GEOLOGIC PROCESSES AND HAZARDS IN HIGH-RELIEF TERRAIN

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INTRODUCTION

Research and study of present geologic processes and formations have traditionally been the domain of geologists. Actual processes like denudation, erosion and deposition, landslides, avalanches, and glacier motion in mountainous areas are of high technical and economic importance, especially with regard to settlement and transportation politics. For this reason, botanists, foresters, agronomists, engineers, and land use planners are not only concerned with but often actively participate in the investigation of geologic processes.

The rate with which these processes go on are not only determined by lithology and geomorphological conditions of an area but they fluctuate with the climatic cycles. Thus temperature and snowfall in high altitudes influence the number and severity of avalanches in a given season and determine advance and retreat of the glaciers. Scientists were able to trace about 20 warm-cold cycles for the last 2 million years, which means that one full glacial cycle lasted on the average 100,000 years. Upon the cycle, however, there are superimposed numerous minor climatic fluctuations of short duration. The last episode of significant glacial activity was the Little Ice Age which lasted from the fourteenth to the nineteenth century. During much of this time villages and hamlets in the European Alps were threatened and actually buried by the advancing glaciers.[1] A warming trend beginning in the middle of the nineteenth century brought happier days to the inhabitants of the Alps, but since about 1945 temperatures are reported to have been falling again so that glacier advance could again endanger human developments in the near future.

Scope and Purpose

The content of this paper summarizes the principles of glacial processes and some aspects of mass movement in mountain regions such as the Alps. Especially avalanche formation and protective measures as practiced in Switzerland shall be discussed. The purpose of this report is to review some surficial geologic processes in high-relief terrain and see how they affect human developments and activities. It is thus, too, a practical exercise in environmental geology, particularly as it relates to land use planning. Furthermore, it is a library research paper required for graduation.

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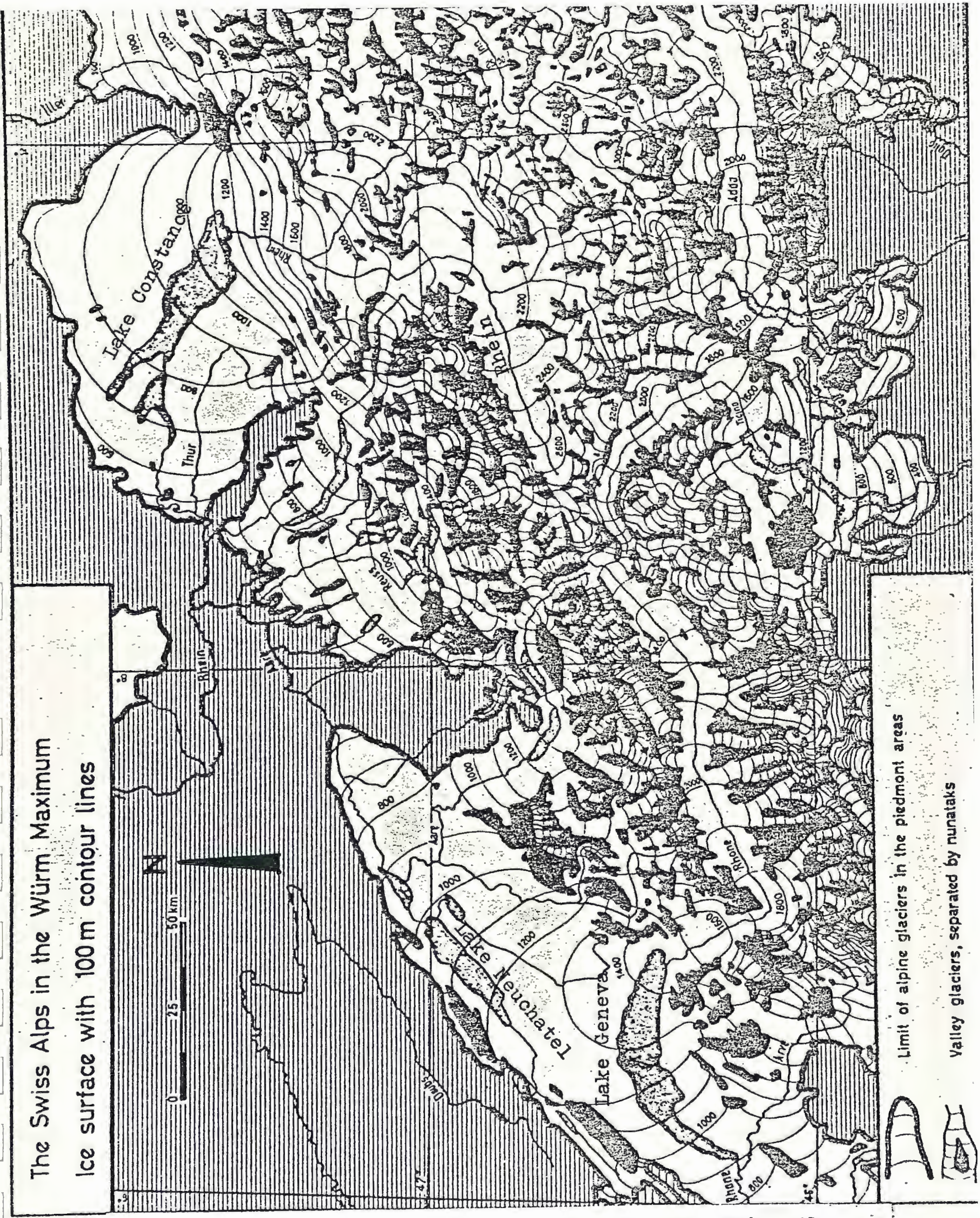
GLACIAL PROCESSES

Ice has exerted its most important geological action in the form of glaciers. At various times during earth history glaciation has been in operation over much wider areas than at present. The Scandinavian ice sheet, for instance, extended as far south as lat 48°, almost reaching the Alps.[2,p.5] At the same time, the glaciers of the Alps, an ice system independent from the Scandinavian ice sheet, stretched beyond the present borders of Switzerland, flowing into southern Germany.

