

Chapter 17

Glacial Geomorphology

Glacial geomorphology concerns the processes by which glaciers shape landscapes and analyzes the resulting glaciated landforms. The information in Chapter 16 about the deformation of ice and the flow of glaciers at various temperatures, with or without associated meltwater, is a necessary prerequisite to understand this chapter, but here the emphasis is on glaciers and meltwater as geomorphic agents that erode the rock mass over which they flow and transport or deposit the resulting debris.

Although the glacier-related geomorphic work of meltwater and wind extend many kilometers beyond a glacier margin, the direct work of glacier ice ends at an abrupt, although shifting, boundary. This is one of the few examples that can be cited where geomorphic processes and landforms change at a sharp boundary. Only at coasts is there such an equally abrupt process-controlled change of landforms.

A problem of glacial geomorphology is that most of the processes work beneath ice and are exceptionally difficult to observe. Glaciated terrain is not a landscape until it is exposed by the disappearance of the glacier. In that sense, all glaciated landscapes are relict even if they were exposed only by the previous year's retreat of a glacier terminus.

Valley glaciers create a unique, distinctive landscape by their vigorous erosional shaping of mountain ranges. As noted later in the chapter, the assemblage of sharp pyramidal peaks, knife-edged ridges, and U-shaped troughs with hanging tributaries is so much a

part of our mental image of glacierized and glaciated mountains that we are hard pressed to deduce what the landforms were on a nonglaciated high mountain range during the "greenhouse" climate of the middle Cenozoic Era (Chapter 4). Even though valley glaciers only occupy and modify preexisting river valleys, those modifications can be extreme.

The Pleistocene continental ice sheets pose another set of challenges to geomorphologists. Their erosional work may have been primarily to strip away a former regolith and modify former drainage systems, and their deposits may be measured only in tens of meters of thickness, but they created the landscapes and soils of northern Europe and North America, where the industrial revolution began and agriculture has become a business instead of a means of individual survival. Soils on the vast glaciated and loess plains are young but fertile, formed in transported sediments rather than in an *in situ* regolith. A large percentage of the readers of this book live on a landscape of relict glacial landforms, little modified by 10,000 to 20,000 years of postglacial time.

GLACIAL EROSION AND SEDIMENT TRANSPORT

Valley glaciers probably derive most of their sediment loads from intense weathering and mass wasting on adjacent slopes. Rock debris that composes lateral and medial moraines seems to stay fairly high in the glacier

cross section. An unknown amount is actually eroded by the moving glacier.

Ice sheets that bury landscapes cannot gain sediment by mass-wasting because nothing is exposed above the ice. The Antarctic and Greenland ice sheets are notably free of any sediment other than minute amounts of windblown dust, except very near their bases. Neither ice sheet appears to be transporting much rock debris, but they may not be good models for the former ice sheets of Europe and North America, which expanded and shrank repeatedly during the Pleistocene Epoch.

Fluvial sediment transport beneath wet-base glaciers has largely replaced other mechanisms as the process most favored to deliver sediment to the margins of glaciers (Evenson and Clinch, 1985; Gustavson and Boothroyd, 1987). More and more, glacial and fluvial processes are being studied in conjunction with each other as a "glaciofluvial system" analogous to the fluvial system introduced in Chapter 10 (Ashley et al., 1985; Drewry, 1986; Gurnell and Clark, 1987; Lawson, 1993). Even more complex are the glaciolacustrine and glaciomarine environments, where a calving ice margin stands in fresh or salt water. Extensive segments of the former ice sheets were grounded in water, and the resulting sediments are the subject of many sedimentological studies (Dowdeswell and Scourse, 1990; Anderson and Ashley, 1991).

Processes and Bedforms

Rock masses that are eroded by overriding glacier ice produce detritus with a wide range of grain sizes that must then be transported away. Left behind are rock surfaces with diagnostic minor landforms and surface markings, collectively called **bedforms**. Unlike water and wind, glacial transport has no competence limit (p. 215), and all grain sizes from clay to giant boulders are transported, often thoroughly mixed. Some of the debris is entrained into the basal ice. A larger fraction is probably transported by water from beneath wet-base glaciers. Erosion, entrainment, and transport form a continuum that ends only when the particles reach a final resting place.

Glaciotectonism. Both the vertical (static) load and the shear stress of the forward motion of glacier ice contribute to deforming or moving underlying rock masses or sediments. Pore water pressure and the resulting *effective stress* (p. 170) applied to the substrate by the weight of overlying ice are critical factors in determining the response of glacially overridden materials (van der Wateren, 1995). Shallow slabs of frozen sediment overlying a deeper horizon saturated with confined water, for example, might be easily thrust forward even by very low basal shear stress beneath over-

riding ice. Ice-push ridges of shale and mudstone, some stacked as imbricate thrust sheets, are well known on the Canadian prairies (Aber et al., 1989, p. 37). Similar forms are widely distributed in Denmark, Iceland, the Netherlands, and elsewhere. Their arcuate map patterns and deformation vectors that are clearly related to former ice motion, and their shallow detachment zones, are criteria that demonstrate a glacial origin.

Crushing and Fracture. The crushing strength of most rocks greatly exceeds any combination of normal and shear stresses that can be applied to the bed by overriding glacier ice. However, at a point contact between a rock particle embedded in the ice and the rock substrate, pressure is adequate to chip or flake both the particle and the substrate, just as mild finger pressure on the broad head of a thumbtack can drive the point into hard wood. A wide range of distinctive fractures can be seen on glaciated rock surfaces, including *chattermarks* and *crescentic fractures* (Drewry, 1986, p. 34). These are made as hard particles rolling at the ice-rock interface elastically compress the rock surface in front of them and stretch the rock behind them to create rhythmic fractures. Similar forms are reproduced by rolling a hard steel ball over glass under a heavy load. Large rock obstructions that protrude upward into ice can also be broken by the excess ice pressure on their up-glacier sides, even though temperate ice will attempt to flow around such an obstruction by enhanced plastic flow and regelation slip (p. 366). Because near-surface rock masses are usually weathered and full of microfractures and pores, fractures occur at stresses lower than the predicted strength of the rocks.

Plucking and Hydraulic Jacking. During regelation slip, water migrates to the downglacier side of a subice obstruction and refreezes (p. 366). Rock fragments, usually small, can be incorporated into the regelation layer and carried away by the glacier. This process is called glacial **plucking**. It is not clear whether regelation ice has enough tensile strength to actually pull away lee-side pieces of rock or whether pieces already loosened by the stress of the moving ice are simply frozen in and carried away (Boulton, 1974, pp. 69–75).

Plucking is most effective on fractured rocks. Jahns (1943) reconstructed the sheeting joints on crystalline metamorphosed rocks near Boston, Massachusetts (p. 122) to show that plucking had removed several times the volume of rock from the downglacier, or **lee** sides of hills than abrasion had removed from the upglacier, or **stoss** (German for *push*) sides.

A related process involving rapid increases of water pressure in subice cavities causes basal ice to move suddenly, pushing joint-bounded blocks of rock forward.

Blocks up to 1 m in length have been observed. The process is called **hydraulic jacking** (Drewry, 1986, pp. 46–47).

Abrasion. The abraded scratches and grooves that characterize glaciated rock surfaces are undoubtedly caused by rock fragments in the ice rather than by the ice itself (Boulton, 1974, p. 43). If the glacier is sliding over its base on a water film, any fragment larger than the thickness of the water film will be rolled or dragged against the rock surface. Glaciated surfaces show a range of abraded microrelief forms from hairline *striations* to *grooves* a meter or more in depth and width. Careful study of glacial striations and grooves can sometimes reveal the direction of ice motion, as when a piece of rock is chipped off and dragged downglacier over the surface, or when a resistant projection protects less resistant rock in its lee from abrasion (Figure 17-1). Regelation slip must be an effective process of abrasion, because if a sand grain is embedded in basal ice, it will move into contact with the rock surface as the enclosing ice melts. Abrasion is ineffective beneath polar glaciers, where there is no sliding at the ice-rock interface (p. 364). The strongest proof that a region has been glaciated by a wet-base glacier is a regional pattern of striated and grooved bedrock surfaces.

Not only bedrock surfaces are abraded; the rock fragments carried in glaciers also are subjected to abrasion. A significant proportion of glacially transported pebbles and cobbles have striated or polished planocurved facets on one or more sides, similar in shape to the sole of an old shoe, showing that they were held in contact with other rocks as they ground past each other.

Abrasional effectiveness depends on the relative hardnesses of the "tool" held by the ice and the rock substrate, the flux of abrasive particles in contact with

the bed, and the effective force with which the particles are held against the bed (Boulton, 1974; Hallet, 1979). At the base of a temperate glacier, the normal stress (ice thickness) is a factor in determining basal shear stress (Figure 16-13), which in turn affects the basal ice velocity. The flux of abrasive particles in contact with the bed is reduced by a debris concentration above a certain relatively low value. Glaciers with significant basal sliding will abrade their beds until the basal load becomes large and the sliding rate decreases, when deposition will begin (Hallet, 1981).

Some polished and striated rock surfaces show patterns of flow that diverge from the regional direction in minute, complex ways (Figure 17-2). Local flow directions show eddylike patterns, helical flow along shallow troughs, and potholes that suggest a highly mobile abrasive substance (Gjessing, 1966). Possibly, the base of some temperate and intermediate glaciers is underlain by water-saturated, fine-grained sediment under high water pressure and therefore low effective stress, that deforms under the shearing stress applied by the overriding ice (Alley, 1989; Alley et al., 1989). The *deformable bed* (p. 368) moves generally in the same direction as the overlying ice but sensitively responds to the locally intensified stress gradients around obstructions. Whether it is capable of abrading the rock surface beneath it, or whether it absorbs all the shearing motion of the glacier within itself, is not known. A deformable bed of saturated, shearing sediment could deform, erode, or accumulate, depending on the stress field within it and within the glacier ice immediately above it (Boulton, 1996). Such saturated basal sediment has been exposed by rapid retreat of some glacier termini and has been observed in tunnels dug along the base of valley glaciers. It also makes constructional landforms when it squeezes up into crevasses near the margins of glaciers.

FIGURE 17-1. Glacially abraded surface of limestone, with a protruding chert concretion, a streamlined tail of limestone, and a shallow marginal moat (photo: A. Meike).

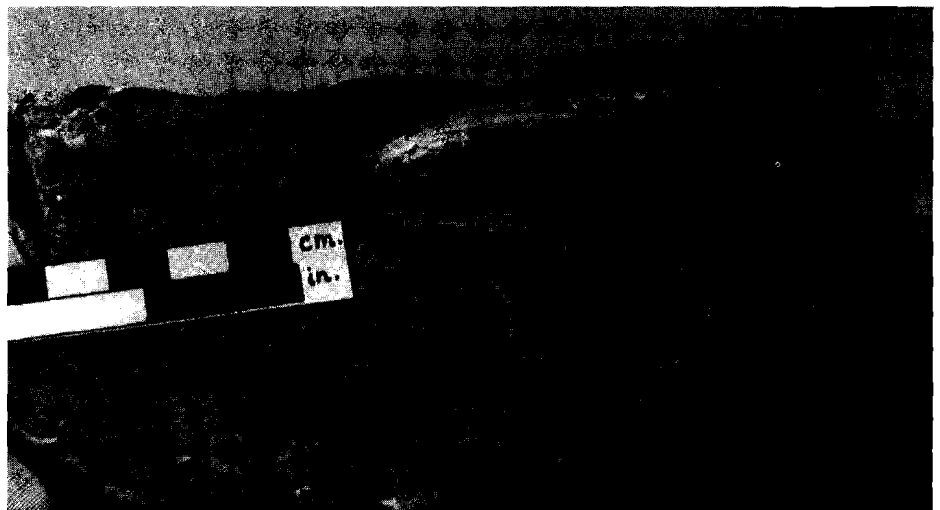


FIGURE 17-2. Glacially scoured metamorphic rock suggesting a highly plastic abrasive medium, near Portør, Norway, on the Skagerrak coast (photo: Just Gjessing).



