10.3.5 The usual interpretation of stoss-and-lee topography is that it developed by erosion by moving ice, involving (minor) abrasion on upstream side and (major) quarrying on downstream side. Jointing patterns in the bedrock are thus important in controlling the shape of the stoss-and-lee topography.

10.3.6 Stoss-and-lee topography is less good in giving the orientation of ice movement (to maybe 10–20°) but is definitive evidence of direction and is less likely to be produced only by the last phase of glaciation.

10.3.7 There is disagreement in the literature over whether stoss-and-lee topography is a transient or steady-state aspect of glacial erosion.

10.3.8 The mechanism for glacial plucking in general is very poorly understood. An obvious requirement is the existence of fractures and joints in the bedrock. (Bedrock with tight and widely spaced joints is just not likely to be quarried.)

10.3.9 Joints and fractures can obviously predate the glacier, either because the rock was jointed long before or was fractured just before. One popular European school of thought holds that most glacial quarrying is the result of fracturing that takes place just before the arrival of the glacier; active frost wedging in a periglacial (that is, near the glacier) climate in advance of the expanding glacier prepares the way.

10.3.10 Another interesting possibility is that the production of joints is occasioned by the presence of the glacier itself. Dilatation joints are produced along surfaces congruent to the bedrock surface by unloading of overlying rock, by erosion, or by artificial means in quarries. In quarries, sometimes a rock face bursts outward, producing such a dilatation joint, because the rock has tendency to expand upon unloading. Some workers think that such dilatation joints can be produced beneath a glacier: as a glacier erodes rock and fills a deeper and deeper depression, hydrostatic pressure at the glacier base, although greater than if the glacier were not there, is much less than it was before the rock was eroded, because the density of ice is much less than that of rock.

10.3.11 What could be interpreted as good evidence of this is the existence of prominent dilatational jointing congruent to the walls of major deep glaciated valleys. (Remember that the valleys must have been produced by the glacier, and not just there from before.) Also, one can make a comparison of the stresses needed: given an ordinarily deep glaciated valley, effective unloading is more than a few hundred meters, and rock bursts resulting in dilatational joints in quarries are known to occur for unloading of less than 10 m.

10.3.12 One still needs to have joints perpendicular to rock surface. There's some suggestion from recent studies that moving glacier ice can produce its own steeply dipping joints: The data that support such an idea are from measurement of joint orientations at a great number of glaciated localities, together with the direction of ice flow from striations. Sets of shear joints and extension joints oriented symmetrically with respect to direction of movement are found. This is highly suggestive, because how could it be coincidental?

10.3.13 A related matter is that of the processes of mobilization of material in the lee of stoss-and-lee topography. What causes the quarrying there? For one thing, there's the possibility of freeze–thaw processes at base of glacier itself. Look at a typical stoss-and-lee hillock in a warm-based glacier (Figure 7-38). Local pressure at base is considerably higher over the stoss side than over the lee side. (This pressure difference has actually been measured.) So, because the pressure melting point of the ice decreases with increasing pressure, ice tends to melt on upstream side, and the meltwater produced flows around to lee side, where it refreezes in the region of lower pressure. If this happens on a larger scale, it could wedge out blocks on the steep lee side of the hillock, given the preexistence of fractures.

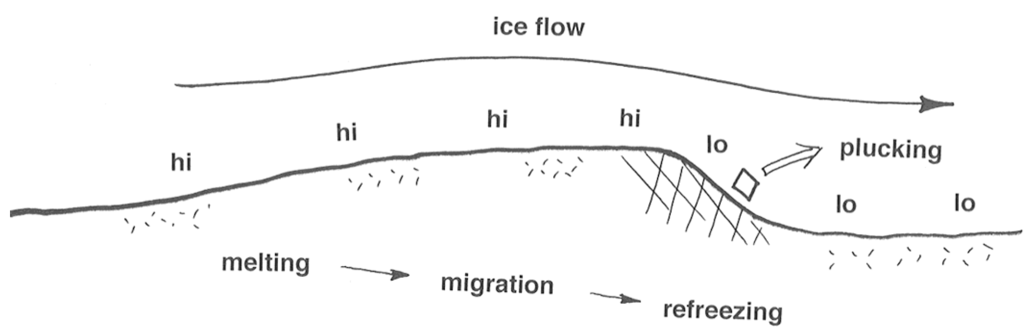


Figure 7-38. A possible mechanisms for plucking, based on melting and refreezing in response to change in pressure over a bedrock knob at the base of a warm-based glacier.

10.3.14 Which is the more important quantitatively, abrasion or plucking? It's generally agreed that plucking is more important than abrasion. But the basis for this view is not extensive.

10.3.15 The classic study was done close to home by Jahns in 1943 on glaciated granite hills of eastern Massachusetts. Well developed sheeting joints presumably predate glaciation and show the original form of the hills. The present form of hills relative to original form shows that there has been much more glacial erosion on the lee sides, by plucking, than on the stoss sides, by abrasion (Figure 7-39).

10.3.16 Another important line of evidence lies in the long profiles of glaciated valleys. Characteristically such profiles show large-scale steps, and even excavation of deep rock basins with up to several hundred meters of closure, now filled with glacial debris or lakes (Figure 7-40). On the assumption of regular preglacial river-valley profiles (a very good assumption), this shows that erosion rate varies greatly from place to place. Also, places where rock basins are excavated tend to have well jointed and therefore easily quarried material, whereas the bedrock highs don't, and are therefore subject only to abrasion.

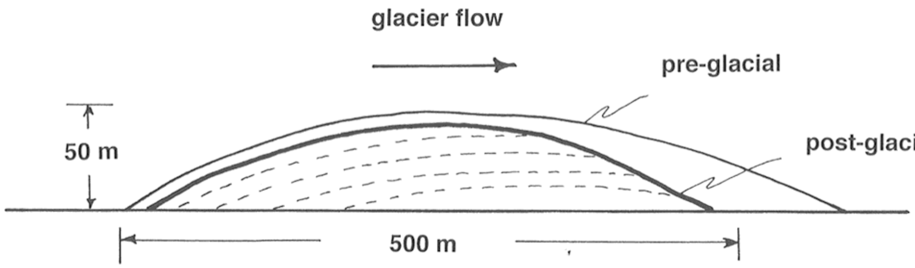


Figure 7-39. Streamwise cross section through a glaciated bedrock hill, showing the relationship between sheeting joints and land-surface topography.

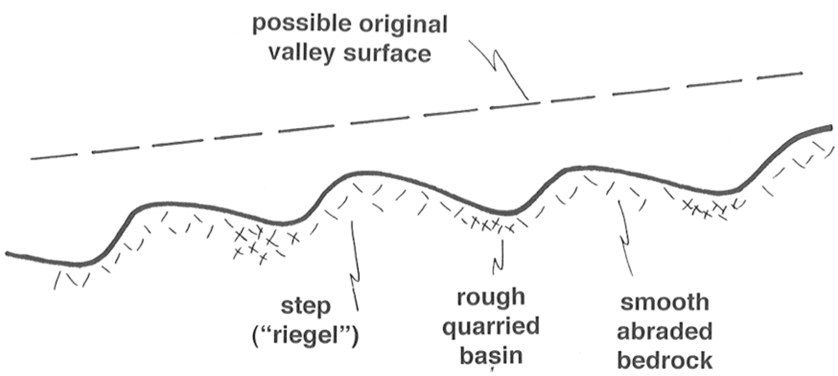


Figure 7-40. Steps and basins along the longitudinal profile of a typical glaciated valley.

10.4 Rates of Glacial Erosion

10.4.1 It would be nice to be able to make an estimate of the rates of glacial erosion, and then to compare these with rates of fluvial erosion. Much effort has been expended in this direction.

10.4.2 There have been four major approaches to this problem:

- Erosion of artificial markers placed beneath a glacier (already mentioned). Problems with this approach:
  - — Time scales are too short.
  - — What about pre-weathering? (But this is a problem only if the marker is placed in an ice tunnel.)
  - — What about spatial non-uniformity? (The sampling grid would have to cover a large area.)
  - — This measures abrasion only, and by the nature of abrasion this involves too small a measurement area.
- Measure sediment transport in subglacial streams emerging from beneath a warm-based glacier. Problems with this approach:
  - — One needs to accumulate very long-term records and make closely spaced observations, because such streams are notoriously highly variable.
  - — It's difficult (or impossible) to measure bed load.
  - — One has to look at moraines too; not all debris goes out in streams (but probably usually most, except in cold glaciers).
- Reconstruct the preglacial surface. Problems with this approach:
  - — How do you know what the preglacial surface was? There's lots of disagreement about the criteria to use. (This approach works best in glaciated valleys, but there are problems even there.)
  - — Effects of glacial erosion are hard to factor out from concurrent fluvioglacial erosion and fluvial erosion during interglacial periods.
- Compute volumes of glacial drift. Problems with this approach:
  - — One needs to work in a very large area.
  - — Glacial drift grades out into fluvial and marine deposits; how do you separate out the glacial component in those deposits?
  - — It's not always easy to identify the form and depth of the base of the drift.
  - — Later erosion of drift deposits must be taken into account.

Method 2 is probably the best, but it works only for temperate (warm) glaciers.

10.4.3 Some very general conclusions:

- Active temperate valley glaciers erode at much greater rates than rivers in similar but nonglaciated areas.
- Ice sheets (warm or cold) moving slowly over low-relief areas produced relatively low erosion even over all of the Pleistocene.