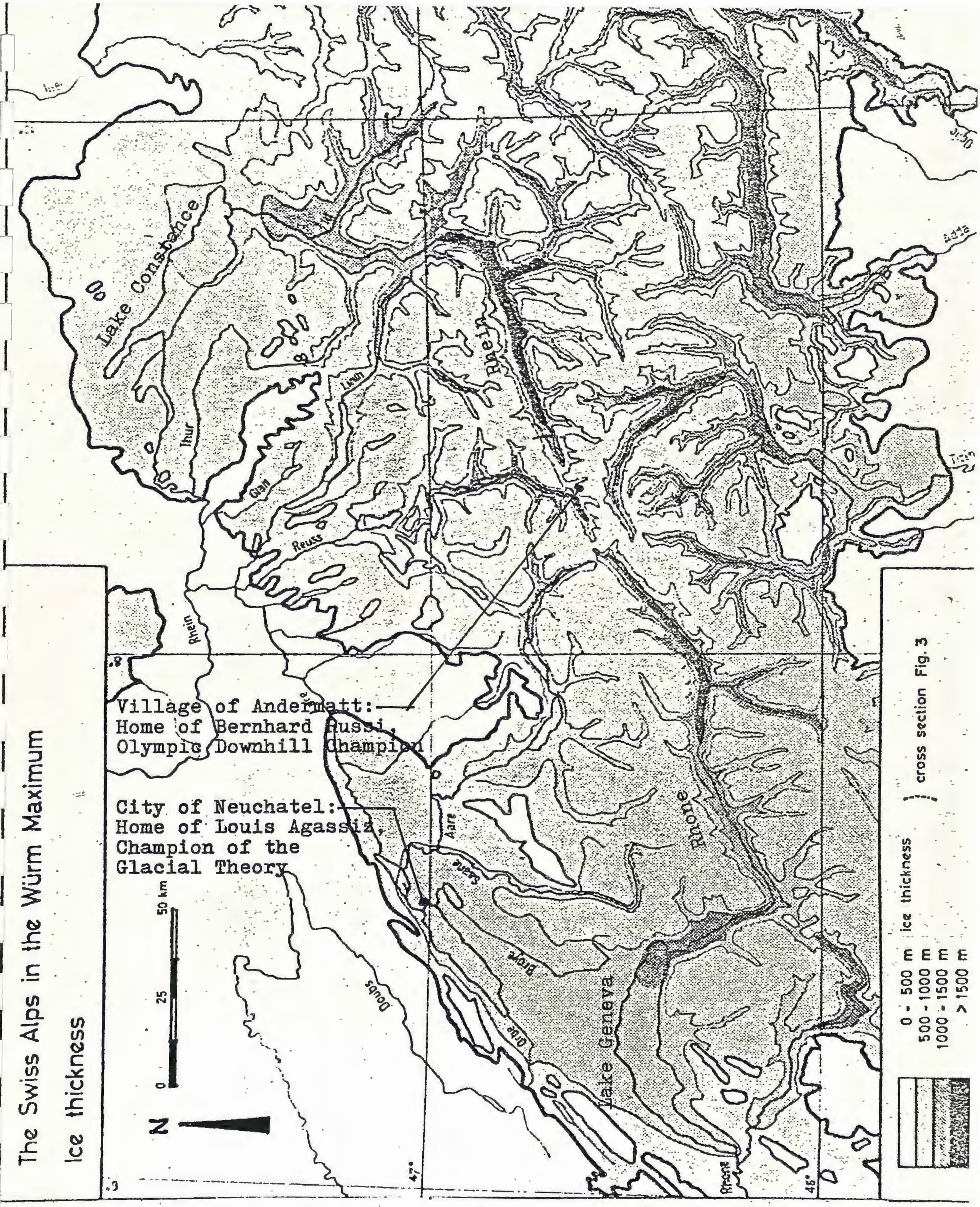
The positions and extent of the former ice sheets are known because the effects of thick ice spreading over the land have been observed in detail and carefully mapped. Repeated glaciations are indicated through superposed glacial deposits which are often separated by ancient soils.

Ice thickness during glacial ages varied enormously from one glacier to another. Large ice sheets are estimated to have reached a thickness of up to 3000 meters.[2] The valley glaciers of the European Alps, however, formed a network of ice with nunataks between the gently dipping ice streams. The Pleistocene glaciers of Switzerland reached their maximum thickness of 1500-1600 meters in the central-alpine valleys of the Rhone, the Rhine, the Ticino, and the Adda rivers (see Maps 1 and 2).

There is great difference between the topographic setting and landscape-carving activities of the generally discontinuous glaciers of mountainous regions and the continuous ice sheets which cover regions of small relief. Some of the processes and hazards of the mountain glaciers will now be discussed in the following sections.



Map 1. Contour map of the ice surface in the Swiss Alps during the last major glaciation (Würm). Contour lines at 100-meter intervals. After H. Jäckli, Reference 3, p. 155.



Map 2. Sketch map of ice thickness during the Würm Maximum in the Swiss Alps.
After H. Jäckli, Reference 3, p. 156.

GLACIER BEHAVIOR

Although a great deal of knowledge on climate and glaciers has accumulated, there is still much to be learned about the initiation of glacial periods, and indeed the problem of their causes has not yet been solved.[4,p.119] It is generally assumed that we are living in an interglacial period at present and the behavior of the world's great ice sheets and glaciers is of direct relevance to mankind. It is estimated that the melting of all present land ice would raise the sea level by some 50 meters and thus flood the major cities around the world.[5,p.3] Immediately significant for people living in the mountains are the fluctuations of the glaciers, particularly any signs of cooling and renewed growth of glaciers. Only a small lowering of the temperature at the snow-line would allow snow to remain throughout the summer and consolidate into small cirque glaciers. Present-day glaciers are carefully observed and studied; boreholes are drilled and tunnels dug in order to examine ice movement relative to the bed. The information gained from these investigations is used to explain ice flow characteristics and landforms that were created by similar processes in the past.

Climate and Ice Formation

Although climatic fluctuation and the formation of ice on the earth's surface are related, their causes are not necessarily the same.[2,p.10] Historically, fluctuations of climate have not always created extensive ice sheets, and the basic factors that initiated glaciations are to a great degree still matters of hypothesis and opinion. Nevertheless, observations on present Alpine glaciers show that advance and retreat of the ice masses respond to significant variations of temperature and precipitation. Thus the warming trend of the last 100 years (1850-1950, approx.) resulted in glacier retreat all over the world. This recent amelioration, however, came to an end around 1945 and the glacier margins in the Alps have since stabilized.

Regimen of Glaciers

Regimen is the material balance of a glacier involving the total accumulation and the gross wastage in one budget year. A positive regimen is the time when a glacier is gaining ice and will advance; in a negative regimen the glacier will retreat. As mentioned before, the regimen of a glacier is intimately related to climate and climatic changes. In terms of the hydrologic cycle, glaciers regulate or influence the water level of the oceans relative to the land. Since the amount of water substance at the earth's surface is nearly constant, the formation of glaciers and lakes on the land diminishes the water in the sea by a corresponding amount. During warm-up periods, on the other hand, water locked up in glaciers and the related lakes will be returned to the oceans. It has been calculated, on the basis of assumptions as to extent and thickness of former ice sheets, that the sea level during glacial ages might have been lowered by about 130 meters.[4,p.55] Today's situation though is less dramatic since we are now either still in an interglacial period or else only at the beginning of a new period of glaciation. In any event, the effects of seasonal as well as longer-term changes of climate give rise to variations in accumulation and ablation.

The snowline is a conspicuous feature seen at the end of a glacier's melting season. It separates the area of accumulation from that of ablation (see Fig 1). Accumulation measures the solid water substance added to the glacier, while ablation measures the portion lost from the glacier. In general, glaciers lose most of their substance by melting, although wind erosion, evaporation, and avalanches contribute significantly to loss of mass. In high mountains, direct solar radiation may account for more than 80 percent of the heat applied to ablation.[2,p.38] In general, variations in ablation affect the net mass budget of a glacier more than variations in accumulation. This is the reason why temperature, not precipitation, is the climatic factor primarily responsible for fluctuations observed in glaciers.[2]