The combination of crushing, plucking, and abrasion exploits structural details of rock masses, especially fractures. A variety of streamlined bedforms, such as *stoss-and-lee topography*, *roches moutonnées*, and *crag-and-tail* hills are variants of the combined abrasion and plucking process (Figure 17-3). They range in length from a few meters to about 1 km. They may form equally well under ice sheets or valley glaciers, but a wet-base status is necessary. All are small hills with abraded upglacier sides, but the *crag-and-tail* variant has a streamlined tail of compact sediment or erodible bedrock sheltered behind a more resistant crag (Figure 17-1), whereas the other two forms have steepened, plucked lee faces. Many of these bedforms may be only slightly modified by glaciation and are primarily inherited from the preglacial relief (Lindström, 1988).

Fluvial Erosion Beneath Glaciers. With the recognition of the widespread occurrence of wet-base glaciers, both at the present time and during former ice ages, the erosional role of flowing water beneath glaciers has received new emphasis. The basal film of water observed beneath temperate and intermediate glaciers is not capable of mechanical erosion, although by virtue of its high confined pressure, high dissolved CO₂ content, and low temperature it can chemically corrode carbonate rocks (Hallet, 1976; Drewry, 1986, pp. 77–82). However, when it is flowing in subglacial channels, the relatively warm and often muddy meltwater can erode both the base of the ice and the bedrock or sediment beneath the ice. *Potholes* in bedrock eroded by vortices in subglacial rivers are well-known glacial erosional landforms. Very large

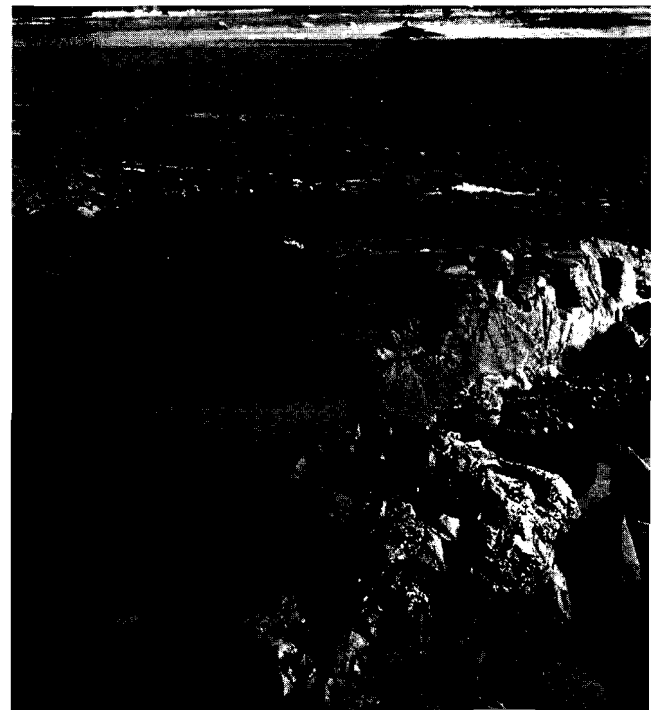


FIGURE 17-3. Intertidal rock knob on Kennebunk Beach, Maine, showing stoss-and-lee topography. Ice moved from left to right, abrading the upglacier (stoss) side and plucking the downglacier (lee) side along joints.

tunnel valleys were eroded into unconsolidated sediments by subglacial meltwater along some margins of the late Pleistocene ice sheets. Those in central Minnesota are up to 1 km in width and 30 km in length (Mooers, 1989). Whether the tunnels entrench into the substrate or melt upward into the ice, and whether they close by ice deformation or become filled with sediment is a complex function of the amount of meltwater, the basal water pressure, ice plasticity, sediment load, and many other variables.

Erosion of subglacial sediment and its transportation to the ice margin by subglacial and englacial streams is claimed to be the source of almost all the outwash in front of the Malaspina Glacier in Alaska (Gustavson and Boothroyd, 1987). An extreme view is that episodic subglacial floods from beneath the southern part of the North American Laurentide ice sheet were responsible for most of the glacial erosional landforms on both bedrock and sediment, including polished and abraded streamlined forms (Sharpe and Shaw, 1989; Shaw, 1989). A continuing debate concerns the relative importance of subglacial water and highly deformed basal sediment in producing the polished, striated, and fluted surfaces on hard rocks so diagnostic of glacial erosion (Figures 17-1 and 17-2).

Erosional and Residual Landforms

Troughs. The erosional landforms of valley glaciers rank with the most spectacular scenery on earth. The European Alps are so thoroughly dissected by valley glaciers that the entire assemblage is sometimes simply referred to as *alpine topography* (Figures 17-4, 17-5, and 17-6). The fundamental landform component is the **glacial trough**, the channel of a former valley glacier. The cross profile of a glacial trough is a catenary curve, not unlike the cross profile of a river channel but quite different from the profile of an entire river valley. Numerical modeling of glacial troughs, based on physical concepts of least work and optimal erosional effectiveness, confirm that their generally parabolic cross profiles are highly efficient for ice flow (Hirano and Aniya, 1988, 1989; Harbor, 1992). With reasonable assumptions of erosional efficiency, a glacier could convert a V-shaped valley into a glacial trough in as little as 10^4 years (Harbor, 1992). Obviously, rock-mass strength (Chapter 7) is a very important factor in the rate and degree of trough formation (Augustinus, 1992, 1995). Computer simulations of trough erosion along a narrow belt of highly erodible rock predict a narrower, deeper trough, as expected (Harbor, 1995).

Troughs are on the order of 1 km or more in width, and commonly hundreds of meters or more in depth. Concave slopes dominate, rising from the valley floor to steep cliffs or stepped-back, ice-abraded shoulders

(Figures 16-3, 17-6, and 18-9). In long profile, glacial troughs may be a series of basins separated by cross-valley ledges of more resistant rocks (Figure 17-5) (Gutenberg et al., 1956). The long profile is structurally controlled by intense glacial plucking and abrasion on more highly fractured or easily eroded rock masses to create the basins. Alternating segments of compressing and extending flow may also differentially erode the glacier bed and enhance initial structurally controlled irregularities, or possibly create harmonic changes in glacier dynamics that cause regularly spaced basins without structural control. Typically, deep-basin erosion is localized by valley-wall constrictions, and sills are associated with widening of the valley walls (Shoemaker, 1986). If not filled with sediment, the basins form chains of lakes, each with a waterfall over the downstream ledge to the next lake. These delightful chains are called **paternoster lakes**, from the resemblance to shining beads on a rosary. It is helpful to visualize a glacial trough as the analog to a river channel. Just as a river channel thalweg may have alternating pools and riffles (p. 220), the basal ice can flow up and down as long as the surface gradient provides the necessary downvalley gravitational potential.

Trough Lakes and Fiords. In a glacial trough that has either a rock sill or a massive end moraine blocking its lower end, a **trough lake** will drown the valley floor. Trough lakes are usually long, steep-sided, deep, and cold, with a minimal inflow of sediment because of the clean abraded valley walls and the upstream settling basins in paternoster lakes on most tributary streams. The Swiss lakes, the Como region of northern Italy, the Lake District of England, the Finger Lakes of central New York (Figure 17-7), and many other regions of large trough lakes have become justly famous as scenic tourist areas. Hydroelectric projects make use of the large storage volumes and abrupt gradients associated with trough lakes in mountainous regions.

The side walls of glacial troughs are straightened by abrasion and plucking. The interlocking spurs that interrupt views along a river valley are cut away, giving a series of **ice-faceted spurs** that align between lateral valleys. Where tributary glaciers joined the trunk ice stream, smaller glacial troughs form, but because the smaller glaciers were less erosive than the trunk glacier, their floors are usually well above the floor of the main valley. Waterfalls cascade down the main valley wall from these **hanging troughs**. The accordance of valley floors, so important to Playfair's law (p. 232), is absent in alpine glacial scenery. However, if we think of glacial troughs as analogous with river channels rather than valleys the comparison is better because tributary river beds are commonly shallower than their trunk channel.

GLACIERS IN AN ALPINE REGION

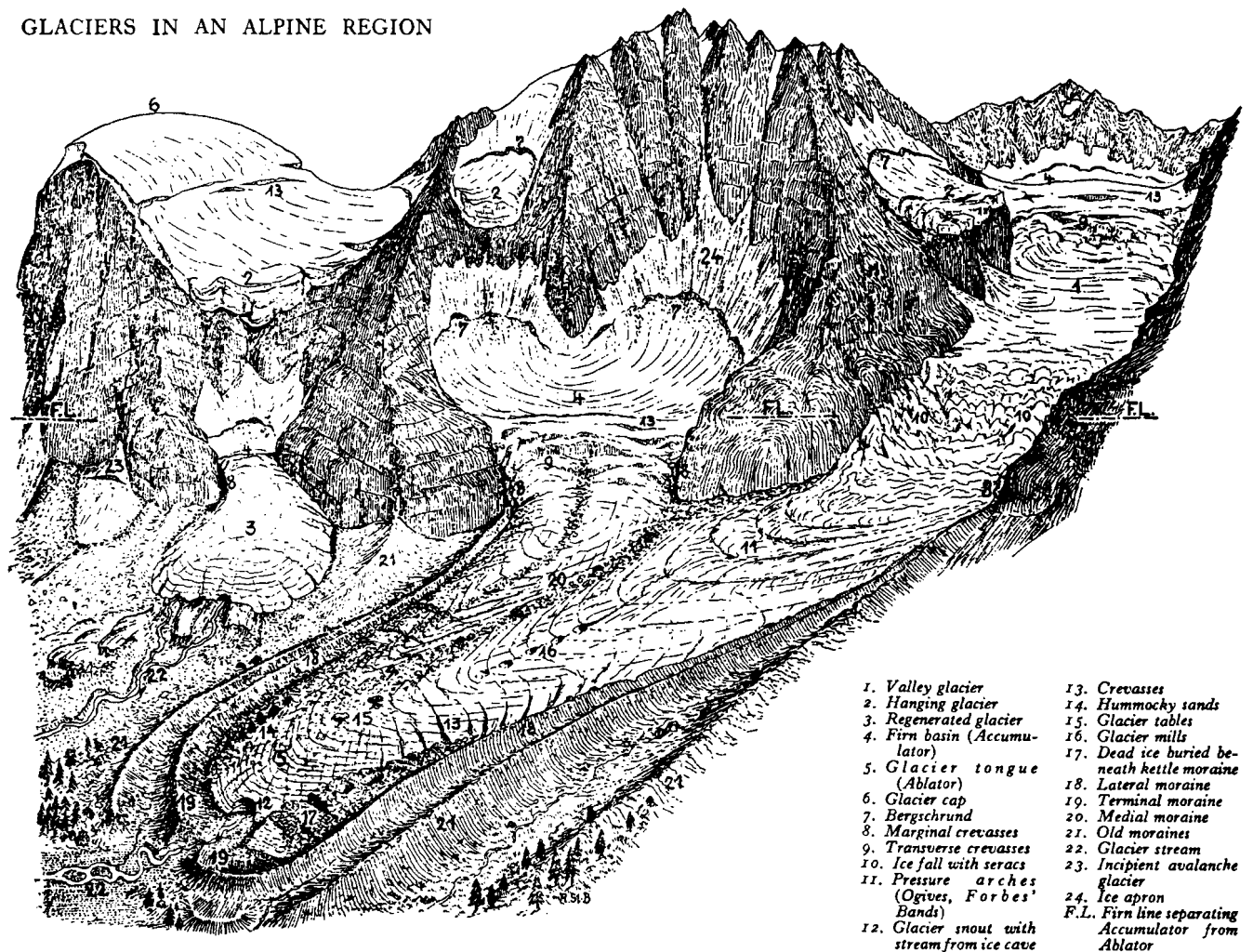


FIGURE 17-4. Glacier landforms in an Alpine region (Streiff-Becker, 1947).