10.4.4 Example: the Antarctic ice sheet seems now to be eroding very slowly. Erosion probably reached a peak early, well before the Pleistocene, when the glaciers were more active and there was more material available, and it's now slow, because of lack of easily erodible material, inactive regimen, and cold base.

10.4.5 Here's some discussion on two additional specific topics in glacial erosion: *glaciated valleys* and *cirques*.

10.4.6 Glaciated valleys. In mountains that are not mainly the work of glaciers, preexisting valleys are modified during glaciation. In low-relief regions covered by ice sheets, often prominent valleys (now partly filled by sediments or lakes) have less obvious relation to preexisting drainage; their position is determined by preexisting drainage, but valleys become more differentiated as the glacier exploits weak rock. The Finger Lakes of New York State are an excellent example of the latter effect.

10.4.7 The transverse profile of glaciated valleys is commonly U-shaped: very steep or nearly vertical side walls, and gently rounded bottom. In some cases there's even a good parabolic fit. Nonglaciated (stream) valleys in mountainous regions, on the other hand, are commonly V-shaped. Glaciated valleys have much steeper sidewalls. See Figure 7-41.

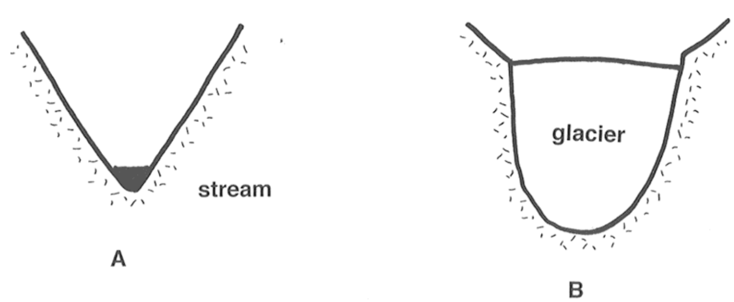


Figure 7-41. Transverse cross sections through A) a nonglaciated valley (V shaped) and B) a glaciated valley (U shaped).

10.4.8 The change from a V-shaped stream valley to a U-shaped stream valley could simply be by widening, or by both widening and deepening. There seems usually to be both widening and deepening. Evidence:

- Hanging tributary valleys (Figure 7-42), which even if glaciated would have far less glacial erosion. Very common.
- The long profile shows steps and rock basins, already illustrated.

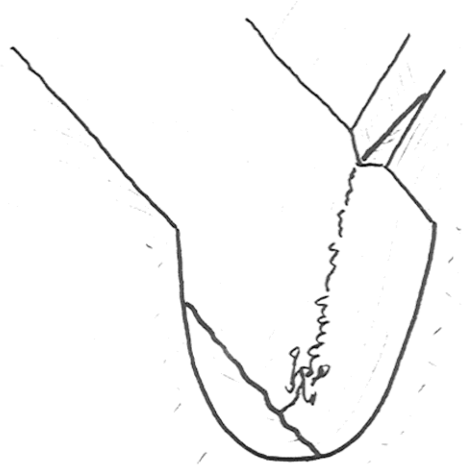


Figure 7-42. Cartoon of a hanging alley.

10.4.9 *Why does a glacier develop a U-shaped valley?* This is difficult to answer. Two possibilities are commonly mentioned in the literature:

- Concentration of nivation (see below) in the lower parts of valleys, as the approaching glacier weakens rocks for removal.
- Development of dilatation joints, especially in lower parts of the valley, where the ice is thicker.

Both of these effects would tend to act more in lower parts of the valley profile.

10.4.10 Cirques. A ***cirque*** is *a rounded basin, partly enclosed by steep cliffs, cut into a mountain slope*. Cirques are about circular, or at least fairly equidimensional, in plan. Their size ranges from tens of meters to several kilometers across. They can be located either at the heads of glaciated valleys or independently as indentations in smooth slopes. Cirques may or may not have (or have had) a glacier in them.

10.4.11 Small cirques can be formed without the presence of a glacier (Figure 7-43). Think about a firn bank that occupies a slight preexisting depression or a shady spot. Freeze–thaw cycles in the warm season result in frost wedging, and then this material moves downslope by a combination of mass wasting and surface runoff (This *combination of freeze–thaw and mass wasting* is called ***nivation***.) This results in a small cirque. In such a small cirque, a firn bank acts as a passive water source.

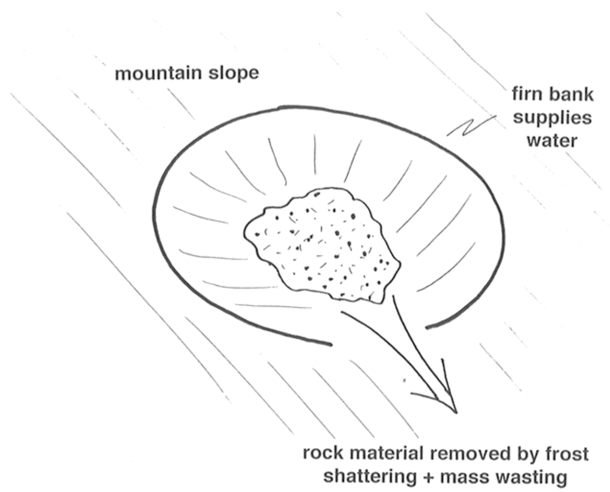


Figure 7-43. Early stage in the development of a cirque.

10.4.12 If the firn bank gets big enough, a small glacier (called a ***cirque glacier***) is formed. Then the moving ice can enlarge the cirque to much greater size. Larger cirques tend to have a bedrock basin (often occupied by a lake after the ice melts, called a ***tarn***, and a sill at the downslope edge (partly rock, partly moraine). Of course, large cirques grade over into the steep amphitheater-like heads of major glaciated valleys. Cirque glaciers are typically half-moon-shaped, and they move in a characteristically rotational way (see Figure 7-44, which shows the results of detailed study of a conveniently small and fairly regular cirque glacier in Norway). Note the almost circular-arc base, the almost planar surface, and the strongly rotational movement. The glacier consists of a series of ice layers separated by discontinuities (ablation textures and mineral-organic dust coatings). Most of the movement is accounted for by rotational sliding around a horizontal axis, but there is some deformation of annual layers as well.

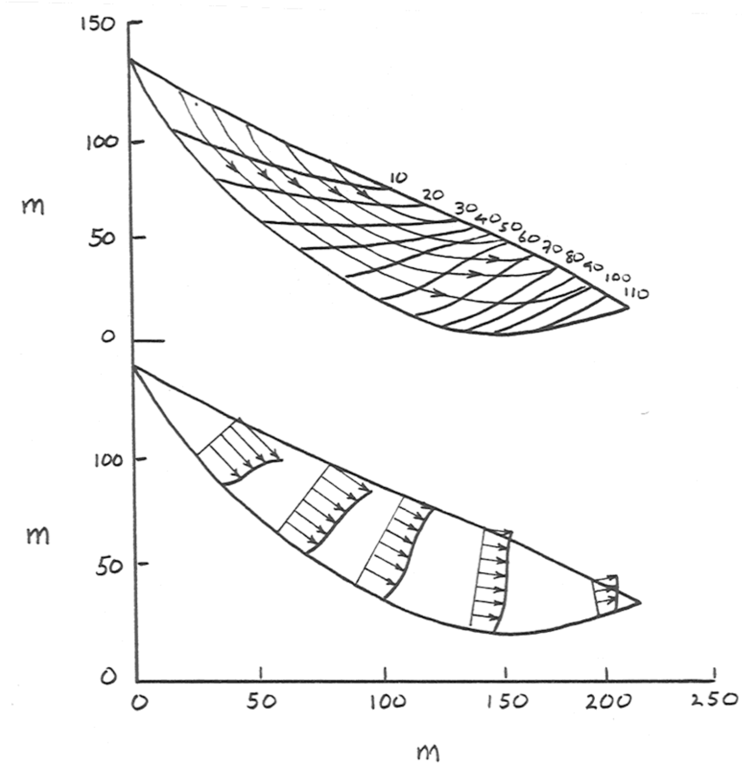


Figure 7-44. Streamwise vertical cross section showing flow lines and velocity profiles in a small cirque glacier.

10.5 Alpine Sculpture

10.5.1 Glaciation in mountainous areas tends to produce characteristic glacially sculptured landforms. The assemblage of such landforms is usually termed ***alpine sculpture***.

10.5.2 As the climate in a rugged mountainous area gradually changes to become more favorable to the development of glaciers, small cirque glaciers form first and then expand to become valley glaciers. The heads of the valley glaciers may expand and merge to form extensive snowfields. Eventually an ice cap covers most or all of the mountainous area, extending out into the lowlands beyond. Then, as the climate ameliorates again, the process operates in reverse, and the glaciers shrink back to valley glaciers, then to cirque glaciers, and finally disappear altogether.

10.5.3 Figure 7-45 shows the sequence of glacially sculptured landforms associated with the cycle of glaciation outlined above. During the expansion of the glaciers, valleys are deepened and widened, leaving U-shaped valleys often separated by sharp ***cols*** and ***arêtes***. Three or more large cirques at the heads of glaciated valleys may meet to for pyramidal faceted peaks known has ***horns***.