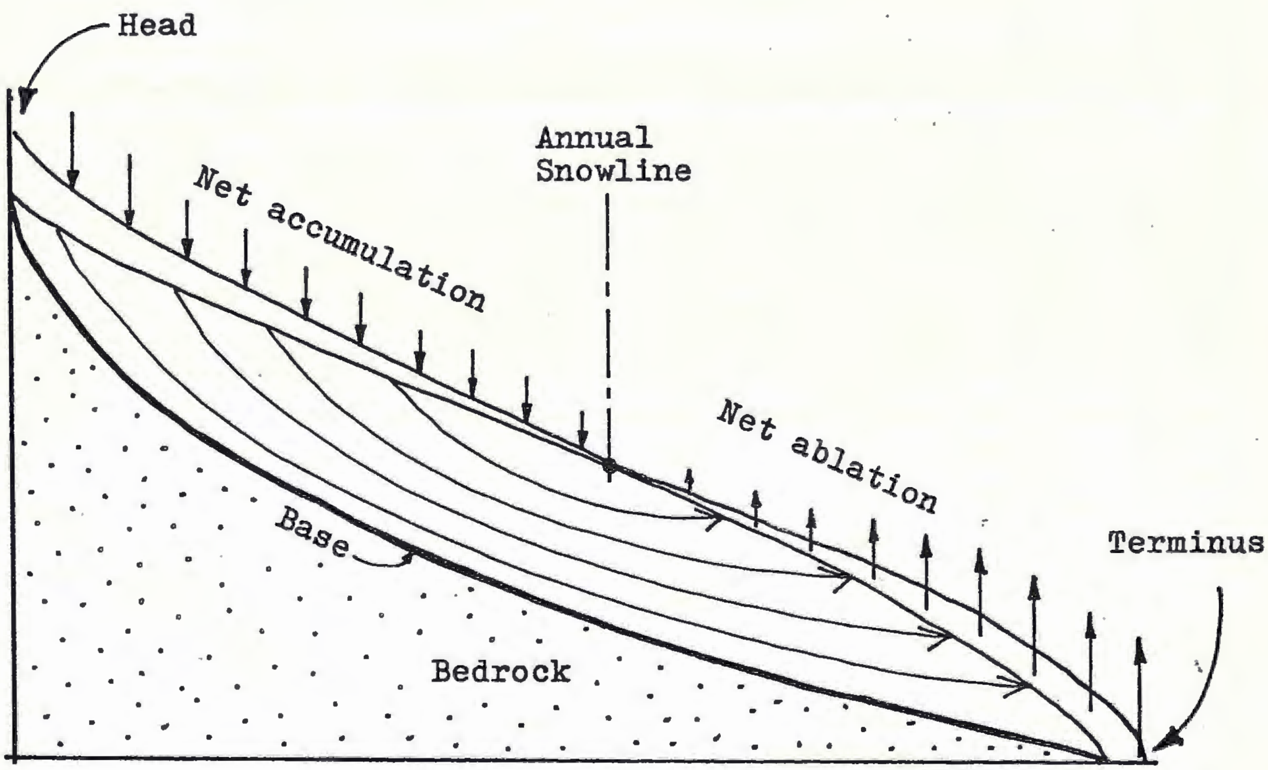


Figure 1. Long section of an ideal valley glacier.
From Reference 2, p. 36.

Types of Glaciers

There are three categories of glacier classification:

1) Thermal; which recognizes temperate glaciers with meltwater at their base, and cold glaciers which are essentially frozen to the ground.

2) Dynamic; distinguishing between active glaciers which are fed by a continuous ice stream from an accumulation zone, passive or stagnant glaciers with retarded ice movement, and dead glaciers in which movement has ceased.

3) Morphological; here, classification is based essentially on size and the characteristics of glacier environment. The most important types of morphologically distinct ice masses are the following:

Cirque glaciers are small masses of ice confined to cup-shaped rock basins in mountainous terrain. This type of glacier must be the most abundant today, for hundreds of cirque glaciers exist in the United States, in the mountains of Norway, and in the Alps.[2,p.28] Map 3 shows several cirque glaciers and the remnants of valley glaciers some of which have their source themselves in a cirque.

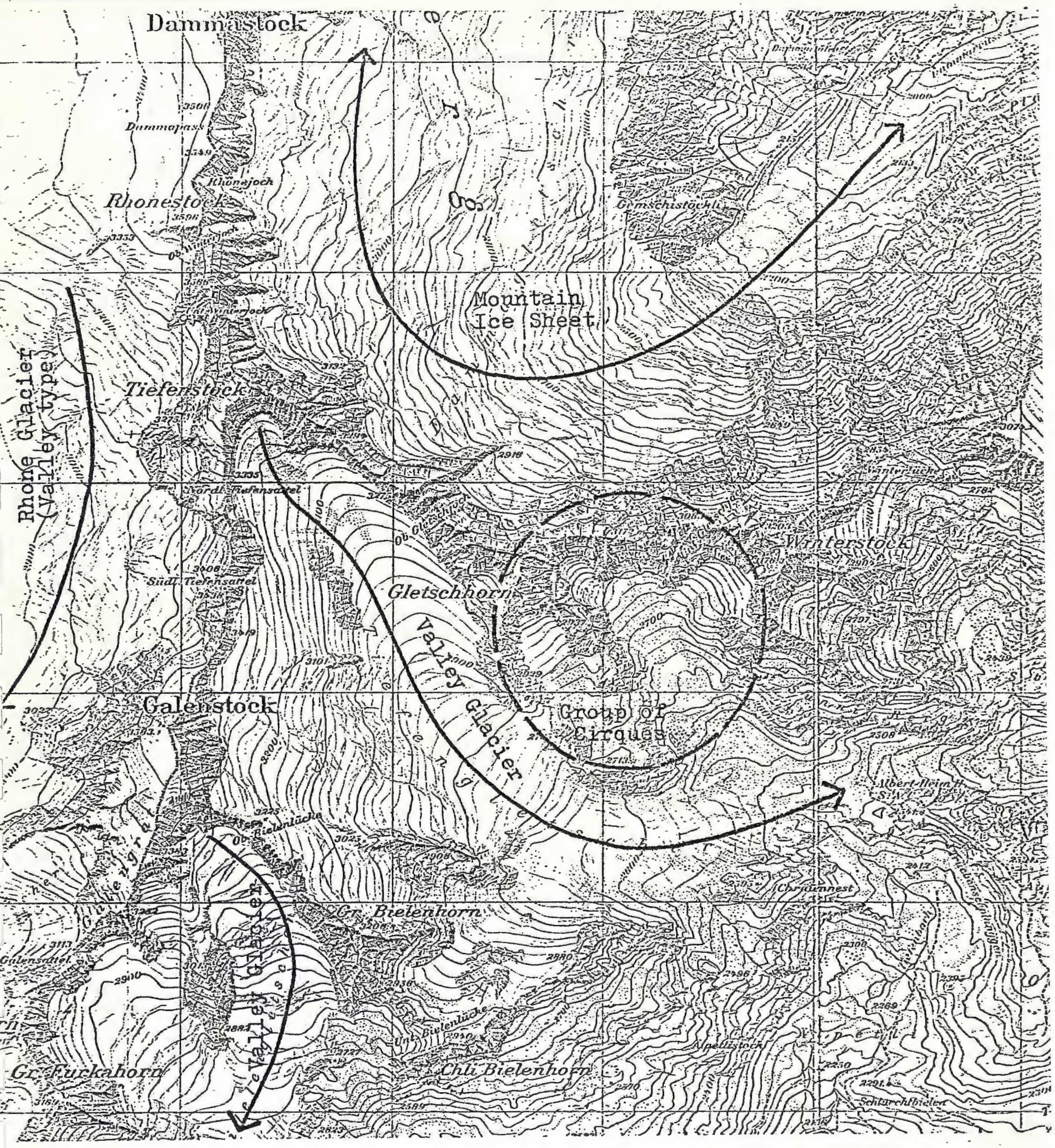
Valley glaciers flow down a valley bounded by exposed rock. Generally their width is small in proportion to their length and many of them have branching tributaries (see Fig 2 at the end of this section). They are also known as mountain or Alpine glaciers. Many of them originate in cirques or groups of cirques (see Map 3); others come from snow-fields (névés) or ice sheets and ice caps.(6,p.278) The valley glaciers mold themselves to the shape of the underlying ground and usually follow pre-existing valleys.(2)

<u>Piedmont glaciers</u> are large aprons of ice formed at the base of mountain ranges where several valley glaciers merge to coallesce into a continuous spread-out mass. Thus the piedmont glaciers occupy broad lowlands and are a rare phenomenon today. Examples of former piedmont glaciers in the European Alps are shown on Map 1.

<u>Ice caps</u> are small ice sheets overlying plateaulike uplands in highland regions. Where their margins encounter well-defined valleys, they give origin to outlet or valley glaciers.

<u>Mountain ice sheets</u> are glaciers that overly or originate in two or more mountain masses exposing high crests and peaks. A qualifying condition is that outflow takes place in two or more directions. Therefore, the Damma glacier in Switzerland shown on Map 3 might be classified as a remnant of a former mountain ice sheet.