Fjords are glacial troughs that have been drowned by the sea (Figure 20-5). Their form differs from that of other troughs in that near the outer coast, many fjords shoal rapidly to only a fraction of the depth that they attain between more restricting walls farther inland, where multiple basins may be 1 km or more in depth. In some fjords, the sill at the mouth is an end moraine, but usually it is bedrock and is attributed to a reduction in intensity of vertical erosion as the valley glacier either began to float (Crary, 1966) or spread laterally as a piedmont lobe. The analysis of extending and compressing flow in glaciers (p. 357) assumes that longitudinal flow is confined between parallel walls so that no lateral expansion or contraction occurs. Where their valley walls widen, as on a piedmont plain, glaciers spread laterally and thin abruptly. Basal erosion decreases dramatically and the trough floor rapidly rises to a terminal sill (Shoemaker, 1986).

Cirques. At the head of each trough, in the accumulation area of the glacier that eroded the valley, are steep-headed, semicircular basins, or **cirques** (Figure 7-4; Color Plate 13). The bowl shape is a consequence of intense frost action and nivation processes adjacent to the ice field plus the flow lines in the zone of accumulation (Figure 16-2) that carry surface rock debris downward toward the base of the glacier, where the bed is heavily abraded. A cirque has much the shape of a human heel print in sand and is often "down-at-the-heel," that is, more deeply eroded as a basin behind an abraded rock sill at the lip of the cirque. A cirque lake, or **tarn**, dammed by the rock sill, occupies the basin in many cirques.

Cirques are a unique and diagnostic feature of alpine glaciation. The association of active cirques with the accumulation areas of modern valley glaciers led to the recognition of similar relict forms very early in the

THE SAME REGION AFTER DISAPPEARANCE OF THE GLACIERS

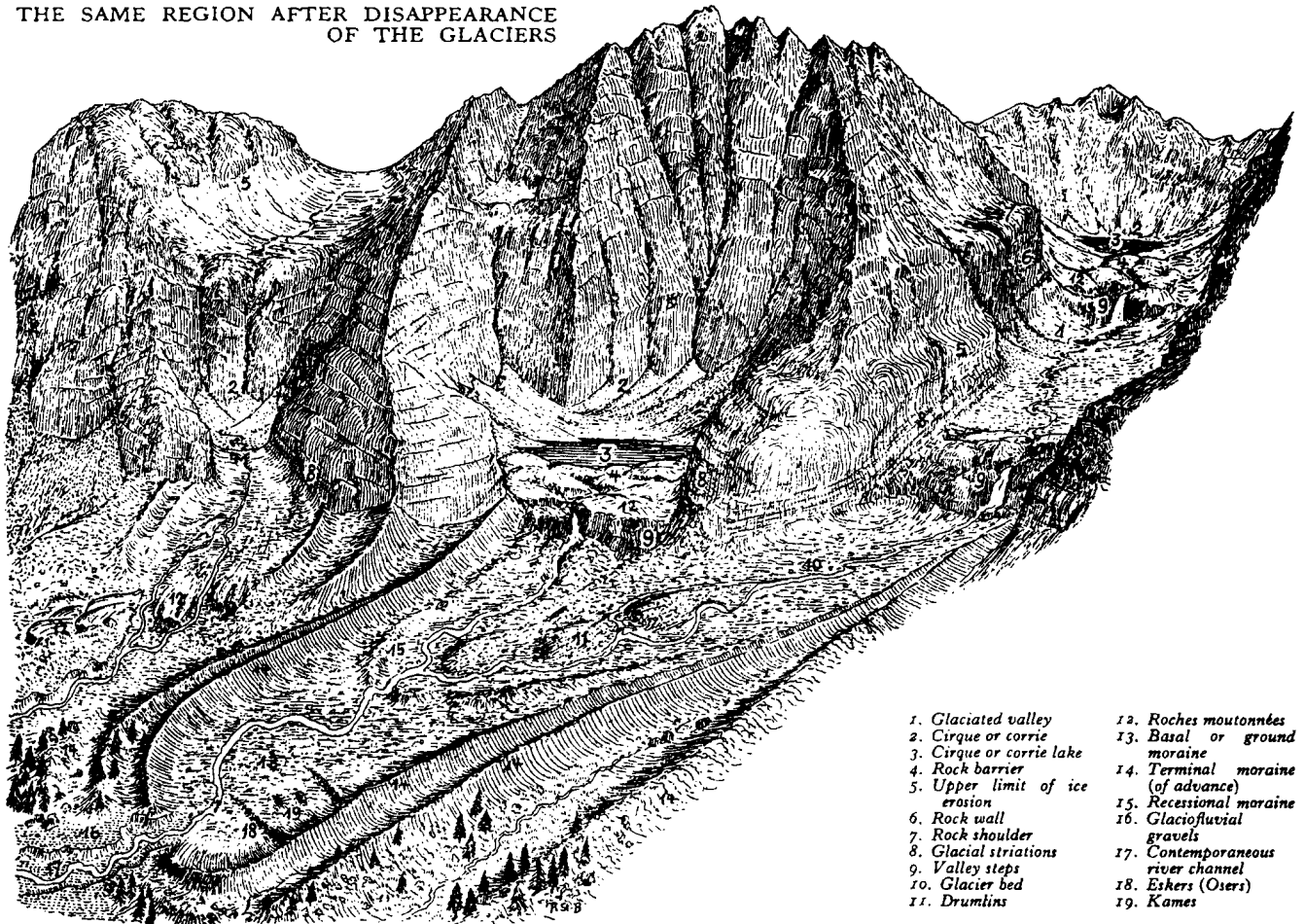
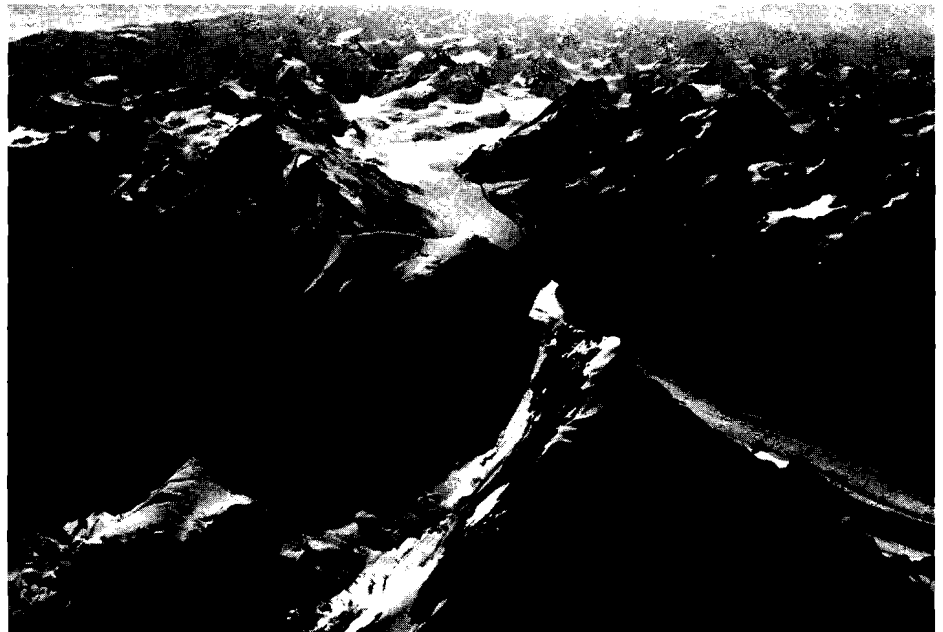


FIGURE 17-5. Glacial landforms following deglaciation of an Alpine region (Streiff-Becker, 1947).

FIGURE 17-6. View northeastward over Mount Cook (3754 m) up the Tasman Glacier, New Zealand. Horns and arêtes are prominent (photo: New Zealand National Publicity Studios).



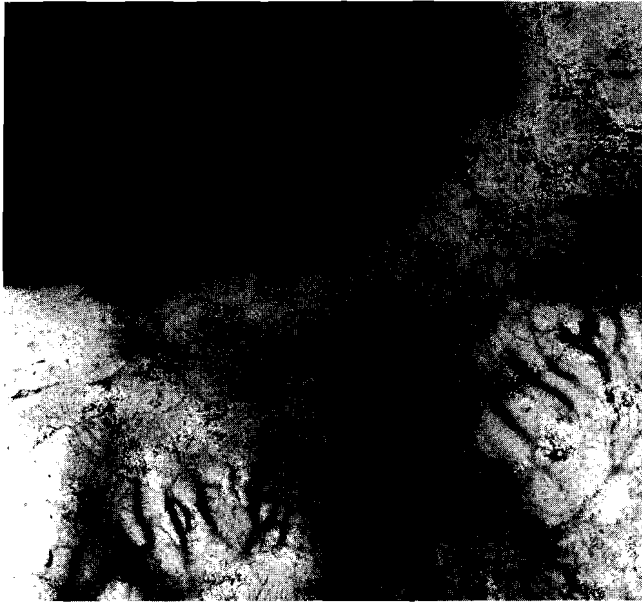


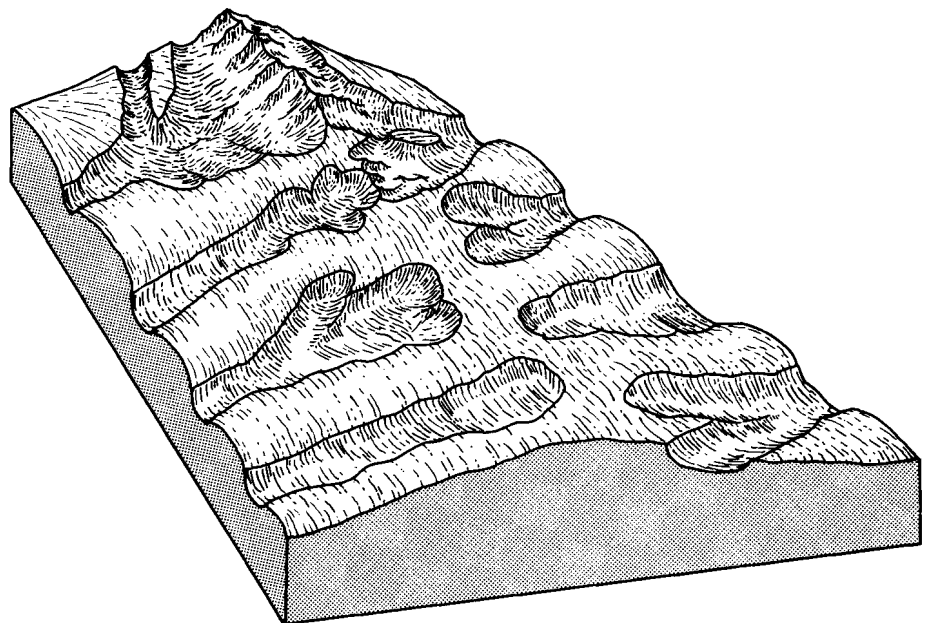
FIGURE 17-7. The Finger Lakes of central New York. All the lakes drain north to Lake Ontario (top) via the Oswego River or Genesee River. Cayuga Lake, the easternmost of the two longest lakes, is 61 km long. A light snow cover enhances the relief (NASA ERTS E-1243-15244).

history of geomorphology. By nature of their position high on mountain slopes, many cirques that fed Pleistocene valley glaciers still carry small cirque glaciers or greatly reduced valley glaciers in them. In North America, they commonly are oriented toward the northeast, most sheltered from the afternoon sun and in a favorable orientation for windblown snow to accumulate (Color Plate 13). Many cirques are compound, with a nested series of progressively smaller and higher forms representing stages in the reduction of icefield size in late-glacial and postglacial time. The altitude of a cirque floor, tarn, or a rock sill is a convenient approximation of the former equilibrium line altitude.

Cols, Arêtes, and Horns. If a plateau or mountain range is lightly but incompletely dissected by glacial troughs and cirques, the former convex upland surface remains as remnants between the steep, concave glaciated forms (Figure 17-8; Color Plate 13). This condition is well shown in the central Rocky Mountains and the Big Horn Mountains, where it has been described as a *scalloped upland*, or *biscuit-board* topography. The latter term describes the landscape well, by analogy to a thick layer of biscuit dough on a cutting board, with arcuate pieces removed by cirque erosion.