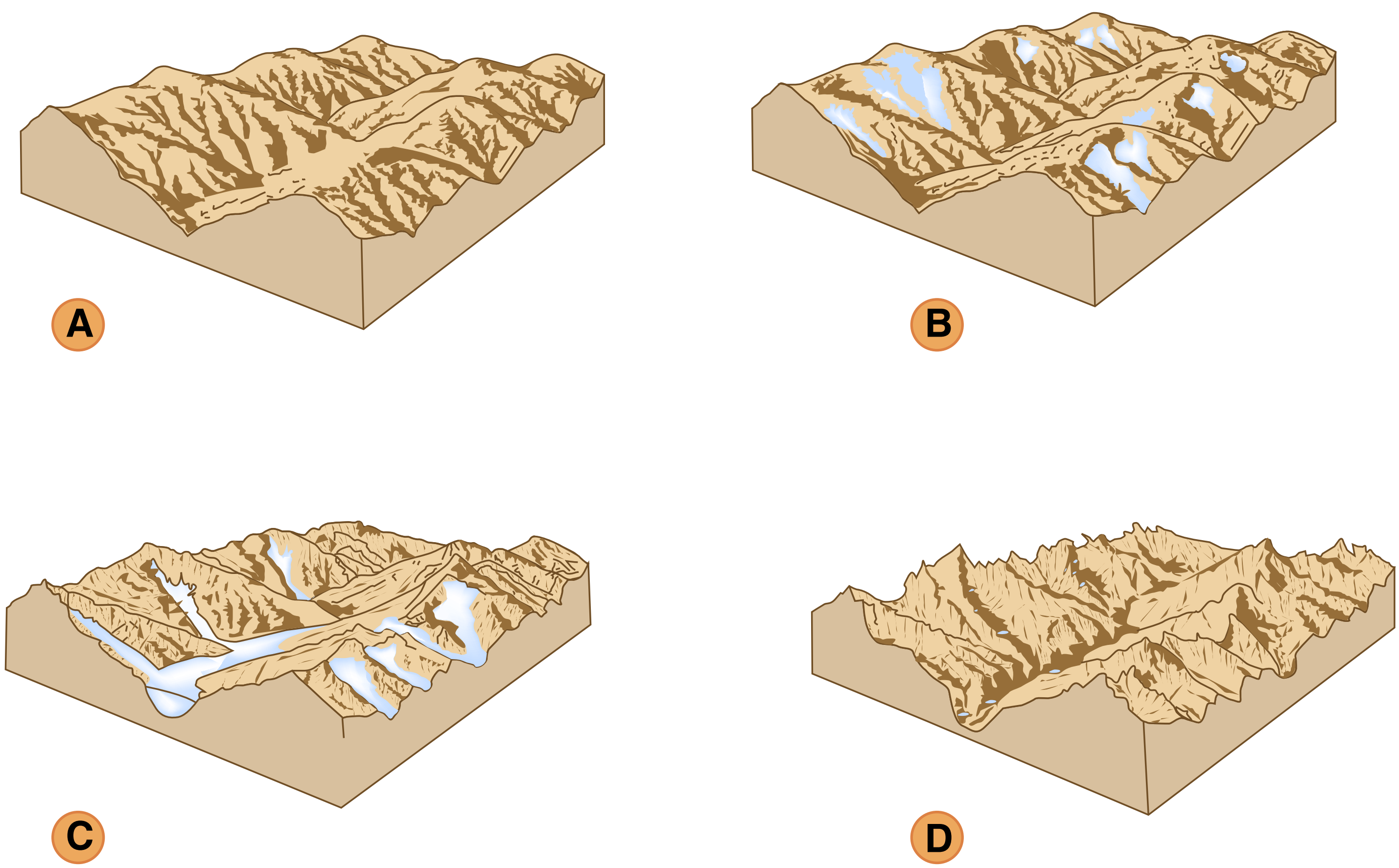


Figure by MIT OCW.

Figure 7-45. Sequence of glacially sculpted landforms associated with the cycle of glaciation and deglaciation of a mountainous area.

10.5.4 At the time of maximum glaciation, much or all of the former sharp glacially sculptured topography is smoothed as the entire area is worn by the moving ice of the ice cap. Then, as the ice caps shrink back to valley glaciers and then cirque glaciers, the mountains are sculpted once again into the landforms noted above.

11. GLACIAL SEDIMENT TRANSPORT

11.1 The term ***load*** is used for *all drift that's in transport by a glacier at a given time*. As with fluvial sediment transport, keep in mind the distinction between the load and the transport rate. The ***transport rate*** of a glacier is *the time rate of passage of sediment past some cross section through the glacier that's stationary relative to the underlying bedrock*.

11.2 The load is generally classified on the basis of where it's transport in the glacier:

- **supraglacial load:** load transported on the surface of the glacier. It gets there by falling onto the glacier. It's thus restricted mainly to valley glaciers. But near the terminus of the glacier, drift can reach the surface by ablation of the ice, and also by upthrusting.
- **englacial load:** load transported within the glacier. The quantities of englacial drift are always much smaller than of supraglacial drift and subglacial drift, because the load is obtained by the glacier at the bottom and the top of the glacier. Solids deposited on the surface of the glacier in the area of accumulation are buried and thus become englacial drift.
- **subglacial load:** load transported at the base of the glacier. This constitutes most of the load, just because most material is entrained at the base. Cold-based glaciers have little subglacial load. The dirtiest warm-based glaciers might have concentrations of up to tens of percent by volume, for thicknesses of a few meters above the base. The way this is known is by examination of now-dead ice at a stagnant terminus.

12. GLACIAL DEPOSITS

12.1 Introduction

12.1.1 Here are some comments on the nature of sedimentary materials deposited by glaciers:

• It *tends to be fresh*. After a glacier picks up the loose preexisting material, it wears away fresh bedrock. And there's no weathering while the material is in transport. So the material deposited beneath the glacier, or dumped at the terminus, tends to be fresh. (But the deposit itself can undergo subsequent weathering.)

• It faithfully reflects the composition of the upglacier source rocks.

• It *varies widely in composition*, because it reflects the composition of the bedrock directly upglacier, which of course can be anything. It consists of both mineral grains and rock fragments.

• It *tends to be poorly sorted*. A glacier is indiscriminate in terms of the particle sizes it carries, so deposits directly from the glacier are likely to be very poorly sorted. Glacial deposits are among the least well sorted of all sediments. But if the material is reworked by water or wind, it can end up being fairly well sorted. (But it's still called glacial sediment if it's recognizably related to glacial action.)

• Particle shape is sometimes characteristic: multifaceted "flatiron" shapes are common among the larger, gravel-size clasts. This characteristic shape is caused by abrasion while in successive orientations as a tool at the base of the glacier. Only some, not all, of the large clasts show this characteristic shape.

• The larger clasts are often striated, just like the underlying bedrock. Beware, however, that other agents of transport (like debris flows) can produce striations on gravel-size clasts.

12.2 Notes on Classification of Glacial Deposits

12.2.1 It's important to understand that classification of glacial deposits is twofold:

- texture of material in general
- form or geometry of deposit

Composition is typically not taken into account in classification.

12.2.2 There's a partial but not nearly total correlation between these two bases for classification. It's common to have the same textural kind of deposit involved in more than one morphological kind of deposit, and the same morphological kind of deposit showing more than one textural type—thus the need for separate ways of looking at glacial deposits. (One could argue that this could be the case for sediments in general, but usually we can't study them together as suitably as in glacial deposits.)

12.3 Glacial Drift

12.3.1 Glacial sediments have long been called ***drift***. That word dates from before the glacial theory, when it was thought that this characteristic material was deposited by flowing water ("drifted" in). Glacial drift is *all material in transport by glacier ice, all deposits made by glacier ice, and all deposits predominantly of glacial origin even though not deposited directly by a glacier*. Glacial drift is thus highly varied in texture, composition, deposit morphology, and origin. In the following I'll discuss the nature of the material first, and then the deposit geometry later.

12.3.2 Glacial drift is usually subdivided into two major categories*:* ***till***, on the one hand, and material variously called ***stratified drift***, ***washed drift***, or ***sorted drift***, on the other hand (Figure 7-46). (A really good term for this second kind of drift has not yet been invented.)

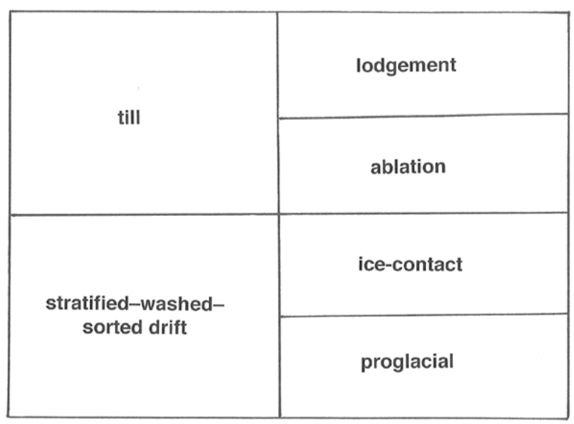


Figure 7-46. Classification of glacial drift.

12.3.3 The distinction between till and s/w/s drift is basically *descriptive*: till is poorly sorted and largely nonstratified, and s/w/s drift is much better sorted and characteristically well stratified. But the distinction is universally viewed as having a clear *genetic* basis: till is deposited directly from glacier ice, without the effect of flowing meltwater, whereas s/w/s drift is material that has been picked up by flowing meltwater and redeposited somewhere else.

12.4 Till

12.4.1 General

12.4.1.1 ***Till*** is a genetic term applied to *all unstratified and unsorted deposits made directly by or from glacier ice*. The lithified equivalent of till, as seen in the ancient sedimentary record, is called ***tillite***.

12.4.1.2 Till is subdivided in turn into lodgement till and ablation till (Figure 7-47). ***Lodgement till*** is *deposited directly from the moving ice beneath the moving glacier*, whereas ***ablation till*** is *deposited at the glacier terminus as the ice melts and drops its load.* The distinction between lodgement till and ablation till is thus a genetic one.