<u>Continental ice sheets</u> are enormous masses of ice which cover major parts of a continent uninterrupted, allowing only the highest mountains to penetrate. Ice flow occurs



Map 3. Glacier types in the Urseren area of the Swiss Alps.
From Blatt 1231 of Landeskarte der Schweiz.
Scale: 1 : 25,000

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outward in all directions from one or more central areas. Present-day examples are the ice sheets of Antarctica and Greenland.

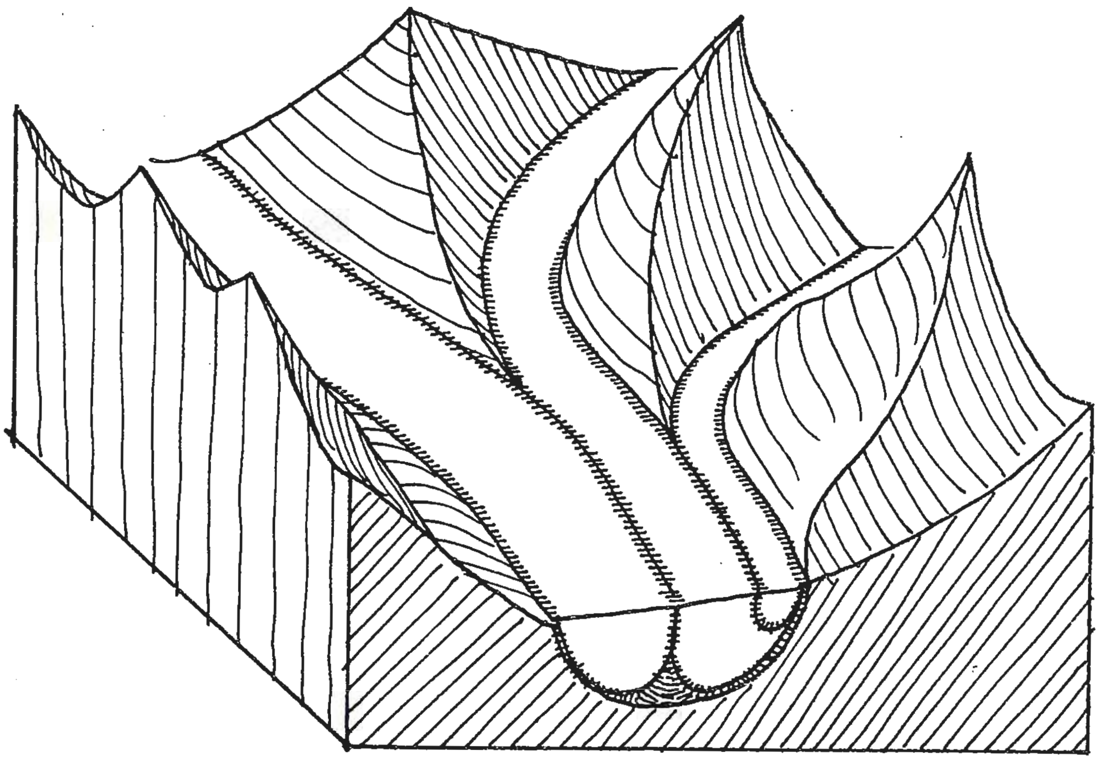


Figure 2. Idealized complex valley glacier, showing that tributary bodies maintain their identities after joining the main glacier.

Adapted from Reference 2, p. 35.

Ice Motion

Because of low bond strength between its molecules, ice is a very weak crystalline solid and flows therefore easily under the gravitational forces acting on its mass. For all practical purposes, there are two kinds of glacier movement recognized, namely sliding over bedrock, and internal deformation or plastic flow. The relative contribution that each makes toward total ice motion varies with bed slope, ice thickness, and ice temperature.

Polar glaciers (cold glaciers) are generally believed to be frozen to the ground so that the velocity component for basal sliding would virtually be zero. Furthermore, a polar glacier exhibits lower internal deformation since cold ice can support greater shear stresses than ice at pressure melt-point.[5,p.126] Temperate glaciers, on the other hand, are approximately at pressure melting-point throughout their thickness (except for upper 15 to 20 m in winter) and have therefore not only a higher internal deformation rate but show also a great deal of basal and side slip. However, the mechanism of basal sliding is not well understood yet, but it is of great geomorphological interest since it could explain the more obscure facets of glacial erosion.[5,p.110]

Over longer periods, the most rapidly moving glaciers are those descending steep slopes and fed by large accumulation areas. However, flow rates vary widely not only among different glaciers but also in different parts of an individual glacier. Average velocities of up to 200 meters per year are not uncommon for valley and outlet glaciers.(4,p.21) Exceptional velocities of several kilometers per year have been observed, but such phenomena are known as glacial surges and will be discussed later.

The distribution patterns of movement in glaciers are related to the morphology of the ice mass and the location of the equilibrium line between the zones of accumulation and ablation (see Fig 3). Movement of a glacier is explained by the physics of stress (force per unit area) and strain (deformation of shape or volume), whereby stress effects strain and strain rates govern velocity. Basal shear stress is given by the equation

$$\tau_b = \rho g h \sin \alpha$$

where τ_b is the shear stress at the base of the glacier.
ρ is the density of the ice.
g is the acceleration of gravity.
h is the thickness of the glacier.
α is the slope of the upper surface.

The above equation shows that shear stress, at a given slope, increases directly with thickness of the glacier, and the value of τ_b can be found when ice thickness and angle of slope are known. According to the theory of perfect plasticity, the yield stress τ_o has a value of about 1 bar. Going back to the above equation, it becomes evident that

$$\frac{\tau_o}{\rho g} = h \sin \alpha = \text{constant},$$

which means that any change in the thickness of the glacier must be compensated by a change in slope. This is an explanation of the natural fact that glaciers become thinner on steeper and thicker on gentler slopes.[5,p.125] At the same time, consistent with the theories of fluid mechanics, ice in thinner stretches flows faster than in thicker parts of a glacier. Figure 4 shows the mutual relationship between velocity and shear stress distribution in glacier ice.