Progressive glacier erosion of cirques and troughs leads to the development of sharp-edged divides between them. As two steep-headed cirques intersect, U-shaped notches cut the divide between their head-

FIGURE 17-8. A scalloped upland on a broad, gentle divide (foreground) evolving by intersection of opposing cirques into a *fretted upland* of arêtes, horns, and cols at higher altitudes (background). Compare Color Plate 13.



walls. Such a gap is a **col** in alpine terminology. A saw-toothed divide between opposing or parallel troughs, consisting of a series of cols and intervening ridge segments, is an **arête**. Where three or more cirques intersect, a higher pyramidal or triangular-faceted peak may rise above the general level of the arête. These peaks are the **horns** (or *Matterhorn peaks*, after the type example) of alpine scenery. A summit area of arêtes, cols, and horns represents the complete destruction of the preglacial surface (Figure 17-6), and is analogous to the complete dissection of a regional fluvial landscape. The most prominent horns are at the intersection of three or more radiating arêtes.

Breached Divides and Through Valleys. If mountain glaciation increases in intensity until a mountain ice cap develops, former regional divides may be breached by glacial erosion, and major diversions of preglacial stream flow result. The Brooks Range of Alaska was glaciated more intensely on the southern, snowier side of the regional divide, whereas only short valley or piedmont glaciers formed on the arid northern Arctic Slope. As the ice divide was south of the buried bedrock divide, erosion by the north-flowing glaciers breached the preglacial regional divide in several places, and major ice streams flowed northward down the steep troughs. Breached divides of this sort become strongly abraded and low. In the Brooks Range, the divide at Anaktuvuk Pass, between the Yukon and Arctic drainage, is only slightly over 600 m above sea level and is on a swampy valley floor among moraines (Porter, 1966). In Patagonia, glacial erosion has captured many rivers on the more arid Argentine (eastern) side of the Andes Mountains and diverted them to the wetter Chilean (western) coast.

Similar breached divides in the glaciated northern Appalachian Mountains were produced by multiple advances of the Pleistocene ice sheets that overrode the preglacial divides between the St. Lawrence-Great Lakes drainage and the southerly draining Delaware, Susquehanna, and Ohio Rivers. Powerful erosion through constricted valleys at the base of the ice sheets produced the Finger Lakes troughs (Figure 17-7) on the north side of the divide but left flat-gradient, swampy **through valleys** between north- and south-flowing tributaries (Tarr, 1905). The present divide is inferred to be close to the preglacial position in most areas, but valley forms have been modified into a series of troughs "like the intersecting passages of some unroofed ants' nest" (Clayton, 1965, p. 51; see also Linton, 1963; Coates, 1966). Undoubtedly, much of the erosion of the through valleys can be attributed to meltwater overflowing into the southerly drainage from large ice-

dammed lakes north of the divide at times when the ice margin blocked the St. Lawrence lowland. Each ice advance must have been preceded by an intense period of fluvial erosion in meltwater channels across the divides, and each retreat was accompanied by a similar interval of overflowing ice-marginal lakes.

Regional Denudation

Valley Glaciers. Table 15-2 and the related text present some evidence that the regional denudation rate in mountains with glaciers exceeds 1 m/1000 yr, the most rapid of any climatic region. Even higher rates, up to 50 m/1000 yr, have been calculated for large, fast-moving temperate valley glaciers in southeastern Alaska (Hallet et al., 1996). The combination of high relief, steep slopes, intense nivation, and temperate glaciers must indeed have a powerful impact on landscapes. The share of the work borne by glaciers alone is hard to evaluate because glaciers are just one aspect of geomorphic processes in high snowy mountains (Harbor and Warburton, 1993). It seems certain that alpine glaciation, in which most of the upland is buried under a mountain ice cap from which valley glaciers radiate, must be one of the most rapid means of destroying mountains. The **gipfelflur**, or "sea of peaks" all at about the same elevation, is a well-known feature of alpine scenery (Figure 17-6), but whether the horns that rise above mountain ice fields represent remnants of an ancient erosion surface or are nothing more than the mean level of intersection of adjacent cirques and troughs cannot be determined. Further, in the Holocene Epoch (a postglacial or interglacial interval), glaciers all over the earth have greatly receded, and the high modern denudation rates of mountainous regions are at least partly due to the many oversteepened slopes relict from glacial erosion and deposition but now exposed to mass-wasting and fluvial erosion (Church and Ryder, 1972).

Ice Caps and Ice Sheets. An extreme range of views can be found concerning the effectiveness of large ice sheets in denuding landscapes. White (1972) argued that the multiple Laurentide ice sheets of eastern North America completely stripped up to 1 km of Paleozoic and younger sedimentary rocks off large areas of the Canadian shield. Other authors (Ambrose, 1964) have noted that the buried relief of the crystalline shield, where it is unconformably overlain by lower Paleozoic or Cretaceous sedimentary rocks, is very similar to the local relief on the shield today. By this line of reasoning, multiple Pleistocene glaciations have done little more than to remove the weathered regolith (Dyke et al., 1989, p. 182). The oldest known glacial deposits in the

central United States contain igneous and metamorphic rocks derived from the Canadian shield, proving that the shield was exposed at least in part prior to the onset of Quaternary glaciation (Gravenor, 1975). The view of minimal regional denudation by ice sheets is supported by the rather small volume of glacial deposits on the periphery of the Laurentide ice sheet and by the fact that most of the glacial sediment was derived from relatively near the ice margin, especially in the southern quadrant in the United States. From the volume of glacial sediment on the North American continent, many authors have argued that a total of no more than 20 m was eroded from the Canadian shield during the Quaternary ice ages. Only 1 to 2 m of resistant shield rock was removed by the latest glaciation although as much as 6 m might have been removed by the earliest ice sheets, which eroded the preglacial regolith (Kaszycki and Shilts, 1987, p. 158). Such rates are of the same order as the rates of fluvial erosion expected for shield terranes (Figure 15-2). They imply that ice sheet erosion is not intense, and that ice sheets could even be regarded as "protection" from fluvial erosion. A less confident estimate is that a total of 120 m has been removed from the glaciated portion of North America by Pleistocene glacial erosion, and that most of the volume is now in the deep sea, not on land. This estimate is based on many assumptions of deep-sea sediment volumes, grain-size distribution, and age (Bell and Laine, 1985). Ice sheets do not seem to be dramatic erosive agents.

Discoveries of bauxite pebbles in glacial deposits on Martha's Vineyard, Massachusetts (Kaye, 1967); deep weathering on Long Island, New York, that is characteristic of the humid tropics (Liebling, 1973; Blank, 1978); and remnants of saprolite in New England and Quebec (LaSalle et al., 1985) all support the inference that in preglacial times, weathering typical of the present tropics or subtropics affected northeastern North America (Chapter 4). Such weathering is known to affect rocks to depths of 100 m or more (Figure 7-14), so it is possible that glacial erosion has accomplished little more than to remove the former regolith from northeastern North America. The very large crystalline **erratics** derived from the Precambrian shield, but widely scattered in southern Canada and the glaciated United States east of the Rocky Mountains, need not represent erosion into fresh, unweathered rock; they are more likely corestones (Figures 7-1 and 7-13) from the preglacial weathered layer (Feininger, 1971). Minimal glacial erosion into resistant rock masses is also inferred from the conformity of roches moutonnées around Boston, Massachusetts, to sheeting joints that are almost certainly preglacial (Jahns, 1943) and from

similar hills in Sweden that have analogs in the relief on sub-Cretaceous and sub-Cambrian unconformities (Lindström, 1988). The nearly total lack of rock debris in the tabular icebergs derived from Antarctica implies that glacial erosion is now inactive there (Warnke, 1970). However, the thick deposits of glaciomarine sediments around Antarctica prove that at some time in the Cenozoic Era, the Antarctic ice sheet eroded a considerable mass off the continent, including a deep regolith (p. 58).

GLACIAL DEPOSITION

Several traditional terms have been used to describe rock debris that is both in transport by modern glaciers and is inferred to have been deposited by former glaciers. The oldest word is *moraine*, used to describe the hummocky or ridged depositional landscapes beyond the margins of alpine glaciers and sometimes also to describe the rock debris in and on glaciers today (Figures 17-4 and 17-5). Some confusion results if the word *moraine* is used to describe both the material and the landforms, so only the latter meaning is used here. **Moraine**, then, is simply a collective term for all depositional landforms of direct glacial origin. One possible classification of moraine and related glaciofluvial forms is outlined on the following pages.

For centuries before the glacial theory of Agassiz (1840), the poorly sorted, unconsolidated surficial deposits of northern Europe were supposed to have been the result of the biblical flood of Noah's time. In Germany, the term *diluvium* was widely used with reference to the "great flood." In the British Isles, the same material was for the same reason called *drift*, a term also used to describe marine sediment in transport along coasts. With the acceptance of the glacial theory to explain these deposits, **glacial drift** (or simply **drift** if no ambiguity is possible) has come to be the general English term for all sediment transported and deposited by glaciers and associated fluvial activity. It is important not to confuse landforms and materials when studying glacial geomorphology, so *drift* is used consistently to describe glacier-transported sediment, and *moraine* is used as a general term for glacier-depositional landforms.¹ The value of the distinction should become apparent in the following pages.

1. See the discussion on p. 12 concerning the analogous use of the terms *terrane* and *terrain*.

Glacial Drift

The following brief glossary defines some of the more common terms for categories of glacial drift and associated sediments. Details of composition, sorting, stratification, and genesis are left to specialized texts (Evenson et al., 1983; Eyles, 1983; Drewry, 1986; Goldthwait and Matsch, 1989; Menzies, 1995, 1996), but this vocabulary is necessary to appreciate the genesis and morphology of morainic and related glaciofluvial landscapes.

1. **Till.** Nonsorted, nonstratified glacial drift, deposited directly from ice without reworking by meltwater. The nongenetic term for any mixed sediment, **diamicton**, is recommended for poorly sorted debris that cannot be proved to be glacier-derived. Some authors object so much to the unique glacial origin implied by the term *till* that they use only the term *diamicton* for all poorly sorted till and till-like sediments, including the following subcategories and many others (Dreimanis, 1989).
 - a. *Lodgment till*—Compact, tough, dense till, often clay-rich, deposited at the base of a flowing temperate or wet-base glacier. Stones may show surface abrasion and preferred orientation parallel to former flow.
 - b. *Melt-out till*—Sediment released from ice by melting. *Basal melt-out till* can be deposited wherever the heat generated by friction, deformation, or geothermal heat flow cannot be conducted rapidly enough through overlying ice. It is compact and nonsorted, but is not sheared by ice flow. *Surface melt-out till* can be derived from either supraglacial, englacial, or subglacial debris that has arrived at the glacier surface by ablation. It is most common in polar glaciers, where ablation is primarily by sublimation. On temperature glaciers, surface sediments are usually reworked by water.
 - c. *Resedimented till*—Surface melt-out till commonly slides, slumps, or flows as it accumulates on ablating ice, especially if it is water saturated. While retaining most properties of till, resedimented till may show crude layering or deformation structure. It is transitional to the following categories of glaciofluvial, or water-laid, drift.
2. **Ice-contact stratified drift.** Drift modified by meltwater during or after deposition. Considerable stratification and sorting is usual. Although water-modified, the large range of sizes, chaotic sedimentary structures, slump structures, faults,

surface topography, and inclusions of till all confirm an origin in contact with, or in close proximity to, melting ice.

3. **Outwash.** Alluvium deposited by meltwater not in close proximity to melting ice. Sorting and stratification are more regular than in ice-contact stratified drift. Similar to other coarse alluvium, except as influenced by the extreme variability of discharge from glacial sources.

In addition to these three categories of drift, the following sediment types are also commonly associated with glaciation:

4. **Glaciolacustrine sediment.** Deltaic and lake-bottom deposits, usually fine grained, formed in settling basins on or near melting ice. Commonly forms silt and clay plains of broad extent, with distinctive erosional and soil-forming properties. Sometimes *varved*, or composed of rhythmic silt-clay couplets each representing an annual cycle of deposition in a seasonally frozen lake.
5. **Eolian sediment.** Sand dunes, sand sheets, and loess derived from other types of glacial drift. Genetically undifferentiated from other eolian sediments except by regional and temporal association with other drift (Derbyshire and Owen, 1996).
6. **Glaciomarine sediment.** "Glacially eroded, terrestrially derived sediment within a variety of facies in the marine environment. Sediment may be introduced into the depositional environment by fluvial transport, by ice-rafting, as an ice-contact deposit, or rarely by eolian transport" (Molnia, 1983, p. 136).