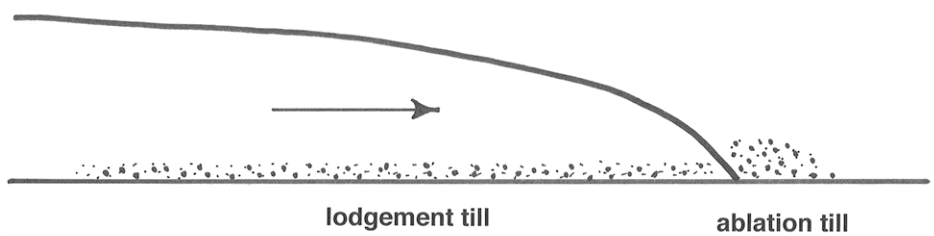


Figure 7-47. Lodgement till and ablation till.

12.4.2 Lodgement till

12.4.2.1 Classic lodgement till is a very characteristic sediment:

- The particle size ranges from large boulders continuously down to clay-size material (mainly ***rock flour***: *ground-up very-fine-grained mineral material*) Tills are often described as *boulder clays*.
- Typically the larger clasts "float" in the matrix, so the sediment could be called *matrix-supported*.
- There's usually *no stratification*: till is typically a structureless deposit.
- Often there's a subtle statistical *preferred orientation*, or fabric, especially of the larger clasts; the long axes tend to be oriented parallel to glacier flow.
- The sediment tends to be *cohesive* even if it's young; the degree of compaction is such that often the stuff has to be dug with a pick instead

of a shovel. Tills, even young ones, are often lithified enough to be jointed.

- It has low porosity and very low permeability.

12.4.2.2 In terms of genesis, it seems clear that till like that described above is deposited particle by particle from the base of an active glacier. Till known or thought to have been deposited in this way is called ***lodgement till***; remember that this is a genetic term.

12.4.2.3 Apparently it's common in the peripheral parts of an ice sheet for the glacier to be overloaded at its base, so instead of the glacier picking up material, it's depositing material. But the relative importance of the various deductively reasonable mechanisms of lodgement are largely unclear.

12.4.2.4 Here are a few comments on the mechanics of subglacial deposition. Three mechanisms of subglacial deposition can be envisioned:

- Pressure melting at the base of the sediment-laden ice releases particles at a rate faster than the rate at which comminution by abrasion plus removal by meltwater can operate, and the excess particles left beneath the ice are plastered onto the depositional surface.
- Plastering happens because the frictional drag on sediment particles in movement over the sediment becomes equal to the tractive force of the glacier ice. This might happen where the ice is decelerating and the friction is decreasing.
- Gradual reduction of basal ice velocity leads to shearing over successive layers of ice–debris mix at the base, and then the water is eventually removed by melting.

12.4.2.6 There have been just a few direct observations of lodgement processes at the base of a glacier. These indicate that at least the first two processes actually happen. And there's no reason to believe that the third can't happen as well.

12.4.2.7 Another important thing about lodgement till is that under some conditions it can undergo flow by shearing after it's deposited. The moving glacier above the sediment exerts a shearing force on the upper surface of the deposit, and if the shearing force exceeds the shear strength of the material, the till flows. Here's a qualitative treatment of the forces involved in such flow.

12.4.2.8 The shear strength of a continuous granular medium like till is a reflection of the frictional resistance the material affords to shearing. Think of the resistance to shearing as a matter of friction: the frictional stress, which tends to make the material shear along shearing planes, can be viewed in the classical way as the product of the normal stress exerted on the plane of shearing and the coefficient of sliding friction. The high water content together with the low permeability of typical lodgement till beneath the glacier means that the effective normal stress on the shearing planes is greatly reduced by the water pressure in the pores, because the water pressure lessens the particle-to-particle contact forces. If that doesn't make sense to you, just think in terms of *buoyancy* or *submerged weight*: fill a drinking glass with loose sand and think about how much the grain-to-grain contact forces are lessened when you fill the glass with water interstitial to the sand grains.

12.4.2.9 The slow flow of subglacial lodgement till has actually been observed directly in a few cases. All you have to do is drive a tunnel under the base of the ice (and keep it open!), and then drive a segmented rod down into the till. Come back sometime later, excavate the rod, and see how its little segments have moved in the downglacier direction. Figure 7-48 shows an actual measurement of this kind.

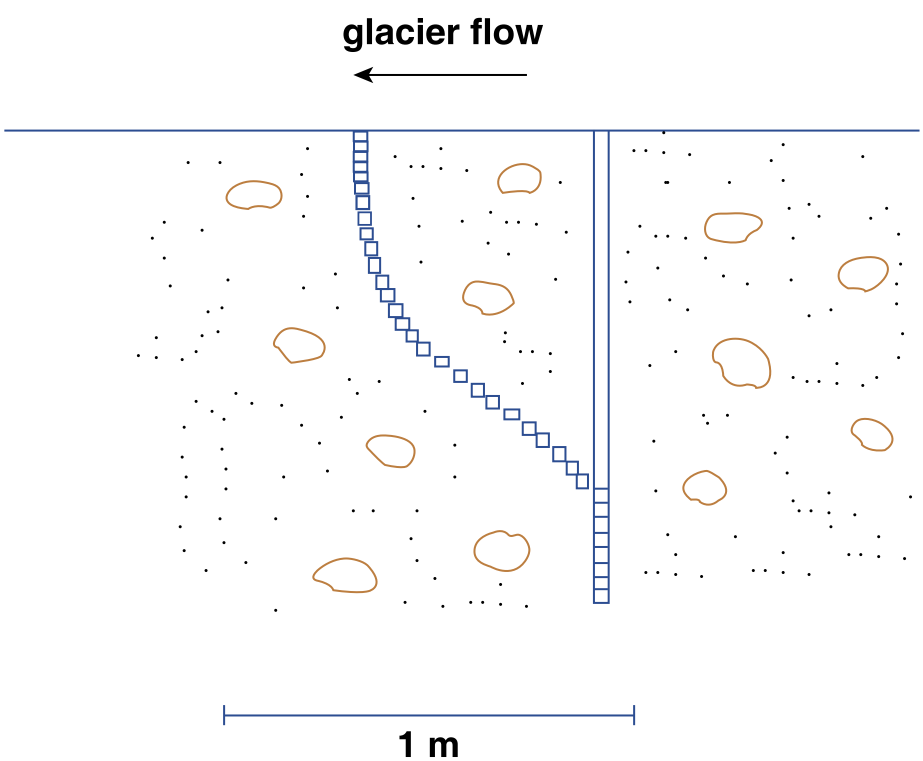


Figure by MIT OCW.

Figure 7-48. Slow flow of subglacial lodgement till.

12.4.3 Ablation till

12.4.3.1 The other kind of till, called ***ablation till*** (also a genetic term, remember) is easier to understand. It's *deposited right at the terminus of the glacier, in the process of ablation there*, not underneath the glacier. This is one of the things that can happen to the load of the glacier that's delivered all the way to the terminus rather than being lodged beneath the glacier.

12.4.3.2 Although I've never seen any definitive pronouncements, my guess is that more sediment is delivered to the terminus than is extracted from the glacier from below, upglacier of the terminus. This material is simply released from the glacier ice as the ice melts—hence the term ablation till.

12.4.3.3 The important thing about ablation till is that it's not as rich in very fine material as lodgement till, because the liquid water melted out of the glacier tends to carry that material away in suspension, leaving behind the coarser material, of gravel, sand, and silt size (along with some fraction of even the finest material as well). So ablation till is more friable and easier to dig than lodgement till. It's not difficult to tell the two apart in this way.

12.4.3.4 You may be thinking that what I've said about ablation till makes it seem closely related to, or indistinguishable from, what I called washed drift. The important point is that the material that was deposited to form the ablation till—what's there for you to see—was deposited directly from glacier ice; only the finest fraction is winnowed by the meltwater and deposited somewhere presumably far away. Of course, there are likely to be gradations between ablation till and washed drift, depending on how much meltwater happens to be around and flowing in the given locality, and therefore on how much of the coarser material is transported and redeposited.

12.4.4 Two-Layer Till Sheets

12.4.4.1 From what was said above about the deposition of till, it should make sense to you that till typically appears as extensive sheets and blankets that mantle the landscape continuously or semicontinuously. (The mantle of till tends to be thicker in valleys than on hills, and thicker on the lee sides of hills than the stoss sides.) The thickness of till sheets ranges from less than a meter to many meters, and in some cases even tens of meters. Till sheets take on a wide and interesting variety of geometrical forms; I'll elaborate in a later section.

12.4.4.2 Commonly, though not nearly always, one sees a ***two-layer till sheet***: *a lower layer of lodgement till, and an upper layer of ablation till.* The upper layer of ablation till tends to be spottier and less continuous than the lower layer of lodgment till, but the two-layer structure is nonetheless characteristic.

12.4.4.3 It's easy to understand how the two-layer arrangement of till comes about (Figure 7-49). As the glacier advances, it's likely to form an underlying sheet of lodgement till. Then, during retreat, ablation till deposited at the retreating terminus of the glacier is spread backward as a sheet across the underlying sheet of lodgement till.