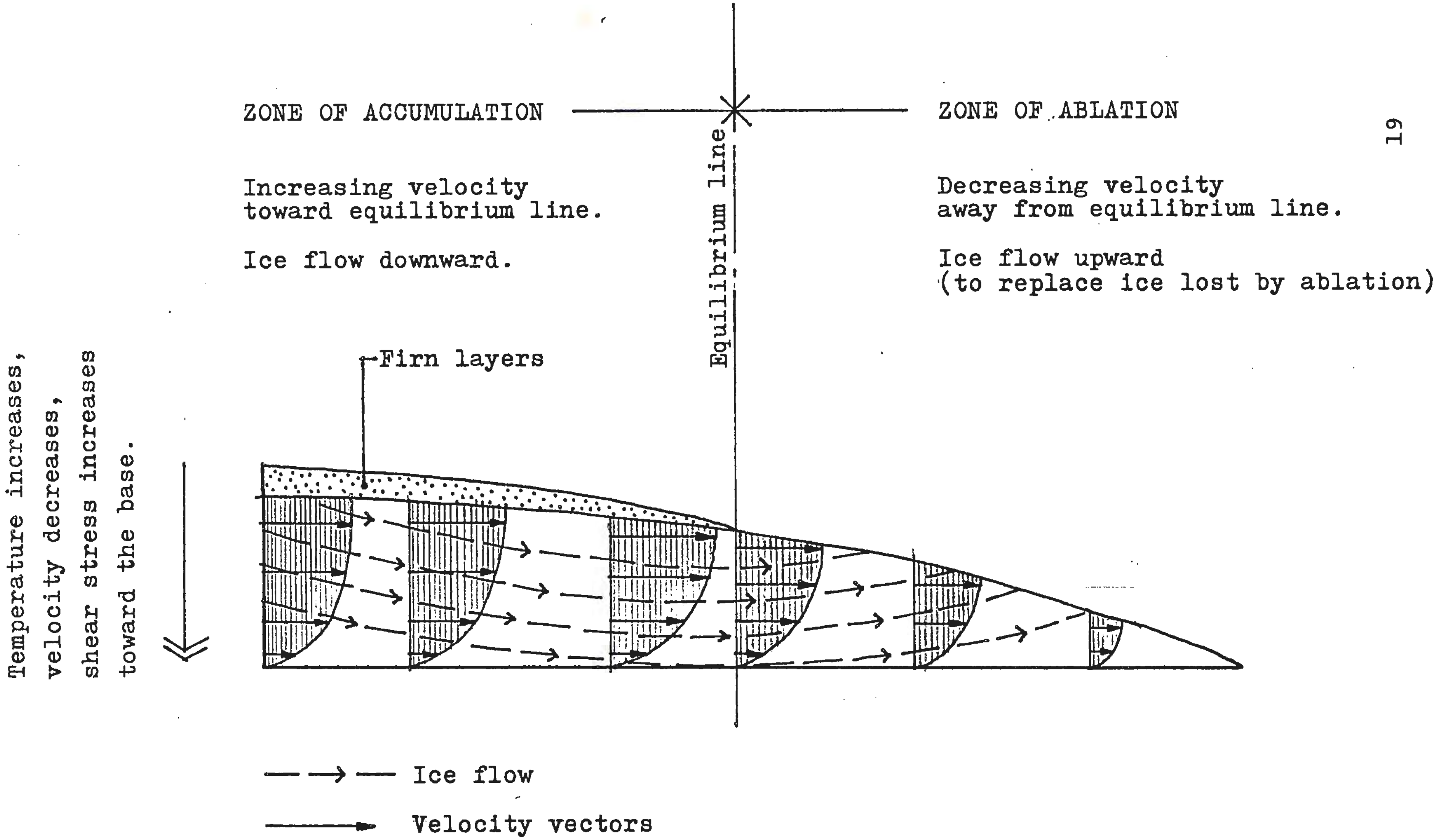
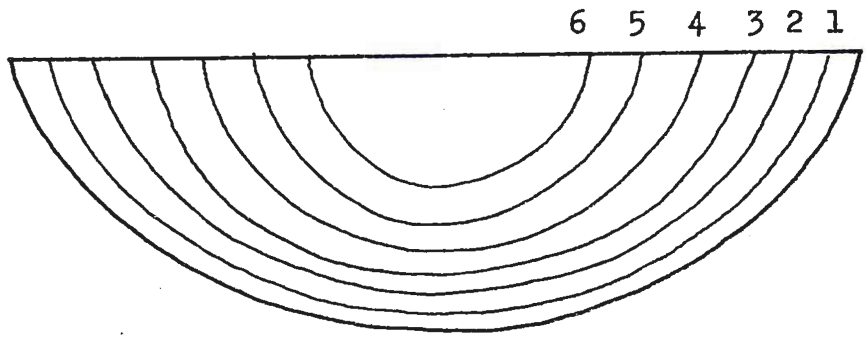
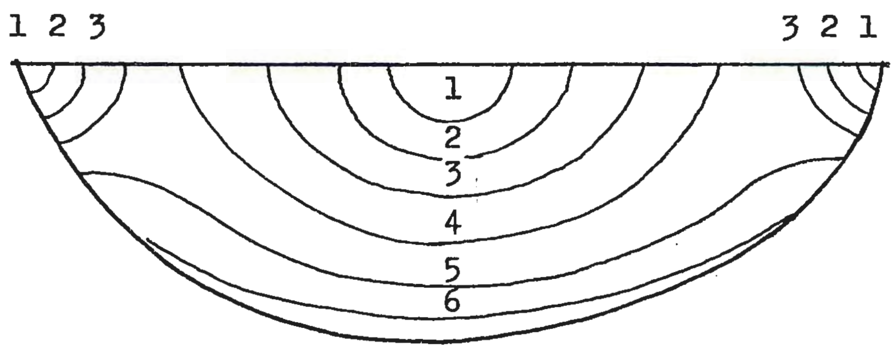


Figure 3. Schematic long section of ideallized glacier showing ice flow and velocity profiles.



(A)



(B)

Figure 4. Theoretical velocity distribution (A) and shear stress distribution (B). Magnitudes are dimensionless and have only qualitative meaning.

Adapted from Reference 5, p. 129.

EROSION AND TRANSPORT

The former glaciation of a given area is established by the evidence of glacial activity such as scratched and polished bedrock, rocks foreign to the localities of occurrence, and unsorted and unstratified deposits of rock fragments of all sizes. As the ice melts and flows away as meltwater, however, it carries sediments beyond the margin of the glacier, leaving a cover of sorted deposit called outwash. The seasonal variations in water supply and the abundance of sediment leads to the development of braided (multichanneled) rivers. In times of dry weather, the fine sediments of the outwash plains can be blown away by winds and be redeposited as sand dunes or blankets of loess (wind-blown silt).

The heavy glacial drift or rock fragments in a glacier are carried at the ice-bedrock interface. These rock inclusions, known as the load of the glacier, are the chief agents of quarrying and abrasion. Some of the drift material is avalanched from steep valley sides and thus transported on the upper surface of the glaciers. But the bulk of the drift is carried at the base of the glacier and is released from the ice by ablation at the terminus. These drift accumulations, known as moraines, have created distinct topographies, characteristic for glaciated areas.

Markings and Ice-Flow Forms

Glaciated regions often display rock surfaces that show clear signs of the former passage of ice over them. The most familiar of these features are polished rock surfaces, striations, channels, grooves, and gouges suggesting the extraction of small chips of rock by some agency.

Striations are fine scratches made by rock particles held in the base of a glacier, whereby the finer scratches might have been made by sand and silt, and with decreasing particle size the scratches grade into a general polish.[2,p.88] The smoothness and brilliance of the polish is determined by the fineness of the abrading grains and the textural and mineralogic properties of the abraded rock. Due to a wide range of grain size in the fragments carried by a glacier, polished rock surfaces also bear larger scratches. These, however, are usually found on harder and coarser grained rocks such as granites and volcanics.

Drift particles of larger diameter seem to carve grooves into the bedrock. Generally found in soft rock such as carbonates, giant grooves of 12 km length, 100 m width, and 30 m depth have been located.[5,p.183] It is assumed that these larger grooves have not been gouged out by single boulders, but rather groups of boulder packed and frozen together. Another assumption is that they are merely ice-modified pre-glacial grooves, or that they are the work of boulder-transporting subglacial streams.[5] In combination with other directional indicators, grooves and striations have been used successfully to reconstruct the ice-flow patterns of former glaciations. However, striations have been found to be made also by snow avalanches, rock avalanches, and debris flow. Therefore, before striations are interpreted as glacial, reasonable evidence for it must be established.[2]

Various other marks of small size are found on glaciated bedrock surfaces. The most common ones are crescent-shaped small-scale features such as crescentic gouges and lunate fractures caused by percussion and sliding of larger rock fragments.[4] Other features of the same category are crescentic fractures, Sichelwannen, conchoidal fractures, and chattermarks. All of these bedrock markings are collectively called friction cracks.[5,p.187] Some of them are shown in Fig. 5 below.