Moraines and other Constructional Drift Landforms

Some moraines and constructional glaciofluvial landforms are genetically related to a specific kind of drift. Examples are *till plains*, which by definition are low-relief areas of eroded or deposited till; *eskers*, which by origin in an ice tunnel or channel must be made of ice-contact stratified drift; or *outwash plains*, which are depositional plains of outwash. Other terms for constructional glacial landforms do not indicate the nature of the drift from which they are built. For example, *end moraines* range in composition from entirely till to entirely ice-contact stratified drift. Almost all moraines contain some mixture of these materials.

In an attempt to sort out the nomenclature of moraines, Prest (1968) proposed a classification based on the orientation of morainic landforms with respect to ice movement (Table 17-1). The classification is

Table 17-1

Classification of Moraine^a

Transverse to Ice Flow (Controlled Deposition)	Parallel to Ice Flow (Controlled Deposition)	Nonoriented (Uncontrolled Deposition)
Ground moraine—corrugated. End moraine; includes terminal, recessional, and push moraine. Ice-thrust moraine. Ribbed moraine. De Geer moraine; includes cross-valley moraine. Interlobate and kame moraine. Linear ice-block ridge; includes ice-pressed and ice-slumped ridge, and crevasse filling.	Ground moraine—fluted and drumlinized; includes drumlin, drumlinoid ridge, crag-and-tail hill. Marginal and medial moraine. Interlobate and kame moraine. Linear ice-block ridge; includes ice-pressed and ice-slump ridge, and crevasse filling.	Ground moraine—hummocky (low relief); includes some ablation moraine. Disintegration moraine—hummocky and/or pitted (mainly high relief); includes dead ice, stagnation, collapse, and ablation moraine. Interlobate and kame moraine. Irregular ice-block ridge and rim ridge; includes ice-pressed and ice-slump ridge, and ablation slide moraine.

Source: Prest (1968).

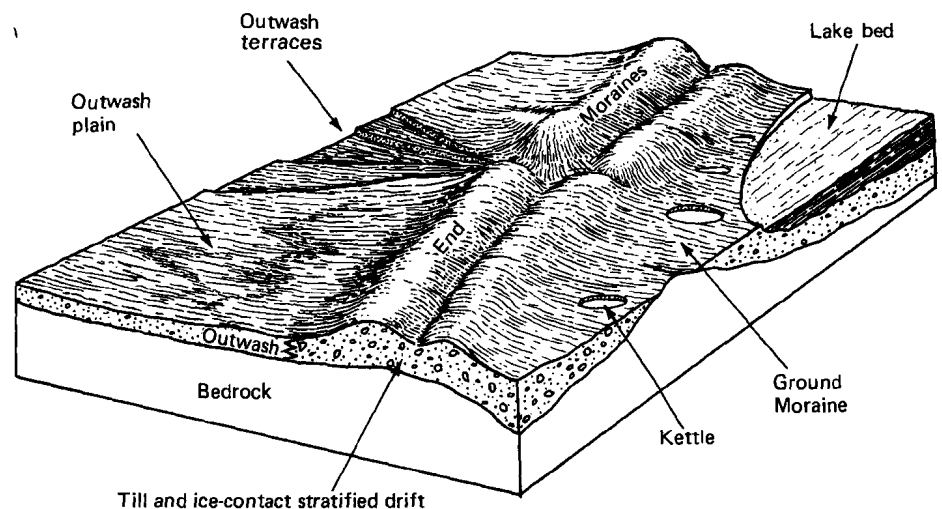
^aIncluded are landforms constructed from till or other drift that was deposited directly from, or in association with, glacier ice.

intended for landforms built by continental ice sheets but includes forms built by valley glaciers as well. The scheme is vaguely similar to the classification of sand dunes (Figure 13-20) in that longitudinal (parallel to ice flow), transverse (perpendicular to ice flow), and nonoriented forms are included. The transverse forms are mostly ice-marginal deposits, built during the last recession of the ice margin, but include some forms that are shaped beneath ice during ice advance. The forms that parallel ice flow, by contrast, are nearly all controlled by actively moving ice, although some accumulate in stagnant, melting ice along channels and crevasses parallel to former flow directions. The nonoriented forms are commonly built of a mixture of melt-out till and ice-contact stratified drift, during a time of regional melting with insignificant control by

ice motion. Longitudinal forms are likely to predate transverse and nonoriented forms in the same region. For instance, younger *end moraines* or *kame moraines* may be banked against, or draped over, *drumlins* and fluted (corrugated) ground moraine. Some of the morainic terms apply equally well to transverse, longitudinal, and nonoriented forms. For example, melt-out till pressed into the base of a crevassed and melting ice sheet may produce ridges oriented in several directions or produce a nonoriented maze of ridges and mounds. Drift is a notoriously mixed sediment; within a single moraine (Figure 17-9), deltaic structures, lodgment till, and deformed ice-contact stratified drift may be found close together.

Glacial landforms tend to occur in characteristic regional patterns because the terrane over which the

FIGURE 17-9. The concept of a glacial series or morphosequence. A group of morainic and related glaciofluvial landforms are either synchronous or sequential in time and are genetically related.



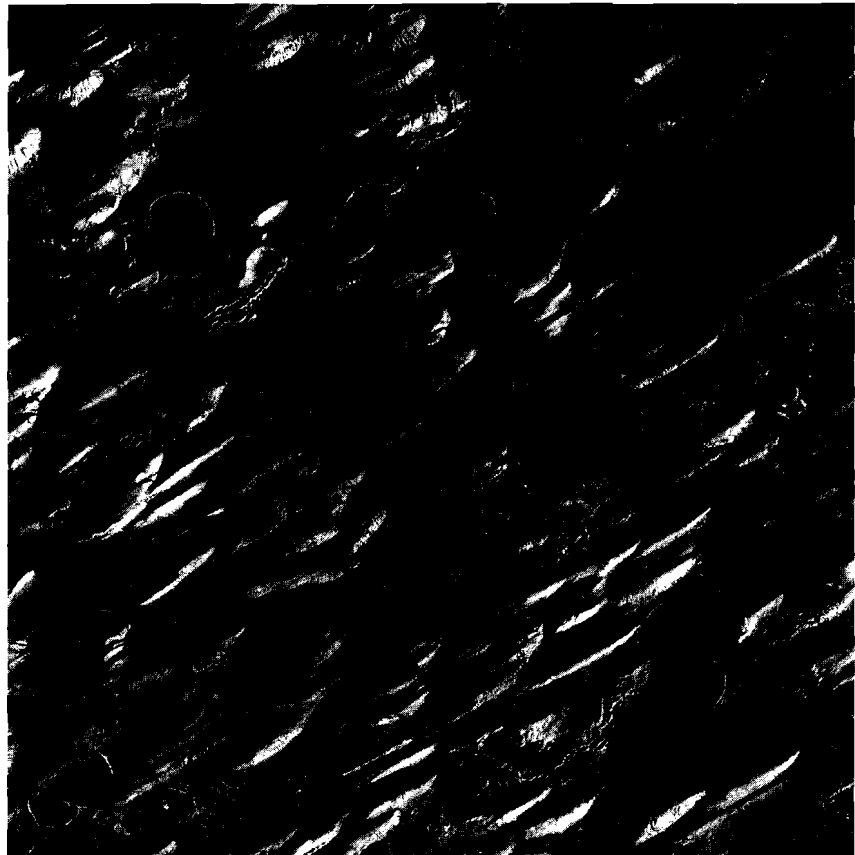
glacier moved and the ablation history largely determine the regional character of both the drift and the landforms. Many morainic forms can be assembled in a genetic sequence that was called a *glacial series* (Figure 17-9) by early workers (Penck and Brückner, 1909; von Engel, 1921), and later incorporated into the *morphosequence* concept (Koteff and Pessl, 1981, p. 5). Each morphosequence refers to a continuum of landforms comprised of glaciofluvial sediments, grading from ice-contact environments to distant outwash. All the forms in a morphosequence are regarded as contemporary and a chronology can be built from successive morphosequences. The principle is a good one. At any one time during the uncovering of a landscape by ice, an end moraine may be forming at the ice margin, fronted by an outwash plain. Ice tunnels and crevasses behind the moraine may be the sites of eskers, crevasse fillings, and kames. Meltwater may be traced through an eroded channel across the end moraine to a fan or delta that coalesces with the outwash plain in front of it. The ice-contact, or *proximal*, part of the end moraine is of till or ice-contact stratified drift. On the *distal* or outwash side of the end moraine, the drift is mostly outwash, with inclusions of ice-contact stratified drift or resedimented till. The entire assemblage represents contemporary events and, when understood as a unit,

gives a good concept of the activity along a receding glacier terminus.

Morainic Landforms Constructed Primarily of Till.

Many of the ice-parallel morainic landforms are streamlined, and when developed in clusters give a vivid impression of ice motion. Similar erosional forms, such as roches moutonnées and crag-and-tail hills, have been described previously (p. 376). Best known of the streamline ice-molded forms are **drumlins** (Muller, 1974). These are elliptical hills, blunt on the upglacier end, with an elongate downglacier tail (Figure 17-10). They occur in clusters, most frequently in areas of fine-grained lodgment till. Internal composition is not a criterion for calling a streamlined hill a drumlin, and known examples range from those that are nearly all rock with a thin veneer of till to those that are entirely till with no rock core. Classic examples are to be seen in northern Ireland, New Brunswick, eastern Massachusetts, southern Ontario, central New York, and southern Wisconsin. In each region, hundreds or thousands of drumlins give the landscape a distinct "grain," or linear trend. The drumlins tend to be of comparable internal structure, size, and shape within each region. The similar streamline shape of drumlins of diverse internal structure illustrates that they are stable forms

FIGURE 17-10. Drumlin field southeast of Lake Athabasca, Saskatchewan, Canada. Ice flowed from the northeast (upper right). Also shown are numerous kettle lakes and a system of anastomosing eskers. Lakes are approximately 1 km in diameter. Low-angle illumination is from the southeast. If the drumlins appear inverted, view the picture upside down (photo: Surveys and Mapping Branch, Dept. of Energy, Mines, and Resources, Canada).



in equilibrium with local conditions at the base of moving ice. The concept of *equifinality*, the tendency of diverse processes to produce similar landforms (p. 14), is well illustrated by drumlins and related streamline erosional forms (compare Figures 17-1 and 17-10). Table 17-1 lists as variants such forms as fluted (ridged) ground moraine (flutes) and elongate "drumlinoid" ridges.

A hypothesis of "dilatant flow" within water-saturated basal till has been proposed to explain drumlin formation (Smalley and Unwin, 1968; Smalley, 1981). A mixture of rock debris, ice, and water, similar to the deformable sediment bed beneath some wet-base glaciers (p. 368), could flow as a viscous fluid as long as it retains water. If, however, an increased pressure region upglacier of a bedrock obstruction or even of a large stone within the mass were to cause local dewatering, the mixture would stiffen and lodge. A similar effect could result if a mass of coarser or better-sorted sediment were dewatered by shear within enclosing viscous sediment (Boulton and Hindmarsh, 1987). The initial obstruction would be enlarged by the mass of rigidified till and continue to grow to some equilibrium streamlined form or until another obstruction in the vicinity caused a similar response. The hypothesis has the advantage of explaining the size, shape, compositional variation, and spatial distribution of drumlins. The process requires a wet-base glacier and a deformable substrate. Local pressure and temperature gradients at the base of the ice, associated with extending and compressing flow, may localize drumlin forma-

tion in certain zones broadly parallel to the ice margin. A stimulating alternative hypothesis is that drumlins are the streamlined remnants of former drift eroded by subglacial meltwater floods (p. 377).

Landforms Constructed Primarily of Ice-Contact Stratified Drift. A great variety of distinctive landforms are made of ice-contact stratified drift. Oriented and nonoriented forms are both common. The terminology is complicated because various terms can be applied to either single landforms or groups of forms.