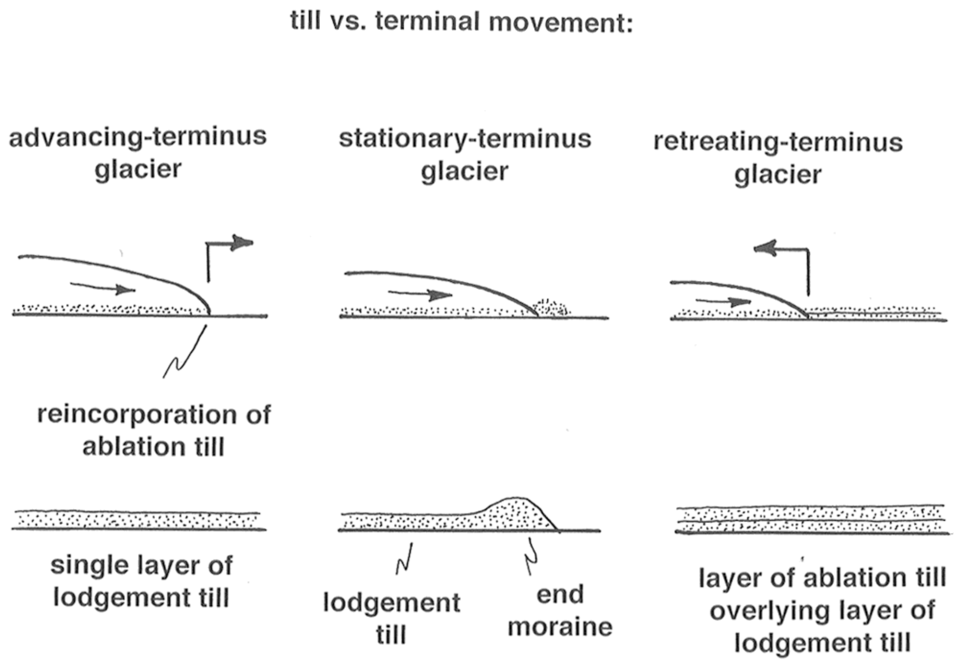


Figure 7-49. Relationship between till deposition and terminus movement.

12.4 5 The Morphology of Till Sheets

12.4.5.1 When till sheets are viewed morphologically or topographically, they are usually termed ***ground moraine*** (note the without-the-article usage).

12.4.5.2 The topography of ground moraine varies widely (Figure 7-50). Often one sees a fairly regular series of ridges and furrows oriented parallel to the direction of ice flow. The relief of these ridges and furrows varies from less than a meter high and only a few hundreds of meters long, to giant features, up to a few tens of meters high and a few tens of kilometers long. The ridges are strikingly straight and parallel. This kind of topography can best be appreciated from the air or on air photos. Ground moraine showing such ridges and furrows is called ***fluted moraine*** or ***fluted ground moraine***.

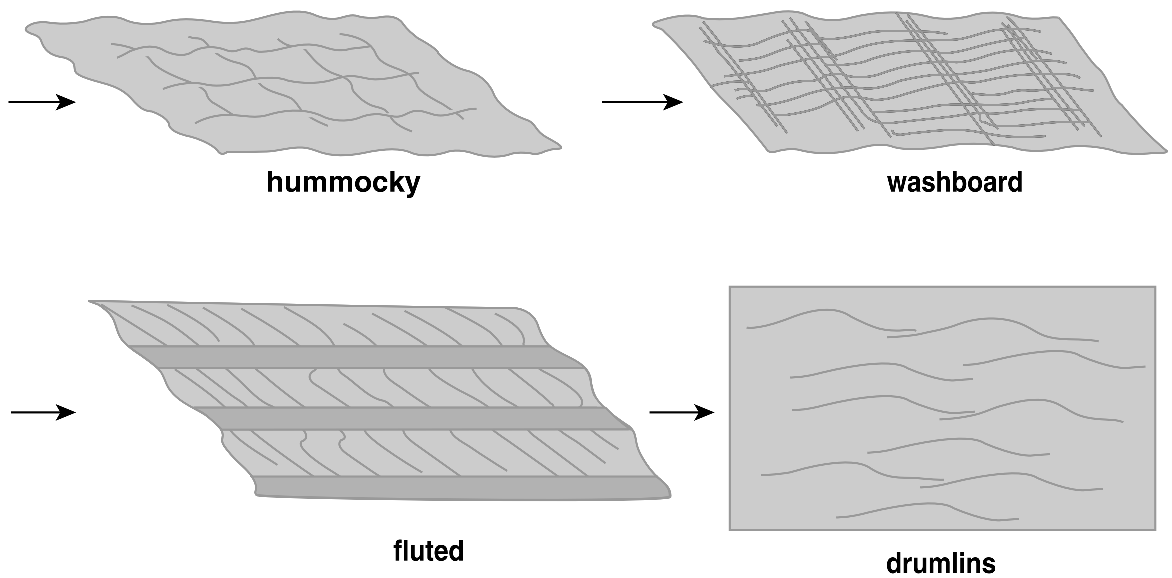


Figure 7-50. Varieties of surface topography of ground moraine.

12.4.5.3 How is the ridge-and-furrow topography produced? Most theories rely on lateral flowage of till beneath the glacier, somehow. The idea here is that till moves laterally in a semiliquid or plastic state subglacially in response to subglacial hydrostatic pressure field. For example, low pressure in the lee of boulders or rock knobs should produce flow into that area, and these low-pressure areas might tend to extend themselves downstream. All such explanations are highly speculative.

12.4.5.4 ***Drumlins*** are another very characteristic kind of ground-moraine topography. You've all seen drumlins: *glacially streamlined hills*, ideally almost half-ellipsoidal in shape. Drumlins are found in large numbers in certain areas, called ***drumlin fields***. Drumlins can be cored by either till or bedrock hills. Often the drumlin consists of a streamlined tail of till extending downstream from a bedrock knob. There's a whole spectrum of features intermediate between discrete and rather blunt drumlins to long and regular longitudinal ridges and furrows. Often what are called drumlins are very elongated hills that might better be called longitudinal ridges.

12.4.5.5 A great many theories on the origin of drumlins has been proposed, but none of them has become really widely accepted. The problem is the classic one that we can't observe them in the process of formation. Actually drumlins may be a polygenetic phenomenon. The usual approach is to invoke some kind of depositional response to subglacial pressure differences, either by flowage of till or preferential lodgment. Clearly a dynamical instability of some kind must be involved, or else the till surface would remain planar. But this kind of approach doesn't seem to be appropriate or necessary to explain rock-cored drumlins with just a veneer of till, especially if the drumlins are more or less isolated.

12.4.5.6 Just to confuse you further, let me point out that some kinds of ground moraine show transverse topographic elements. Such ground moraine is usually called ***ribbed moraine*** or ***washboard moraine***. It's clear that such topography is produced subglacially, not at the glacier terminus. But again, understanding is in a poor state: there are lots of theories, all of them speculative.

12.4.5.7 Finally, it's only fair to point out that much ground moraine shows no strongly organized topographic elements, just a seemingly random collection of high and low areas, often rather subdued. Apparently in many cases none of the various mechanisms leading to flow-parallel or flow-transverse features are at work.

12.5 Stratified Drift

12.5.1 Introduction

12.5.1.1 The term ***stratified drift*** refers to glacial drift that has been reworked by glacial meltwater and then deposited either in direct contact with glacier ice or at some point more or less far away from the glacier, in a wide variety of depositional environments. In a kind of order-of-magnitude sense, stratified drift is just as common and important as glacial till.

12.5.1.2 The term stratified drift is not an especially good one. What's called stratified drift is indeed usually prominently stratified, but the term doesn't do well at capturing the essence of the stuff. Alternative terms, perhaps somewhat better but not a lot better, are ***sorted drift*** and ***washed drift***.

12.5.1.3 Here are some miscellaneous important points about stratified drift:

- It's mostly sand or sand plus gravel; the fines are carried away in suspension.
- It occurs in large to small but basically *isolated deposits*. Sometimes it's in two-dimensional belts, but it's never in extensive sheets like till sheets. The main reason is that streams are by their nature localized.
- Stratified drift commonly forms deposits with considerable *relief*, so it's often seen as hills.

- It almost always shows excellent and striking *stratification*, because it's laid down by flowing water stratum by stratum, and conditions of deposition usually vary with time.
- It commonly shows *cross-stratificatio*n on various scales, because of the existence of ripple and dune bed configurations under a wide variety of conditions of flow and sediment.
- It commonly shows *deformation* features, because it's often deposited against steeply sloping surface of ice, and, when the ice inevitably melts, the sediment slumps and slides.
- It's an important *source of sand and gravel* for making concrete (in glaciated regions that is; in nonglaciated regions, sand and gravel is usually a lot harder to come by).

12.5.1.4 Watch for sand and gravel pits and quarries for examination of cuts through stratified drift. The trouble is, the good viewing doesn't last for long.

12.5.1.5 In this section I'll deal mostly with the mechanics of transport and deposition of stratified drift, and the characteristic form or geometry of the drift deposits.

12.5.2 Classification

12.5.2.1 It's not easy to classify stratified drift, because there's basically a continuous gradation among the various types. But there are some characteristic types, and we'll deal with these.

12.5.2.2 There are two major categories of stratified drift:

- **ice-contact stratified drift:** deposited in direct contact with, or at least in the immediate proximity, of, glacier ice, active or inactive.
- **proglacial drift:** drift that is carried out well beyond the glacier terminus by flowing water. This passes over into sediment that is not recognizably of glacial origin.

12.5.3 Ice-Contact Stratified Drift

12.5.3.1 First off, I want you to have two rather different mental pictures of a retreating glacier, which is ultimately going to leave the stratified-drift deposits we see (Figure 7-51). I'll give call two contrasting modes ***active-ice retreat*** and ***stagnant-ice retreat***. (These terms are expressive but not really official.)