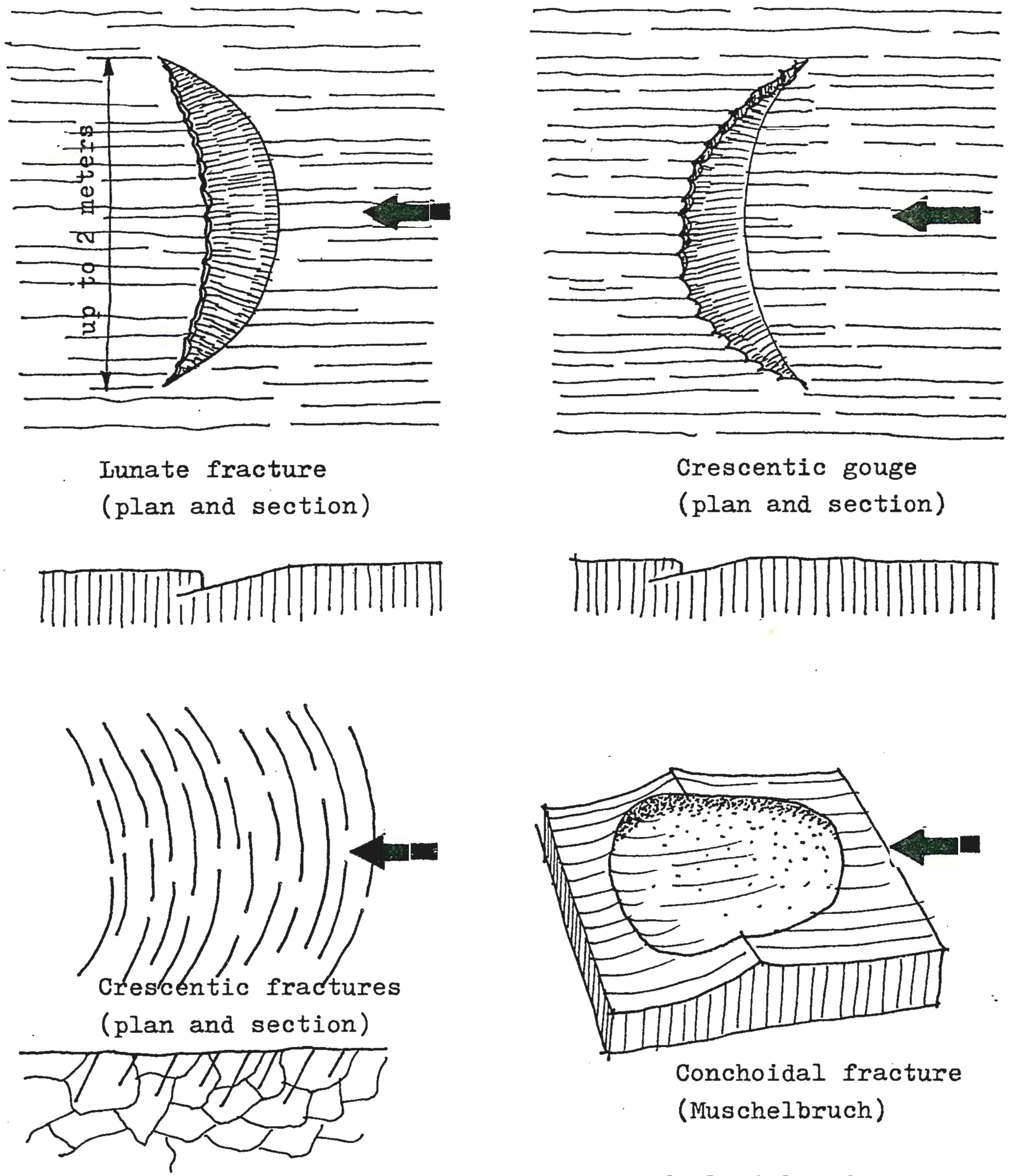


Figure 5. Some small-scale features of glacial and fluvioglacial erosion. Horizontal arrow indicates direction of ice motion.
From Reference 5, p. 190.

Another feature of glacial erosion are large bedrock outcrops which have been abrased on their upstream side and plucked or quarried on the lee side. These forms are tapered and have a cylindrical shape, and their size varies from a few meters to several hundred meters. They can serve as accurate indicators of ice flow since the smoothed and streamlined stoss side is always pointing upstream. Known as whalebacks or roches moutonnées, they usually occur in groups dominating considerable areas. The asymmetric arrangement of such bedrock bosses and small hills in a strongly glaciated district has been termed stoss-and-lee topography.[2]

Cirques

The cirque is one of the most characteristic forms of glacial erosion in high-relief areas. It consists of a rounded basin partially enclosed by steep cliffs and sometimes containing a small lake or cirque glacier. Cirques seem to be formed by erosive processes associated with the presence of ice in them. Like other landforms, cirques vary in size tremendously. They can be small and shallow depressions a few tens of meters across, or they may have widths of several kilometers and backwall heights of hundred or even thousands of meters.

Form and size to which a cirque will grow depends on many factors: proximity to other cirques, homogeneity and structure of rock material, competence of the rock to withstand failure, and surely climatic conditions and duration of glaciation. Depending upon these factors, the length-height ratio of a cirque may have a wide range, but a typical median is about 4:1 or 3:1. Figure 6 below shows the long profile of an idealized cirque.

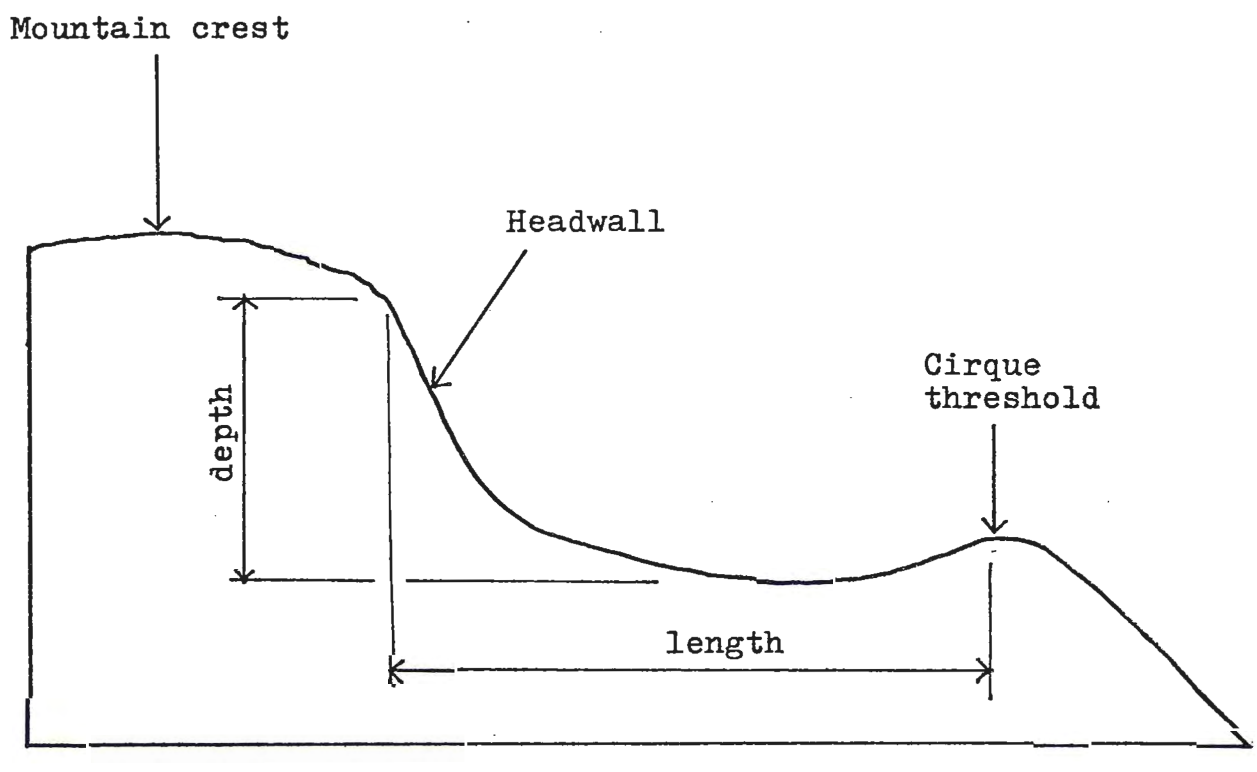


Figure 6. Cirque morphometry; longitudinal profile of a typical alpine cirque.
From Reference 5, p. 209.