Kame is representative of the complexities of this category of landforms. A kame is an isolated hill or mound of stratified drift deposited in an opening within or between ice blocks or in a moulin (Figure 17-11). If the opening was between ice and a valley wall, the resulting form is called a **kame terrace**. **Kame deltas** are flat-topped hills with an ice-contact proximal face and a semicircular plan view, deposited in standing water at an ice margin. A **moulin kame** is believed to form within stagnant ice at the base of a *moulin* or glacier-karst swallow hole (p. 358). It is clear that a variety of environments in or adjacent to melting glaciers can produce kames.

Another important landform of ice-contact stratified drift is the **esker** (Figure 17-10). An esker is a sinuous or meandering ridge of stratified sand and gravel, commonly with very well-rounded pebbles and cobbles. Eskers form in open channels on glaciers and in tunnels within or beneath them. If large quantities of

FIGURE 17-11. Internal structure of a kame near West Durham, Maine. Coarse gravel foreset beds at bottom were deposited by a prograding delta among ice blocks. Overlying glaciofluvial sand and gravel were probably deposited by an aggrading meltwater river.



sediment-laden meltwater are discharging through a karstlike hydrologic system within a glacier, the sediment may be deposited on the floors of various tunnels or channels while the ice walls or roofs are widened or raised by melting. Eskers show sinuosities comparable to the dimensions of rivers and clearly show by internal structures that they are water-laid. However, they cross bedrock ridges, rising and falling as they are traced across moraine landscapes. Confined flow driven by the pressure gradient within the overlying ice must be invoked to permit them to be deposited on uphill gradients. Most trend broadly perpendicular to major end moraines and parallel to former ice flow and ice surface gradient because water flowed under hydrostatic head toward the ice margin (Shreve, 1985) (Figure 16-15). Eskers seem to be systematically related to regions of crystalline bedrock with discontinuous, permeable drift cover, rather than with regions of fine-grained till (Clark and Walder, 1994). The esker systems of northeastern United States, eastern Canada, and southern Finland and Sweden are noteworthy.

A **kettle** is a basin of nondeposition within morainic terrain, formed where an ice block was buried by till or ice-contact stratified drift and then melted. Many areas of moraine are chaotic mazes of kettles among kames and **crevasse fillings**, the latter representing sediment castings made in ice molds that then melted. Extremely complex internal structures, representing multiple episodes of melting and drift deposition, characterize many areas of kames and kettles (Figure 17-12). Small kettles develop angle-of-repose side slopes and circular outlines very similar to those of collapse dolines (Figure 8-10). Larger ones are complexly segmented by crevasse fillings (Figure 17-10). Kettle lakes are favorite scenic attractions and a fisherman's paradise. Thoreau's Walden Pond may be the best known.

Landforms Constructed Primarily of Outwash. Meltwater alluvium aggrades the floors of glaciated valleys downstream from melting glaciers. Meltwater streams have steep gradients and braided patterns appropriate for their massive bedloads and fluctuating discharges. Where outwash is confined between valley walls downstream from a glacier, the aggraded surface is called a **valley train** (Figure 17-13). On the margin of an ice sheet in areas of low relief, broad **outwash plains** form from confluent alluvial fans that emerge from multiple meltwater valleys through end moraines. The Icelandic term **sandur** (plural: sandar) is sometimes applied to outwash plains. The Icelandic sandar are notable for major episodic floods that inundate hundreds of square kilometers during events called *jökulhlaups* ("glacier bursts"), when ice dams are floated by subglacial water pressure and release masses of impounded water from the margins of

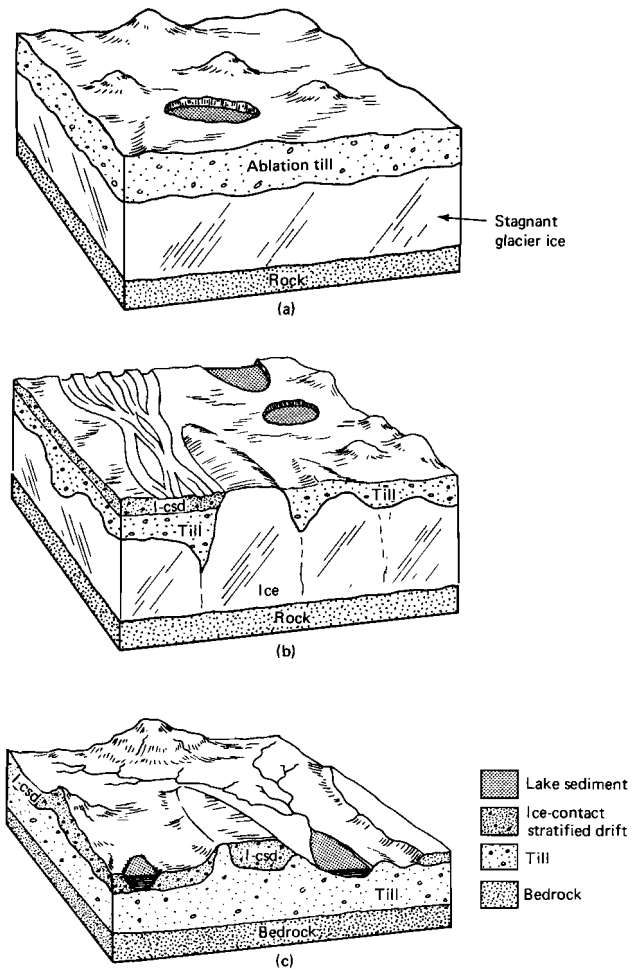


FIGURE 17-12. Hypothetical sequence of events in the evolution of a flat-topped kame. (a) Cover of melt-out till over stagnant ice. (b) Till slumps and washes into deepening crevasses, exposing ice on a chaotic ablation surface. (c) Glaciofluvial sediment washed into a depression when ice melted, later became the crest of a kame, surrounded by a rim of till. Other ice-contact stratified drift deposits are slumped and tilted by final melting.

mountain ice caps. Active valley trains and outwash plains are excellent source areas for dune sand and loess (Figure 13-26), because they carry massive sediment loads, but may nearly dry up at night or during winter, when melting is slower. Eolian deflation may then remove sand and silt from the dry river bed, but the sediments are renewed with each runoff event.

Landforms Constructed of Glaciolacustrine, Eolian, and Glaciomarine Sediments. When ice-marginal lakes drain, a variety of minor depositional landforms remain. The deeper parts of the lake will typically become a lacustrine plain of relatively fine-grained sediment. Slumps and mudflows may accompany the

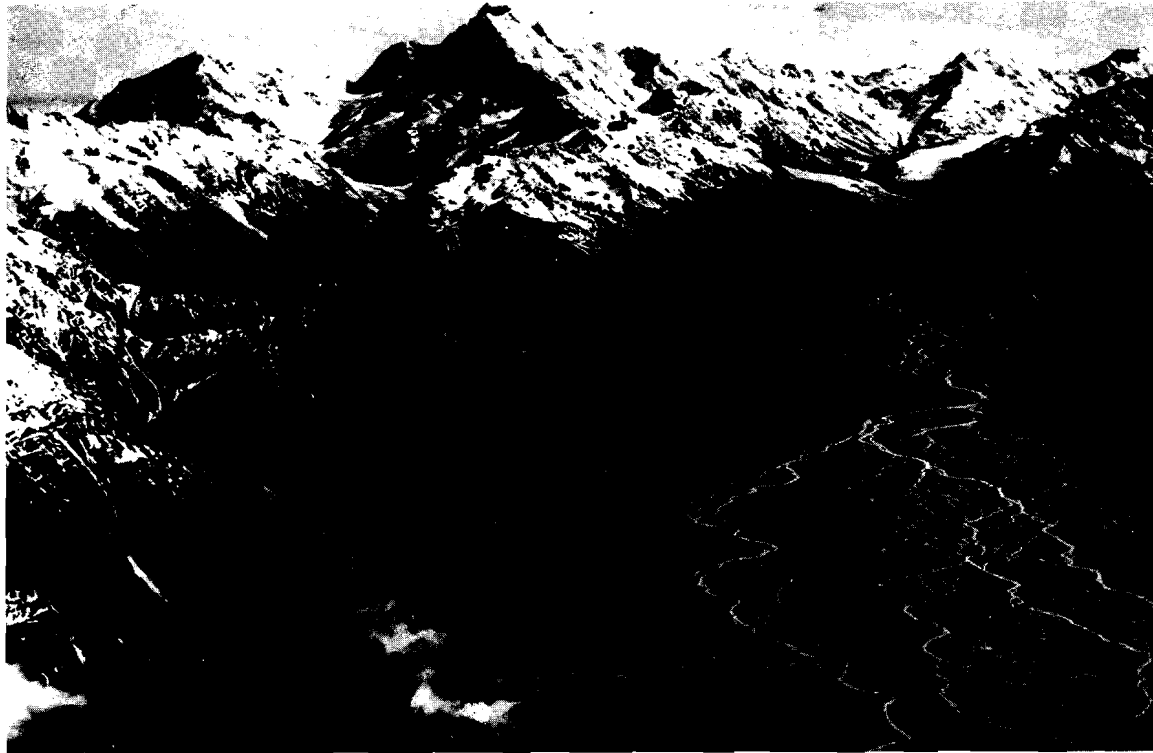


FIGURE 17-13. View upstream over the braided valley train of the Tasman River, South Island, New Zealand. Tasman Glacier, mostly covered with melt-out till, in right distance. Mount Cook is high peak on the left (photo: V. C. Brown & Son, Photographers).

draining of the lake. Beach ridges and deltas may define former shorelines. Eolian landforms include dune fields, sand plains, and loess sheets that drape over or bury prior topography (Chapter 13). Because of isostatic uplift, extensive areas of emerged glaciomarine sediments are common near coasts. They create a complex set of landforms that include submarine end moraines, ice-contact deltas, eskers, and emerged beach ridges (Figure 17-14). The quick-clay terranes of glaciomarine sediments are noteworthy for their slope stability problems (Figure 9-9).

INDIRECT GEOMORPHIC EFFECTS OF GLACIATION

Glacial Isostasy

A special form of epirogenic movement of particular significance in the Pleistocene Epoch is **glacial isostasy**. The theory of isostasy holds that the pressure at the base of all vertical sections of the earth, above some deep level of compensation, is equal. Nature's "great experiments in isostasy" (Daly, 1940, p. 309) were the alternate

loading and unloading of large continental areas with ice sheets during late Cenozoic time (Figure 17-15). Seawater must evaporate and accumulate on land as ice in order for ice sheets to grow. Thus each glaciation involved the mass transfer of a layer of water perhaps 120 m in thickness from the 71 percent of the earth's surface that is ocean, and the concentration of that water mass, as glacier ice as much as 3 to 4.5 km in thickness, on about 5 percent of the earth's surface. We infer that the ocean floor responded to the removal of water load by isostatically rising (Tushingham and Peltier, 1991). It is much easier to prove that the glaciated areas larger than about 50 km in diameter were basined beneath their loads of ice because when the ice was melting, lakes and seas often formed shorelines near the glacier margins. The shorelines of these former ice-marginal water bodies show regular tilting, always concentric upward toward the center of former maximum ice thickness. Uplift of 120 m is known from areas on the eastern side of Hudson Bay, Canada, in a total time of less than 6000 years (Andrews, 1991, Figure 10).