• **active-ice retreat:** there's a fairly steep ice slope in the downglacier area; the ice flows actively all the way to the terminus; there's a well defined terminus even though the glacier is retreating; meltwater discharges directly off or from the glacier and forms a well defined stream, except when the terminus lies in a glacial lake.

• **stagnant-ice retreat:** the glacier surface has a fairly gentle slope near the terminus; excessive ablation reduces the thickness of the glacier over a substantial distance, thereby giving rise to a broad belt at the downstream end of the glacier where the ice can no longer flow; the dead melting ice then sits around over a large area, and the active terminus is now located far upglacier; meltwater streams flow between, around, under, and over this dead ice, depositing stratified drift, and the dead ice sheds sediment as it melts as well.

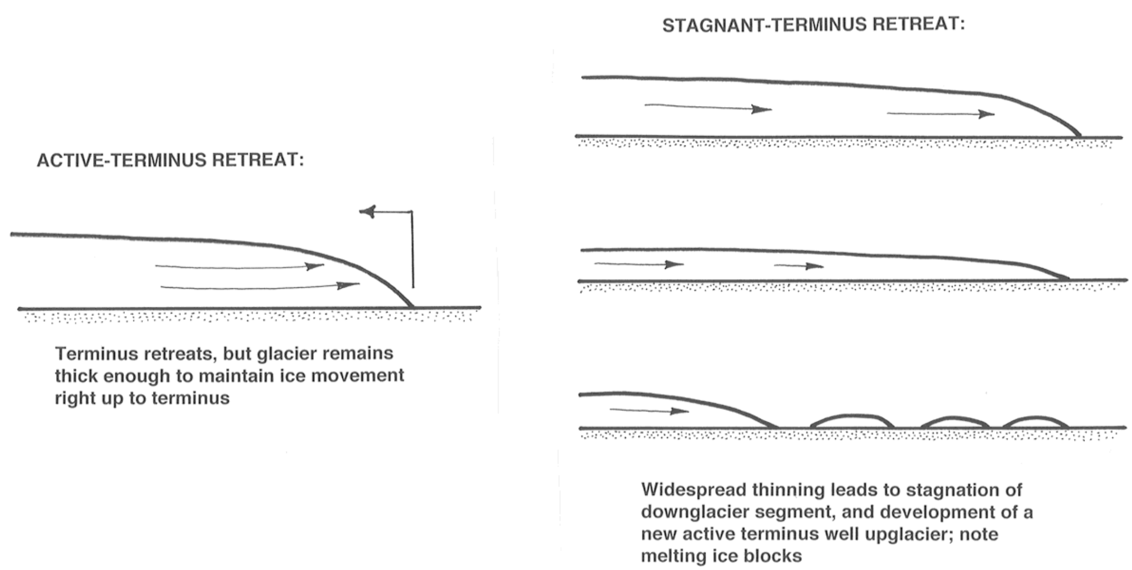


Figure 7-51. Active-terminus retreat vs. stagnant-terminus retreat.

12.5.3.2 An important point: most retreating glaciers in modern times show active-ice retreat rather than stagnant-ice retreat, either because they are valley glaciers or outlet glaciers, which usually remain well integrated even during retreat, or, if they are ice sheets with broad terminus fronts, they calve into the ocean. But stagnant-ice margins seem to have been very common at times of rapid ice-sheet retreat at the end of the Pleistocene, because there were widespread ice sheets on flat, low land in temperate regions. So we suffer somewhat from a lack of highly realistic modern models for deposition of the stratified drift we see resting on the landscape today.

12.5.3.3 Think first about deposition of sediment in meltwater streams before those streams leave the glacier.

- *supraglacial streams* could be carrying sediment that's deposited as a bed and then abandoned when the stream disappears; the sediment would then eventually let down onto the land surface by melting. Apparently this is not common.
- *englacial streams*: same thing.
- *subglacial streams*: these are by far the most important in carrying sediment, both because streams tend to migrate to the base of the glacier and because that's where most of the sediment is carried by the glacier.

12.5.3.4 Subglacial streams certainly deposit a lot of sediment as they leave the glacier. But how about deposition upstream of the glacier terminus? There may or may not be deposition there, depending upon the particular relationship between sediment load and carrying capacity. Figure 7-52 shows two possibilities. Sketch A shows a subglacial stream with uniform transport capacity; all the sediment is carried through over bare rock, and no sedimentary record of the tunnel remains after melting of the glacier. Sketch B, however, shows a situation in which sediment transport rate decreases downstream for some reason, to leave a deposit on the floor of the tunnel. In this case, upmelting of the tunnel roof is likely, to make room for the deposit as well as the flow. A sediment ridge that follows the course of the stream is left after the glacier melts.

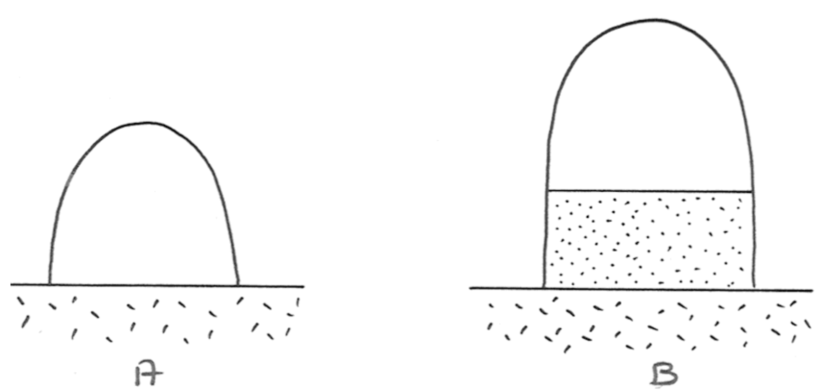


Figure 7-52. Subglacial stream tunnels A) without and B) with a sediment bed.

12.5.3.5 Remember that flow in subglacial tunnels is mechanically like pipe flow, in that there's likely to be no free surface, except very near the glacier terminus or at times of unusually low meltwater discharge. The floor of the tunnel is either bedrock or a sediment bed, and the walls and roof are of glacier ice.

12.5.3.6 What does the cross section of the subglacial flow tunnel look like? A natural first guess would be semicircular, with a horizontal planar floor. But if there's a full bed of sediment, it's possible for the floor to be either concave upward or convex upward instead of planar (Figure 7-53). The meager evidence from the exhumed record of ice-tunnel deposits suggests that the sediment bed is planar or convex upward, because the stratification overall tends to be as shown in Figure 7-54.

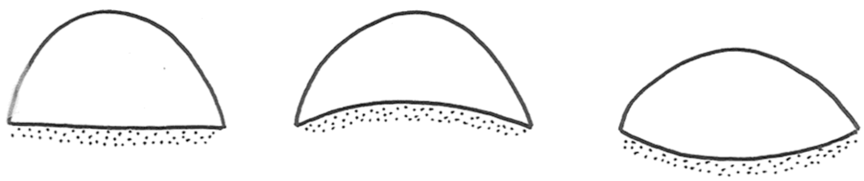


Figure 7-53. Possible cross-sectional shapes of subglacial ice tunnels.

Figure 7-54. Planar and convex-up deposits of subglacial ice tunnels.

12.5.3.7 There's a problem with this interpretation, however, because we never have direct independent evidence that the flow was really in a closed conduit. What about the possibility that the stream was flowing open to the sky in a canyon between two ice walls? That would always lead to a transversely horizontal bed rather than a convex-up bed.

12.5.3.8 One can account hydrodynamically for a transversely arched bed by appealing to *secondary circulations* in the flow (Figure 7-55). Such secondary circulations are known to be present in closed conduits of noncircular cross section. There are various theories on the origin of such secondary circulations, but they are not yet well understood. The circulation pattern shown in Figure 7-55 would lead to a component of sediment movement on the bed from the walls to the center, until an equilibrium transverse profile is reached in which there's a balance between this tendency toward upslope transport and the pull of gravity back down the slope toward the corners of the cross section.

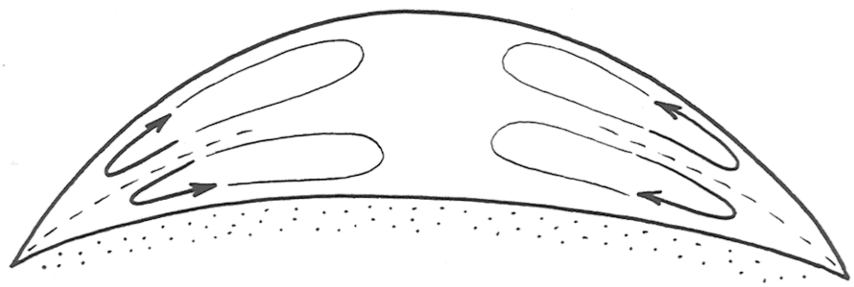


Figure 7-55. Secondary circulations sin flow in subglacial ice tunnels.

12.5.3.9 Deposits of the kind formed in ice tunnels are called ***eskers***, a nongenetic term for *any long and more or less isolated ridge of stratified drift, oriented at a small angle to the overall direction of ice flow, and presumably deposited by meltwater beneath or at the terminus of the glacier*. Most eskers seem to have been deposited by meltwater streams either within the glacier, as discussed above, or as they emerge from beneath the glacier.