Although rock structure may influence shape and development, many cirques seem remarkably little affected by geological structure, and the outline may cut across well-marked lithological boundaries with almost no change.(5,p.211) As cirquess grow, mountain masses are reduced and, indeed, cirque recession in many areas is responsible for the destruction of uplands. Receding cirques create a series of arêtes and horns like those striking features in the Swiss Alps (see Map 3). As the processes of glacial erosion, rock avalanching, and fluvioglacial transport persist, the crests (arêtes or Gräte) will disintegrate until the horns are the lonely survivors in this continuous drama of geomorphologic transformation. The Matterhorn in Switzerland, for instance, is a world-famous example of that kind of monumental landform evolution.

Valleys

Another spectacular result of glaciation are the mountain valleys, those great glacial troughs carved out to depths of hundreds or even thousands of meters. Most, if not all, of these valleys were prepared by pre-glacial rivers, and the glaciers that later flowed through them originated in various ways. As the snowline fell with increasing cold, many of these valley glaciers must have originated in cirques, while others establihsed themselves as mountain ice caps on upland plateau areas from which they formed outlet glaciers into the valleys below (see Map 3).

Glaciated valleys develop characteristic transverse and longitudinal profiles. In general the glaciated valley has a steplike long profile. The steps or Riegels are highest and steepest in the headward part, decreasing progressively down the valley. Between the steps there are usually unabraded rock basins which frequently contain lakes or are infilled with alluvium. Other features associated with glaciated valley long profiles are the hanging tributaries that occur along the valley sides, believed to be the result of differential erosion rates. Figure 7 shows an idealized long profile of a glaciated valley.

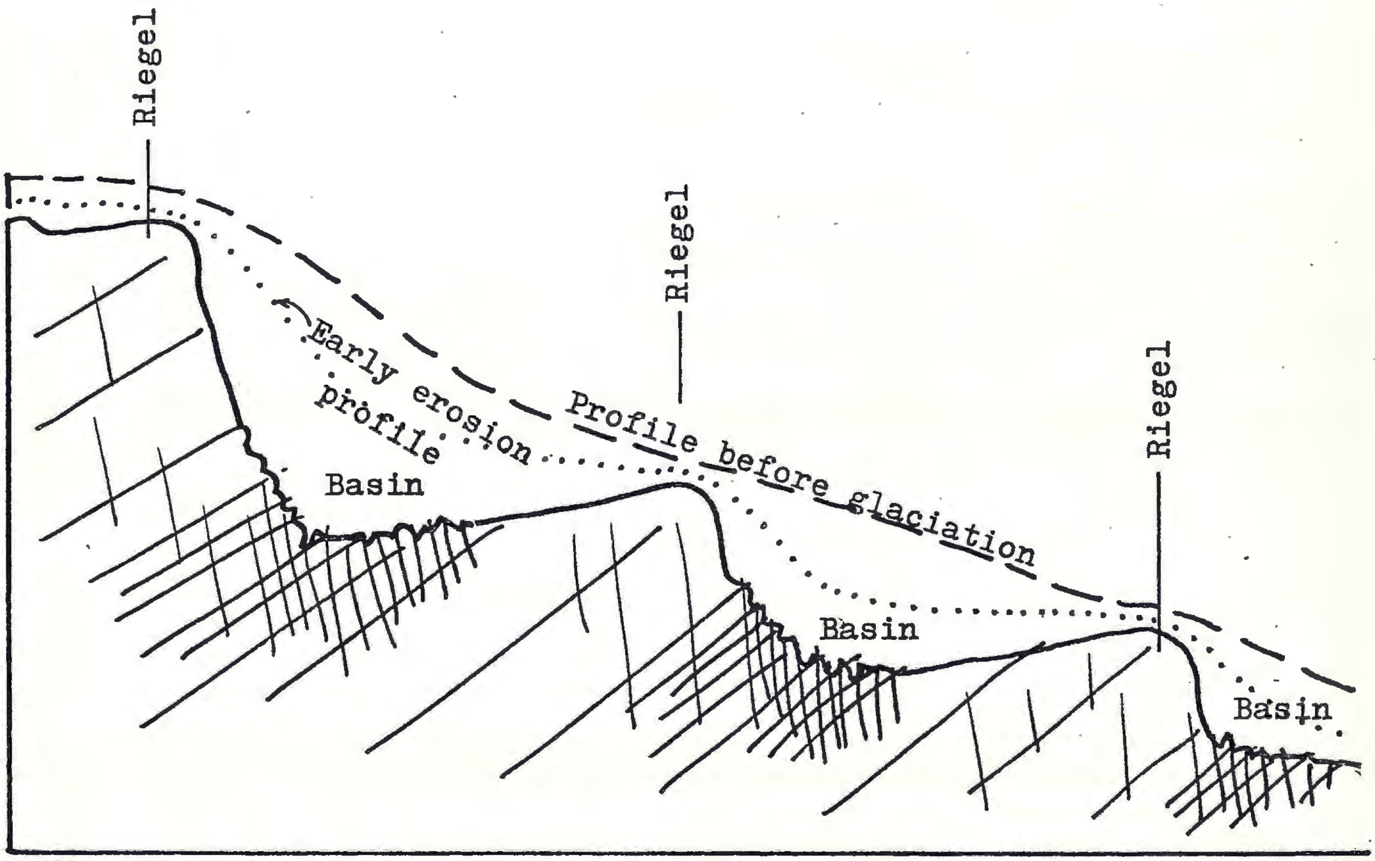


Figure 7. Long section of part of a glaciated valley, showing riegels and basins. The steps are shown to be controlled by unequal distribution of joints and other bedrock weaknesses.

Adapted from Reference 2, p. 128.

The cross-profile of a glaciated valley, in contrast to a river-cut valley, is typically U-shaped. The side walls are usually steep, in places approaching the vertical, whereas the floors are flatter than in river valleys. Deepening of the valley may exceed widening or vice versa, but the general cross-profile in most cases is U-shaped or parabolic. In the Swiss Alps, some glaciated valleys have composite cross-profiles, in that a narrow U is set within a higher and broader U, unrelated to lithology. These features are called two-story valleys which are believed to represent pre-glacial stream valleys later modified by glaciation.[2,p.130]

Meltwater Channels

Temperate glaciers are able to support extensive and deeply penetrating systems of englacial and subglacial drainage. Surface water (supraglacial water) produced by melting of snow and ice on the surface of the glacier is by far the greatest source of supply to the englacial and subglacial drainage systems.[5,p.327] The permeability of temperate ice is mainly due to fractures and tunnels which eventually lead to the bottom of the glacier. This water is the chief factor for basal sliding of the temperate glaciers and thus makes possible the cutting and plucking of bedrock. Large subglacial streams issuing with great force from caves at the snouts of glaciers are regarded as potent agents of erosion and transport. Subglacial streams can attain high velocities and discharges loaded with rock particles washed out from basal moraine with which they may be capable to effect active down-cutting even in hard bedrock. A good example of that kind of down-cutting is the Lütschine gorge in Grindelwald, Switzerland.

DEPOSITIONAL LANDFORMS

Glaciers and their meltwaters deposit a wide range of materials in many different forms. Some regions, like the Swiss lowland (Mittelland), are dominated by depositional features while others (e.g. the high Alps) have a striking erosional character. However, the geomorphological image of many glaciated highland areas is marked by both erosion and deposition, to a higher or lesser degree.

In the foothills of the Swiss Alps, for instance, glacial rivers have cut deep into the relatively soft bedrock (Molassefels) and deposited the gravels of the different glaciations.[7,p.136] The oldest gravel loads were laid down when the valleys still had shallow depths and, consequently, they covered extensive areas. For this reason these early deposits are called cover gravel (Deckenschotter). Renewed advances during later glaciations resulted in new rivers which further deepened the valleys and deposited more gravel on a lower level. The Riss glaciation was the most severe of the pleistocene glaciations in the European Alps (as was the Illinoian glaciation in North America) and had two episodes of great advance. As a result, the valleys were twice eroded and carved deeper and subsequently filled with gravel.[7] The last ice age, the Würm glaciation, had smaller glaciers and thus much less meltwater and gravel than during the Riss episode. Therefore, the glacio-fluviatile activities of the Würm glaciation were not capable to completely erode the Riss deposits and cut deeper into the bedrock (see Fig. 8).