The rate of postglacial uplift is rapid at first, then progressively slower. It approximates an exponential decay curve with a *relaxation time* (the time required for the deformation to decrease to $1/e$, or 37 percent, of

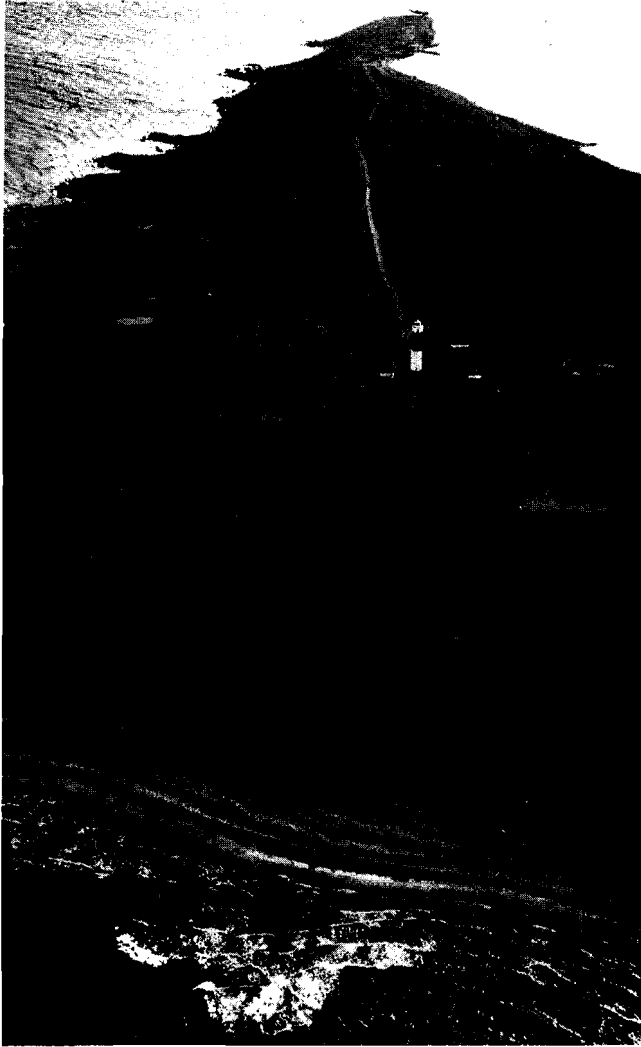


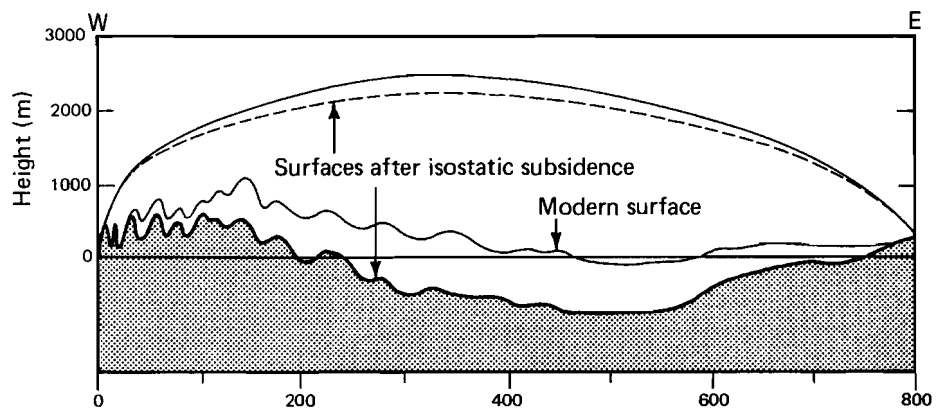
FIGURE 17-14. Emerged Holocene beach ridges on Ostergarnsholm, eastern Gotland, Sweden. The contemporary uplift rate is about 2 mm/yr (Figure 17-16) (photographer and copyright holder: Arne Philip, Visby, Sweden. Courtesy IGCP Project Ecostratigraphy).

the original value) of from 1000 to 5000 years. Initial uplift rates of 9 to 10 cm/yr have been reported for Greenland (Washburn and Stuiver, 1962; Ten Brink, 1974) and for arctic Canada (Andrews, 1991). The maximum known current uplift rate is 9 mm/yr, recorded near the north end of the Baltic Sea (Figure 17-16) (Kakkuri and Chen, 1992).

Where uplift rates are so rapid, landforms such as storm beaches are raised in a few decades sufficiently high to be unreached by the next major storm, and wide, descending flights of beach ridges mark the margins of seas and lakes (Figure 17-14). River gradients must also be affected by the tilting associated with isostatic uplift, and lakes or swamps may have formed or drained, but the extremely deranged drainage patterns of glaciated regions make the subtleties of tilting hard to recognize. Rivers flowing away from the center of uplift should show simultaneous rejuvenation as their gradients are increased by tilting; rivers flowing toward the center of maximum uplift, as do many of the rivers that drain into Hudson Bay, should have their gradients reduced by the greater differential uplift of their lower reaches; this contributes to the abundance of lakes and swamps in glaciated regions. The differential tilt on postglacial shorelines is typically of the order of 1 m/km, or 0.1 percent, which is comparable to the gradient of many river segments, so the effect on stream hydrology cannot be neglected.

The Great Lakes of North America form a natural laboratory for studying postglacial uplift and tilting (Teller, 1987). Abandoned shorelines and contemporary water-level gages prove that the lakes are deepening on their southern shores and shoaling on the northern. For example, Lake Michigan and Lake Huron form a continuous water level through the very deep Straits of Mackinac. The northeastern parts of the combined lakes are now rising about 30 cm/100 yr relative to their southwestern parts (Figure 3-7). Furthermore, the outlet of the combined lakes near Detroit is rising at

FIGURE 17-15. Estimated surface profile of the Fennoscandian ice sheet along a line eastward from Trondheim, Norway, to Helsinki, Finland. Note the very large area of ice that was grounded below sea level in the Baltic basin (Robin, 1964, Figure 7).



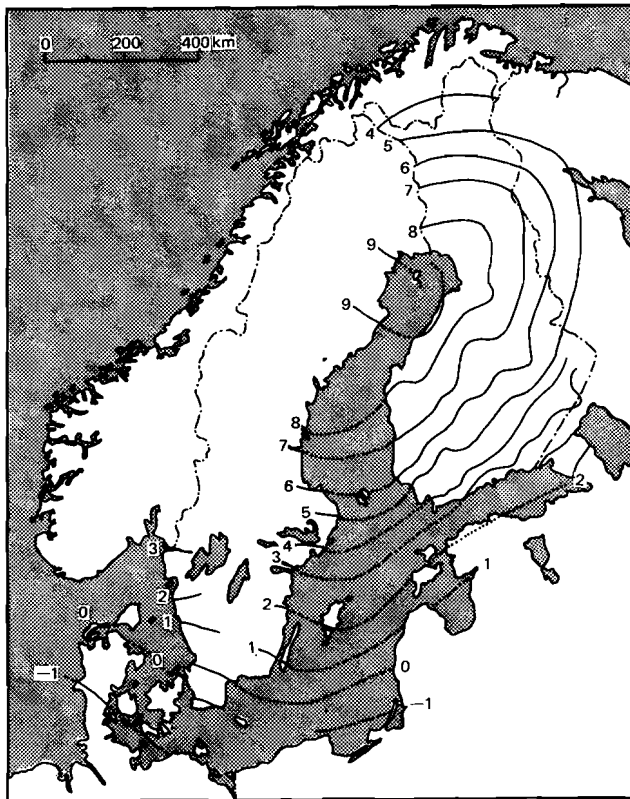


FIGURE 17-16. Contemporary emergence rates in the Baltic basin and vicinity (mm/yr). Emergence is relative to sea level, which is also rising 1 mm/yr or more (Flint, 1971, Figure 13-2).

about 7.5 cm/100 yr relative to a former glacial overflow at Chicago. The discharge of Lake Michigan and Lake Huron will be diverted naturally from the St. Lawrence system to the Mississippi in the next 3200 years if control dams in the Chicago Sanitary and Ship Canal do not prevent the changes (Clark and Persoage, 1970, p. 632). The magnitude of the potential change suggests that widespread regional drainage diversions are possible by postglacial uplift. However, it should be emphasized that the uplift is a *recovery*, or a return to a previous condition, and not the equivalent of net tectonic uplift.

Glacial Eustatic Fluctuations of Sea Level

The volume of water abstracted from the sea at times of ice-sheet expansion and returned to the sea during interglacial intervals of the Pleistocene Epoch has had a worldwide effect on landscape evolution. Global changes of sea level are called **eustatic** changes, meaning that the entire ocean is affected; if they are due to fluctuating ice volume, they are called **glacial eustatic**

sea-level changes. Estimates of the maximum glacial eustatic lowering of sea level during the latest full glacial interval, as compared to today, vary from 80 to 140 m, with the best estimate being about 120 m (Figure 18-4). The lowering was enough to lay bare about one-half of the continental shelf area so that all rivers had to extend and deepen their valleys to the lower base level. At present, all rivers enter the sea either in estuaries or across thick deltaic fills. Both features are a direct consequence of glacial eustatic sea-level fluctuations but are so much a part of the late Cenozoic fluvial geomorphic system that we find it hard to deduce what river mouths would be like if sea level had been stable for the last few million years (Chapter 11).

Glacial eustatic fluctuations of sea level have affected all the coastal regions of the earth, with much greater geomorphic impact than postglacial uplift, which principally affected only the deglaciated areas. The rising sea level of the last 20,000 years has submerged all coasts except those with rapid glacial-isostatic or tectonic uplift. During the most rapid phases of deglaciation, from about 15,000 to 7000 years ago, sea level may have risen at times at a rate of 10 mm/yr. The consequences of this submergence to coastal morphology are considered at length in Chapters 19 and 20.

REFERENCES

- ABER, J. S., CROOT, D. G., and FENTON, M. M., 1989, *Glaciotectonic landforms and structures*: Kluwer Academic Publishers, Dordrecht, The Netherlands, 200 pp.
- AGASSIZ, L., 1840, *Études sur les glaciers*: privately published, Neuchâtel, Switzerland, 346 pp.
- ALLEY, R. B., 1989, Water-pressure coupling of sliding and bed deformation: II. Velocity-depth profiles: *Jour. Glaciology*, v. 35, pp. 119-129.
- , BLANKENSHIP, and 2 others, 1989, Water-pressure coupling of sliding and bed deformation: III. Application to Ice Stream B, Antarctica: *Jour. Glaciology*, v. 35, pp. 130-139.
- AMBROSE, J. W., 1964, Exhumed paleoplains of the Precambrian shield of North America: *Am. Jour. Sci.*, v. 262, pp. 817-857.
- ANDERSON, J. B., and ASHLEY, G. M., eds., 1991, *Glacial marine sedimentation; Paleoclimatic significance*: *Geol. Soc. America Spec. Paper* 261, 232 pp.
- ANDREWS, J. T., 1991, Late Quaternary glacial isostatic recovery of North American, Greenland, and Iceland: A neotectonics perspective, in Slemmons, D. B., Engdahl, E. R., and 2 others, eds., *Neotectonics of North America*: *Geol. Soc. America Decade Map Vol. 1*, pp. 473-486.
- ASHLEY, G. M., SHAW, J., and SMITH, N. D., eds., 1985, *Glacial sedimentary environments*: *Soc. Econ. Paleontologists and Mineralogists Short Course No. 16*, 246 pp.