12.5.3.10 Eskers have rather variable features, probably reflecting a polygenetic origin (Figure 7-56). Here are some common characteristics of eskers:

- kilometers to well over 100 km long
- up to a few tens of meters high, and up to a few hundreds of meters wide
- either flat-topped or arched
- almost always all sand and gravel, well stratified; no till
- the sediment is often slumped at the margins
- often sinuous, like streams
- sometimes show tributaries and branching
- sometimes pass over bedrock divides (by as much as 200 m!) at the lowest saddle point
- sometimes end in what look to be fans or deltas
- show evidence of high-velocity flow

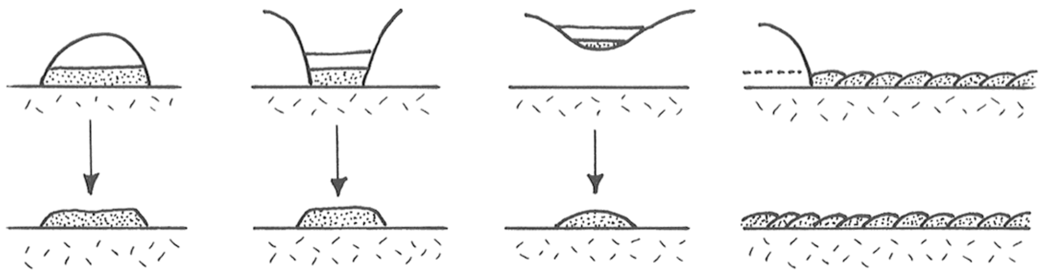


Figure 7-56. Possible mechanisms for development of eskers.

12.5.3.11 Many, maybe most, eskers seem to have been deposited by subglacial streams in closed flow; this accounts well for their passing over bedrock divides, for the arched bedding, and for the existence of tributaries. But of course an open-to-the-air esker formed on glacier ice could have been let down onto the bedrock divide by melting of the ice.

12.5.3.12 Many, but certainly not most, eskers must have been deposited at the downstream end of a meltwater stream where the stream reached the glacier terminus. The long length of the esker is then accounted for by year-by-year shifting of the site of deposition upglacier, as the terminus retreats. Some eskers are *beaded* (that is, they consist of a chain of closely spaced mounds of glacial deposits), and this mode of origin accounts well for the beading. Keep in mind that whatever the mode of origin the long length of eskers is accentuated by the retreat of the glacier.

12.5.3.13 Aside from eskers, many kinds of deposits are left by melting of ice at the terminus of a retreating glacier. These might generally be called ***ice-disintegration deposits***. These are suites of interrelated but distinctive depositional features produced by wasting of the thin, stagnant, marginal part of a glacier, mostly by separation of the ice into isolated masses. The marginal zone of the glacier in which such features are produced might be as much as ten kilometers wide. Instead of being carried to an active terminus, the drift accumulates by various processes associated with running water, in channels or in lakes, and in various positions relative to the melting ice: in channels or openings between or beneath ice blocks, and over the surface of wasting ice.

12.5.3.14 The *surface morphology* of these deposits varies widely. There's a mostly continuous spectrum or gradation of named features: *kames*, *kame terraces*, *kame deltas*, *collapsed masses*, and *eskers* (already discussed). All of these can be observed forming today near the termini of glaciers that are retreating in an inactive-terminus mode. One of the best examples is the Malaspina piedmont glacier in Alaska.

Kame terraces. In areas of active meltwater production in valley glaciers, meltwater streams often flow down the valley along the sides of the glacier. This often leaves linear deposits of stratified sand and gravel on the bedrock slopes of the valley after the glacier melts (Figure 7-57). Often these deposits are at more than one level, as a function of different stillstands of the terminus. It's easy to confuse kame terraces with stream terraces in a nonglaciated valley.

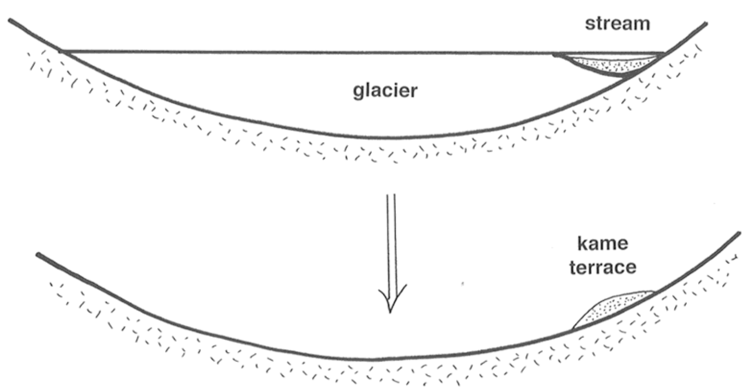


Figure 7-57. Development of a kame terrace.

Kames. The term ***kame*** is used for a whole variety of hills, knobs, and ridges (positive relief forms), more or less isolated from one another, of stratified sand and gravel *deposited against or adjacent to wasting stagnant ice* (Figure 7-58). The sediment may be derived in part from the adjacent ice itself, but the larger kames are probably formed by deposition of sediment carried from upstream by meltwater streams. Kame deposits are characteristically in part deformed, because of the collapse of sediments when the supporting ice melts away.

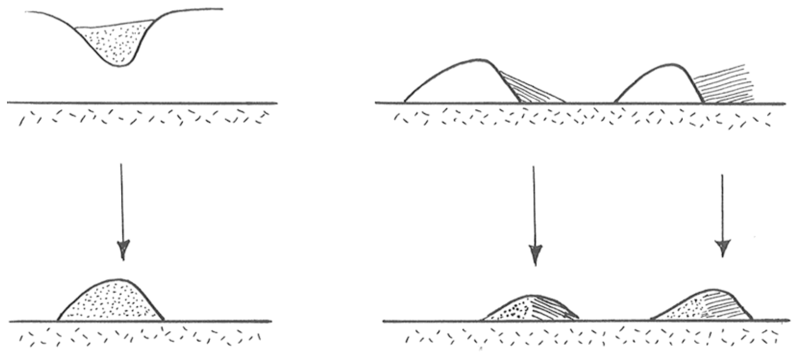


Figure 7-58. Development of kames.

Kame Deltas. Small glacial lakes are common right next to melting glacier ice, because of local damming of meltwater drainage by either the ice itself or sediment deposited by the ice. Streams flowing between, or from the surfaces of, the melting ice masses tend to deposit deltas in the lakes. After the ice melts, the delta forms an isolated and often flat-topped mass of sediment (Figure 7-59).

Collapsed Masses. Another common ice-disintegration feature, for which there is no very felicitous term, consists of irregular masses of sediment deposited on top of wasting ice and then let down onto the land surface as the ice melts. The so-called collapsed masses tend to be blanketlike, but they are highly varied, with sharp variations in thickness. This is sort of a "wastebasket" category.

Kettles. ***Kettles*** are *depressions formed when a thick layer of drift is deposited around or over an isolated ice mass and then the ice later melts* (Figure 7-60). If the groundwater table later lies above the floor of the depression, the kettle has a lake within it, called a ***kettle lake***. Kettles often are present in large numbers in areas of outwash plains; such an outwash plain is called a ***pitted outwash plain***.

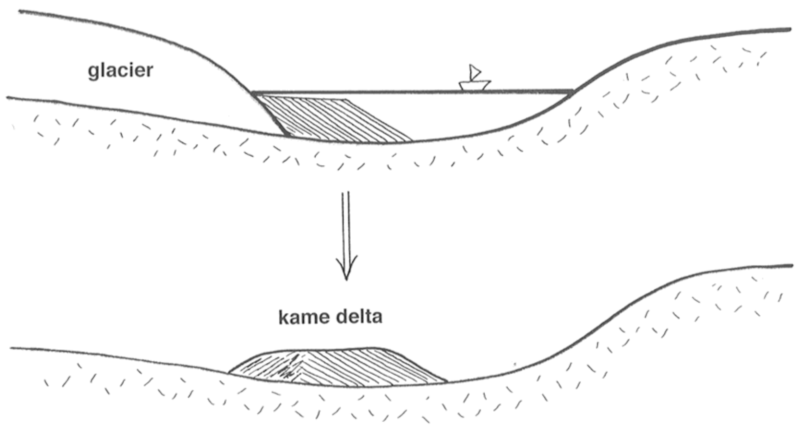


Figure 7-59. Development of a kame delta.

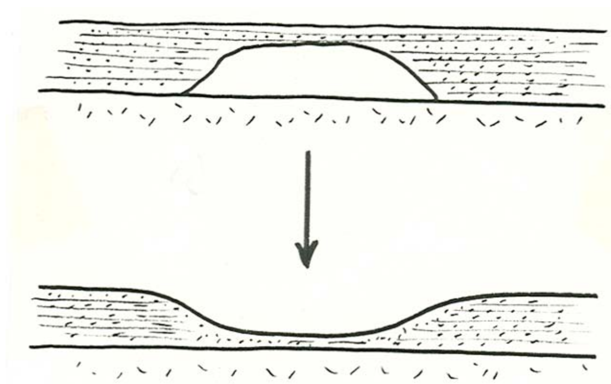


Figure 7-60. Development of a kettle.