- AUGUSTINUS, P. C., 1992, The influence of rock mass strength on glacial valley cross-profile morphometry: A case study from the southern Alps, New Zealand: *Earth Surface Processes and Landforms*, vol. 17, pp. 39–51.
- , 1995, Glacial valley cross-profile development: The influence of in situ rock stress and rock mass strength, with examples from the Southern Alps, New Zealand: *Geomorphology*, v. 14, pp. 87–97.
- BELL, M., and LAINE, E. P., 1985, Erosion of the Laurentide region of North America by glacial and glaciofluvial processes: *Quaternary Res.*, v. 23, pp. 154–174.
- BLANK, H. R., 1978, Fossil laterite on bedrock in Brooklyn, New York: *Geology*, v. 6, pp. 21–24.
- BOULTON, G. S., 1974, Processes and patterns of glacial erosion, in Coates, D. R., ed., *Glacial geomorphology*: State Univ. of New York Publications in Geomorphology, Binghamton, New York, pp. 41–87.
- , 1996, Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation: *Jour. Glaciology*, v. 42, pp. 43–62.
- , and Hindmarsh, R. C. A., 1987, Sediment deformation beneath glaciers: Rheology and geological consequences: *Jour. Geophys. Res.*, v. 92, pp. 9059–9082.
- CHURCH, M., and RYDER, J. M., 1972, Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation: *Geol. Soc. America Bull.*, v. 83, pp. 3059–3072.
- CLARK, P. U., and WALDER, J. S., 1994, Subglacial drainage, eskers, and deforming beds beneath the Laurentide and Eurasian ice sheets: *Geol. Soc. America Bull.*, v. 106, pp. 304–314.
- CLARK, R. H., and PERSOAGE, N. P., 1970, Some implications of crustal movement in engineering planning: *Canadian Jour. Earth Sci.*, v. 70, pp. 628–633.
- CLAYTON, K. M., 1965, Glacial erosion in the Finger Lakes region (New York State, U.S.A.): *Zeitschr. für Geomorph.*, v. 9, pp. 50–62 (discussion, v. 10, pp. 475–477).
- COATES, D. R., 1966, Discussion of K. M. Clayton, "Glacial erosion in the Finger Lakes region (New York State, U.S.A.)": *Zeitschr. für Geomorph.*, v. 10, pp. 469–474.
- CRARY, A. P., 1966, Mechanism for fiord formation indicated by studies of an ice-covered inlet: *Geol. Soc. America Bull.*, v. 77, pp. 911–929.
- DALY, R. A., 1940, *Strength and structure of the earth*: Prentice-Hall, Inc., Englewood Cliffs, New Jersey, 434 pp.
- DERBYSHIRE, E., and OWEN, L. A., 1996, Glacioaeolian processes, sediments and landforms, in Menzies, J., ed., *Past glacial environments: Sediments, forms and techniques*: Butterworth-Heinemann Ltd., Oxford, UK, pp. 213–237.
- DOWDESWELL, J. A., and SCOURSE, J. D., eds., 1990, *Glacimarine environments: Processes and sediments*: Geol. Soc. [London], Spec. Pub. 53, 423 pp.
- DREIMANIS, A., 1988, Tills: Their genetic terminology and classification, in Goldthwait, R. P., and Matsch, C. L., eds., *Genetic classification of glacial deposits*: A. A. Balkema, Rotterdam, pp. 17–83.
- DREWRY, D., 1986, *Glacial geologic processes*: Edward Arnold (Publishers) Ltd., London, 276 pp.
- DYKE, A. S., VINCENT, J. S., and 3 others, 1989, The Laurentide ice sheet and an introduction to the Quaternary geology of the Canadian shield, in Fulton, R. J., ed., *Quaternary Geology of Canada and Greenland*: Geol. Soc. America, *The Geology of North America*, v. K-1, pp. 178–189.
- von ENGELN, O. D., 1921, The Tully glacial series: N.Y. State Museum Bull. 227-228, pp. 39–62.
- EVENSON, E. B., and CLINCH, J. M., 1985, Debris transport mechanisms at active alpine glacier margins: Alaskan case studies, in Kujansuu, R., and Saarnisto, M., eds., *INQUA till symposium, Finland 1985*: Geol. Survey Finland, Spec. Paper 3, pp. 111–136.
- EVENSON, E. B., SCHLÜCHTER, C. H., and RABASSA, A., eds., 1983, *Tills and related deposits*: A.A. Balkema, Rotterdam, 454 pp.
- EYLES, N., ed., 1983, *Glacial geology: An introduction for engineers and earth scientists*: Pergamon Press, Inc., Elmsford, New York, 409 pp.
- FEININGER, T., 1971, Chemical weathering and glacial erosion of crystalline rocks and the origin of till: U.S. Geol. Survey Prof. Paper 750-C, pp. 65–81.
- FLINT, R. F., 1971, *Glacial and Quaternary geology*: John Wiley & Sons, Inc., New York, 892 pp.
- GJESSING, J., 1966, On "plastic scouring" and "subglacial erosion": *Norsk Geografisk Tidsskrift*, v. 20, pp. 1–37.
- GOLDTHWAIT, R. P., and MATSCH, C. L., eds., 1988, *Genetic classification of glacial deposits*: A. A. Balkema, Rotterdam, 294 pp.
- GRAVENOR, C. P., 1975, Erosion by continental ice sheets: *Am. Jour. Sci.*, v. 275, pp. 594–604.
- GURNELL, A. M., and CLARK, M. J., eds., 1987, *Glacio-fluvial sediment transfer: An alpine perspective*: John Wiley & Sons Ltd., Chichester, UK, 524 pp.
- GUSTAVSON, T. C., and BOOTHROYD, J. C., 1987, A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide Ice Sheet: *Geol. Soc. America*, v. 99, pp. 187–200.
- GUTENBERG, B., BUWALDA, J. P., and SHARP, R. P., 1956, Seismic explorations on the floor of Yosemite Valley, California: *Geol. Soc. America Bull.*, v. 67, pp. 1051–1078.
- HALLET, B., 1976, Deposits formed by subglacial precipitation of CaCO₃: *Geol. Soc. America Bull.*, v. 87, pp. 1003–1015.
- , 1979, A theoretical model of glacial abrasion: *Jour. Glaciology*, v. 23, pp. 39–50.
- , 1981, Glacial abrasion and sliding: Their dependence on the debris concentration in basal ice: *Ann. Glaciology*, v. 2, pp. 23–28.
- , HUNTER, L., and BOGEN, J., 1996, Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications: *Global and Planet. Change*, v. 12, pp. 213–235.

- HARBOR, J. M., 1992, Numerical modeling of the development of U-shaped valleys by glacial erosion: *Geol. Soc. America Bull.*, v. 104, pp. 1364-1375.
- , 1995, Development of glacial-valley cross sections under conditions of spatially variable resistance to erosion: *Geomorphology* v. 14, pp. 99-107.
- , and Warburton, J., 1993, Relative rates of glacial and nonglacial erosion in alpine environments: *Arctic and Alpine Res.*, v. 25, pp. 1-7.
- HIRANO, M., and ANIYA, M., 1988, A rational explanation of cross-profile morphology for glacial valleys and of glacial valley development. *Earth Surface Processes and Landforms*, vol. 13, pp. 707-716.
- , 1989, A rational explanation of cross-profile morphology for glacial valleys and of glacial valley development: A further note: *Earth Surface Processes and Landforms*, vol. 14, pp. 173-174.
- JAHNS, R. H., 1943, Sheet structure in granites: Its origin and use as a measure of glacial erosion in New England: *Jour. Geology*, v. 51, pp. 71-98.
- KAKKURI, J., and CHEN, R., 1992, On horizontal crustal strain in Finland: *Bull. Géodésique*, v. 66, pp. 12-20.
- KASZYCKI, C. A., and SHILTS, W. W., 1987, Glacial erosion of the Canadian shield, in Graf, W. L., ed., *Geomorphic systems of North America: Geol. Soc. America, Centennial Special Vol. 2*, pp. 155-161.
- KAYE, C. A., 1967, Fossiliferous bauxite in glacial drift, Martha's Vineyard, Massachusetts: *Science*, v. 157, pp. 1035-1037.
- KOTEFF, C., and PESSL, F., JR., 1981, Systematic ice retreat in New England: *U.S. Geol. Survey Prof. Paper 1179*, 20 pp.
- LASALLE, P., DeKIMPE, C. R., and LAVERDIERE, M. R., 1985, Sub-till saprolites in southeastern Quebec and adjacent New England: Erosional, stratigraphic, and climatic significance, in Borns, H. W., Jr., LaSalle, P., and Thompson, W. B., eds., *Late Pleistocene history of northeastern New England and adjacent Quebec: Geol. Soc. America Spec. Paper 197*, pp. 13-20.
- LAWSON, D. E., 1993, Glaciohydrologic and glaciohydraulic effects on runoff and sediment yield in glaciated basins: *US Army Corps of Engineers Cold Regions Res. and Engin. Lab. Mono. 93-2*, 108 pp.
- LIEBLING, R. S., 1973, Clay minerals of the weathered bedrock underlying coastal New York: *Geol. Soc. America Bull.*, v. 84, pp. 1813-1816.
- LINDSTRÖM, E., 1988, Are roches moutonnées mainly preglacial forms?: *Geografiska Annaler*, v. 70a, pp. 323-331.
- LINTON, D. L., 1963, The forms of glacial erosion: *Inst. Brit. Geographers Trans.*, v. 33, pp. 1-28.
- MENZIES, J., ed., 1995, *Modern glacial environments: Processes, dynamics and sediments: Butterworth-Heinemann Ltd.*, Oxford, UK, 621 pp.
- , 1996, *Past glacial environments: Sediments, forms and techniques: Butterworth-Heinemann Ltd.*, Oxford, UK, 598 pp.
- MOLNIA, B. F., 1983, Subarctic glacial-marine sedimentation: A model, in Molnia, B.F., ed., *Glacial-marine sedimentation: Plenum Press, New York*, pp. 95-144.
- MOOERS, H. D., 1989, On the formation of the tunnel valleys of the Superior Lobe, central Minnesota: *Quaternary Res.*, v. 32, pp. 24-35.
- MULLER, E. H., 1974, Origins of drumlins, in Coates, D. R., ed., *Glacial geomorphology: State Univ. of New York Publications in Geomorphology, Binghamton, New York*, pp. 187-204.
- PENCK, A., and BRÜCKNER, E., 1909, *Die Alpen im Eiszeitalter: Tauchnitz, Leipzig*, 3 v., 1199 pp.
- PORTER, S. C., 1966, Pleistocene geology of Anaktuvuk Pass, Central Brooks Range: *Arctic Inst. of North America Tech. Paper 18*, 100 pp.
- PREST, V. K., 1968, Nomenclature of moraines and ice-flow features as applied to the glacial map of Canada: *Canada Geol. Survey Paper 67-57*, 32 pp.
- ROBIN, G. D., 1964, *Glaciology: Endeavour*, v. 23, pp. 102-107.
- SHARPE, D. R., and SHAW, J., 1989, Erosion of bedrock by subglacial meltwater, Cantley, Quebec: *Geol. Soc. America Bull.*, v. 101, pp. 1011-1020.
- SHAW, J., 1989, Drumlins, proglacial meltwater floods, and ocean responses: *Geology*, v. 17, pp. 853-856.
- SHOEMAKER, E. M., 1986, The formation of fjord thresholds: *Jour. Glaciology*, v. 32, pp. 65-71.
- SHREVE, R. L., 1985, Esker characteristics in terms of glacier physics, Katahdin esker system, Maine: *Geol. Soc. America Bull.*, v. 96, pp. 639-646.
- SMALLEY, I. J., 1981, Conjectures, hypotheses, and theories of drumlin formation: *Jour. Glaciology*, v. 27, pp. 503-505.
- , and UNWIN, D. J., 1968, Formation and shape of drumlins and their distribution and orientation in drumlin fields: *Jour. Glaciology*, v. 7, pp. 377-390.
- STREIFF-BECKER, R., 1947, Glacierization and glaciation: *Jour. Glaciology*, v. 1, pp. 63-65.
- TARR, R. S., 1905, Drainage features of central New York: *Geol. Soc. America Bull.*, v. 16, pp. 229-242.
- TELLER, J. T., 1987, Proglacial lakes and the southern limit of the Laurentide Ice Sheet, in Ruddiman, W. F., and Wright, H. E., JR., eds., *North America and adjacent oceans during the last deglaciation: Geol. Soc. America, The geology of North America*, v. K-3, pp. 39-69.
- TEN BRINK, N. W., 1974, Glacio-isostasy: New data from West Greenland and geophysical implications: *Geol. Soc. America Bull.*, v. 85, pp. 219-228.
- TUSHINGHAM, A. M., and PELTIER, W. R., 1989, Ice/3G: A new global model of late Pleistocene deglaciation based upon geophysical predictions of post-glacial relative sea level change: *Jour. Geophys. Res.*, v. 96, pp. 4497-4523.
- WARNKE, D. A., 1970, Glacial erosion, ice rafting, and glacial-marine sediments: Antarctica and the Southern Ocean: *Am. Jour. Sci.*, v. 269, pp. 276-294.

WASHBURN, A. L., and STUIVER, M., 1962, Radiocarbon-dated postglacial delevelling in northeast Greenland and its implications: *Arctic*, v. 15, pp. 66-73.

van der WATEREN, F. M., 1995, Processes of glaciotectonism, in Menzies, J., ed., *Modern glacial environments:*

Processes, dynamics and sediments: Butterworth-Heinemann Ltd., Oxford, UK, pp. 309-335.

WHITE, W. A., 1972, Deep erosion by continental ice sheets: *Geol. Soc. America Bull.*, v. 83, pp. 1037-1056.