12.5.4 Proglacial Drift

12.5.4.1 ***Proglacial drift*** is drift deposited in various ways, by water or even by wind, more or less far removed from the glacier. The idea that proglacial drift is abundant makes sense. Look at a typical active warm glacier. There's abundant meltwater, so large meltwater streams originating well upglacier from the terminus are important during the melting season. These streams characteristically carry high sediment loads. In such situations, most of the drift deposited by the glacier is deposited downstream from the glacier terminus. This happens whether the glacier is advancing, retreating, or stationary; if the glacier is advancing, then the glacier later overrides some of its own proglacial drift. Here are the important kinds of proglacial drift:

- outwash (glaciofluvial)
- lake deposits (glaciolacustrine)
- marine deposits (glaciomarine)
- loess (wind-blown silt)
- sand dunes (usually reworked outwash)

Glaciofluvial deposits (these are also called glacial ***outwash***)

- Mainly *sand and gravel*; fine material (silt and clay) is either carried farther downstream into areas not recognizably glacial, or blown away by the wind
- Essentially *fluvial* in nature; channel pattern and sedimentary structures are not directly or recognizably glacial
- Typically shows *rapid downstream decrease in grain siz*e, because of overall aggradation
- Outwash streams are typically *braided* (factors: high sediment load; no bank stability)

The typical form taken by outwash is a fan or cone of fluvial sediment, deposited with the apex or head at the ice margin. The reasons for fan development are basically the same as for alluvial fans in nonglacial situations: abrupt decrease in gradient as the stream leaves the glacier and abrupt relaxation of channel constraints leads to decrease in capacity and therefore deposition. Outwash deposits tend toward *two distinctive forms*:

- **Outwash plains:** broad linear front, many outwash streams, series of coalescing fans. This leads to deposition of a broad wedge or blanket, thinning and fining away from the glacier (Figure 7-61). Also called a ***sandur***, from the Icelandic (plural: ***sandar***). One of the best sandar is at the margin of Vatnayökull, the largest ice cap in Iceland.
- **Valley trains:** braided outwash filling a glaciated valley downstream of an active valley glacier (Figure 7-62).

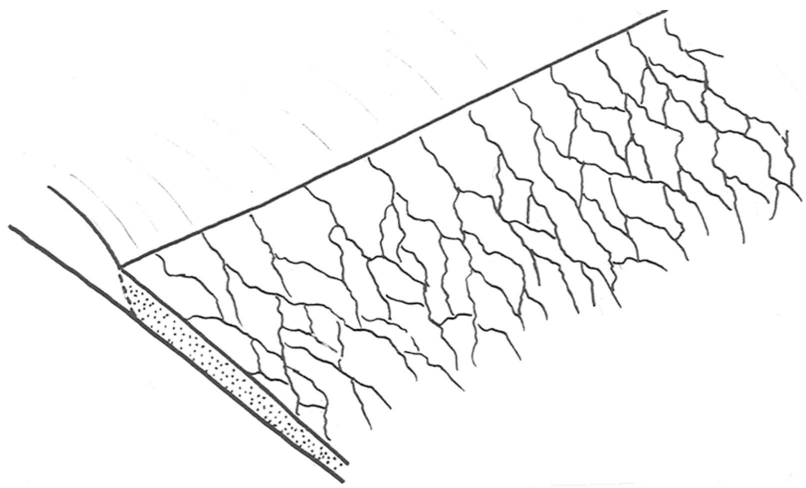


Figure 7-61. A glacial outwash plain (a sandur)

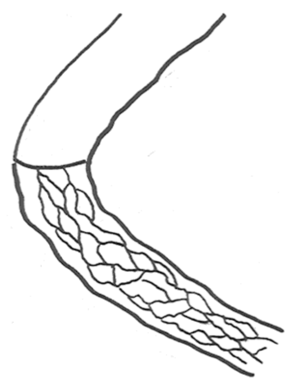


Figure 7-62. A valley train.

Commonly where both meltwater and drift are very abundant (as in wasting of a warm ice sheet in the Pleistocene), the outwash plain or fan is built up to be so thick that it is built headward over the terminus of the glacier itself and merges with ablation moraine. Often it buries stagnant ice near the terminus, leading to a pitted outwash plain as the buried ice melts. Earlier end moraines are often buried as well. This is well displayed on Cape Cod and Long Island.

Glaciolacustrine Deposits

Glaciers create three kinds of lakes:

- glacier-margin lakes dammed by the ice itself (short-lived)
- glacier-margin lakes dammed downstream by earlier-deposited glacial sediment (longer-lived but still temporary)
- bedrock depressions made by glacial erosion, later filled with water (long-lived)

Sediments deposited in glacier-margin lakes are very common, because valleys are dammed by ice or by ice-disintegration deposits.

- Probably *more common during retreat that advance*
- Invariably *temporary*, because ice dams melt soon, and glacial-sediment spillways are degraded by erosion
- Range in *size* from very small and lasting only years, to enormous, lasting thousands of years.

How does one explain the relatively long lives (some thousands of years) of some glacier-margin lakes? If the retreating glacier forms a long-term but moving dam in a valley, and drainage is over a rock divide, the lake remains either until the ice melts or until a sediment-formed divide is cut to the level of the bedrock spillway. Example: Glacial Lake Hitchcock in central Connecticut and Massachusetts.

Some points about sedimentation in glacial lakes (Figure 7-63):

- *Deltas* of coarse sediment are common
- Varves (*annual coarse–fine couplets deposited on the lake bottom*) are formed by differences in summer and winter suspended-sediment deposition (Figure 7-64).
- Dropstones are possible, from floating ice derived from the glacier.

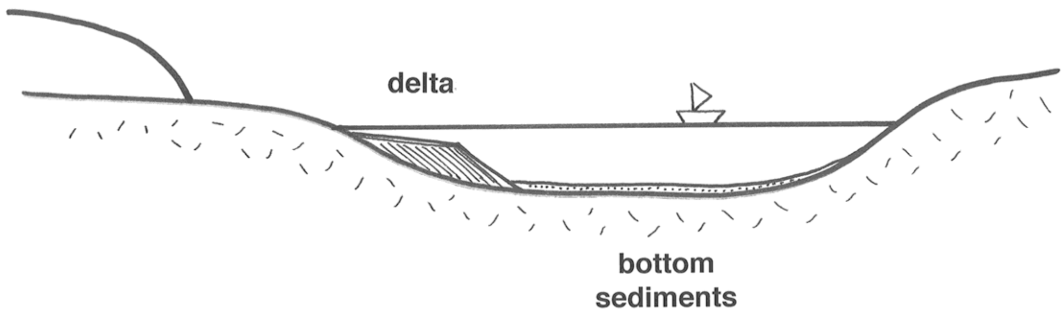


Figure 7-63. Sedimentation in glacial lakes.

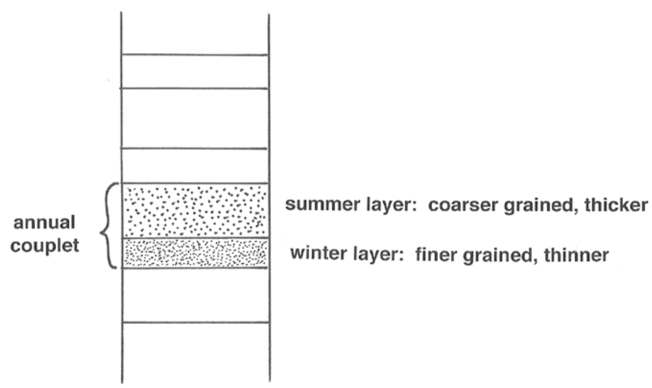


Figure 7-64. Glacial varves.

Loess

The term ***loess*** is used for *blankets of wind-deposited silt on the land surface*. Much but by no means all loess is derived from wind erosion of broad outwash plains marginal to Pleistocene ice sheets. Some is derived also from extensive wind erosion in large deserts, not associated with glacial deposits.

Loess is unconsolidated to semiconsolidated (by slight to moderate simple cementation), and usually buff to yellow to tan in color, reflecting an oxidized state. It is unstratified to only vaguely stratified. It has the interesting property of standing in vertical slopes even though it's easily dug with a shovel (interlocking angularity of grains, plus slight cementation?), and sometimes it even shows columnar jointing. It consists of relatively well sorted and angular grains usually in the fine silt to coarse silt range (average grain size 0.01–0.05 mm). Quartz is usually the dominant mineral. The lack of stratification is probably due to bioturbation by plants and animals, together with the relative uniformity of supply.

Loess forms blankets from less than a meter to many tens of meters thick (over 200 m in the central parts of China). Thickness is well correlated with grain size. In North America, loess is widespread in east-central to west-central US (Kansas, Nebraska, Iowa, Missouri, Wisconsin, Illinois, Indiana, Kentucky, Ohio) and also the Pacific Northwest. Coverage in North America: 1.6×10^6 km^2.

Glaciomarine Deposits

Icebergs produced by calving of large glaciers into the ocean often contain abundant drift, if the glacier is an active warm-based glacier. As the icebergs drift in the ocean and melt, they release this load, which settles to the sea floor along with fine sediment derived from elsewhere.

Glaciomarine deposits are characteristically well stratified but poorly sorted at the same time. The good stratification presumably comes about by annual and longer-term fluctuations in sediment supply from the icebergs. A distinctive feature of glaciomarine deposits is the presence of ***dropstones***: unusually large iceberg-derived clasts which bow down the sediment upon impact and which are then buried by later strata that arch over the dropstone.

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Patterson, W.S.B., 1994, The Physics of Glaciers, Third Edition. Pergamon, 480 p. (the source on the physics of glaciers most accessible to the nonspecialist)

Strahler, A.N., 1966, A Geologist's View of Cape Cod. The Natural History Press, 115 p. (a small, readable book, old but still good, that deals in part in a very engaging way with the glacial geology of Cape Cod; long out of print)

Sugden, D.E., and John, B.S., 1976, Glaciers and Landscape. Arnold, 376 p. (a good but aging source, not highly technical)