Table 1

MAJOR PLEISTOCENE GLACIATIONS

North America	European Alps
Wisconsin Glaciation	Würm - Eiszeit*
Interglaciation	Zwischeneiszeit**
Illinoian Glaciation	Riss - Eiszeit (I + II)
Interglaciation	Zwischeneiszeit
Kansan Glaciation	Mindel - Eiszeit
Interglaciation	Zwischeneiszeit
Nebraskan Glaciation	Günz - Eiszeit

*Eiszeit is the German word for Glaciation.

**Zwischeneiszeit is the German word for Interglaciation.

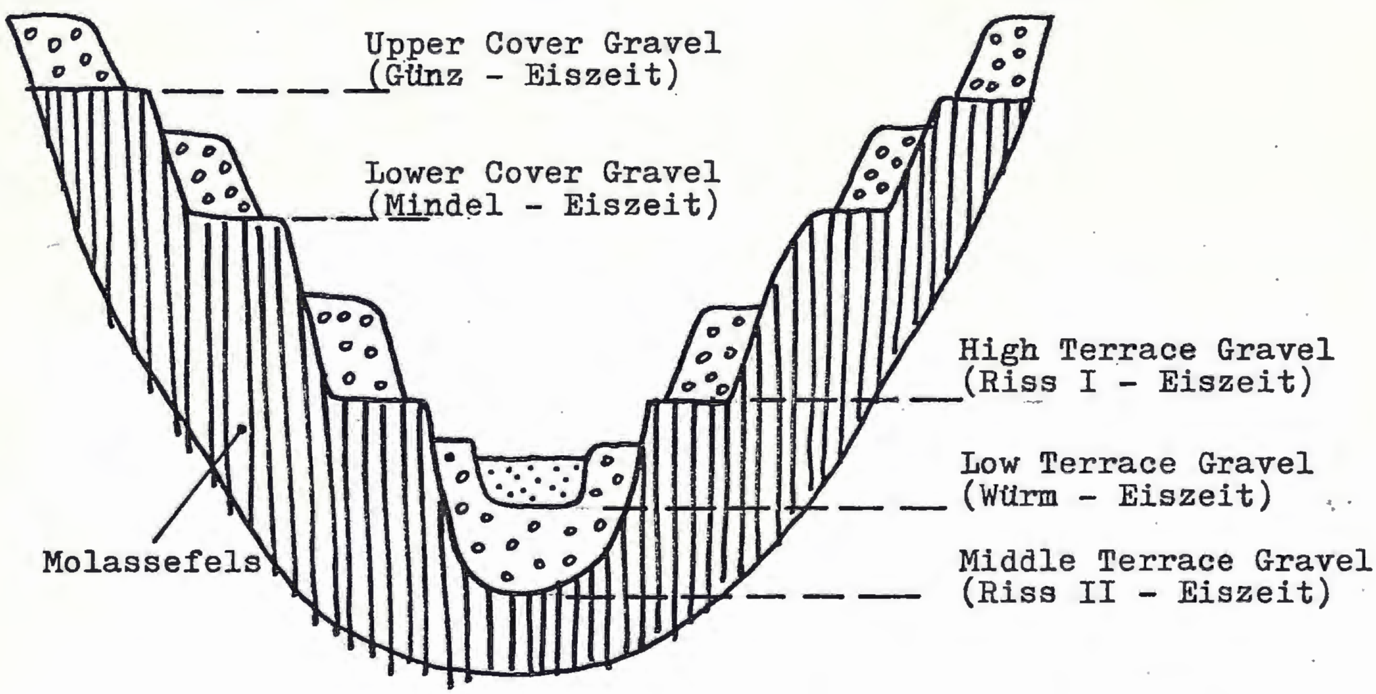


Figure 8. Schematic profile through a valley in eastern Switzerland, showing the gravel deposits of the different Pleistocene glaciations in the Alps.

Adapted from Reference 7, p. 136.

Some of the more classical landforms associated with glacial and fluvioglacial deposition will briefly be discussed in the following paragraphs.

Moraines and Drumlins

The most important deposition features of the Ice Age are the moraines. On their way from the high alpine mountains to the lowland the glaciers carried tremendous amounts of boulders and gravel with them. Part of this material fell onto the surface of the glacier from rockfalls and avalanches, and in part it was abrasion material transported at the base of the glacier. Some of the huge boulders, especially those carried on the surface of the glacier, were transported tens or hundreds of kilometers onto lower altitudes where they got laid down as so-called erratics. Most of the material, however, was broken and grinded down at the base of the glaciers during the long descent into the lowlands. Some of the structure characteristics within a moraine are the following:[7]

1) The rock components are not sorted as to size; their diameters range from boulders of several meters to clay particles.

2) Particle shapes of the components are not uniform. Individual gravel pieces may be scratched at their surface; some may even be polished.

3) Lack of uniformity with regard to lithology. However, if abundant gravel and boulders can be associated with a specific source rock, they may serve as indicators (Leitgesteine).

4) Moraines show no stratified structure. However, there is a distinction to be made between lodgement till and ablation till within a moraine. The former is thought of being deposited under pressure, from the base of the glacier. It is therefore a very compacted till in which stones tend to orient their long axes parallel with the direction of flow. The ablation till, on the other hand, is deposited from drift upon or within the terminal area of a shrinking glacier. It is thus dumped onto the ground, is loose, and shows less orientation. In addition, fines are washed away selectively.(2,p.171)

Terminal moraines are the most obvious morphological features created by glaciers in lowlands. They are long hummocky ridges, or belts of hills, produced by prolonged deposition at the end of glaciers. At present there are few valley glaciers with well-defined terminal moraines because most glaciers are retreating, and terminal morraines can only form effectively when the glacier is advancing or stationary under active ice conditions.(5,p.435) For this reason most upland valleys in Switzerland do not have impressive end moraines; but they are found abundantly in the lowlands. End moraines (terminal moraines) stand at ninety degrees to the ice flow direction or they have a crescentic configuration, always marking the position of the former glacier fronts.

Lateral moraines are formed along the flanks of glaciers and consist of surface debris derived from the rock walls of the valleys. In some parts of the Alps, lateral moraines form well defined ridges and thereby record the lateral extension of the glaciers. Younger glaciations usually have altered, covered, or erased the moraines of former ice ages so that in Switzerland, for instance, some of the best examples of both end and lateral moraines are those worked up by the glaciers of the Würm-Eiszeit.(7,p.134)

<u>Ground moraines</u> were (and are) created at the sole of the glaciers. They are not thrown into bold relief, but are broadly distributed masses of gravel which follow the surface of the bedrock. They have filled irregularities in the ground such as troughs, grooves, and channels, and they appear as undulating plains with low relief devoid of transverse linear elements.[2,p.199] Ground moraines are primarily the result of the abrasive action taking place at the sole of the glaciers. Thus the material of these moraines, speaking of mountain glaciers, represents a mixture of alpine rubble (Geröll) and abraded bedrock.[7]

With progressive transport the rock materials at the sole of the glaciers got more and more ground up so that the moraines are rich in fines such as silt and clay. Frequently the clays occur as banded varves (Warventon),[7] probably as a result of seasonal meltwater fluctuations. Furthermore, there are abundant polished rocks and scratched push gravel (Geschiebe), and rounded particle shapes are much more common than in lateral or terminal moraines. All of these factors together with the enormous weight of the ice mass compacted the ground moraines into a deposit of very low permeability; actually in many places it is reported to be impermeable.[7,p.135]

Ground moraines in Switzerland occur primarily in the bottom of valleys where they reach thicknesses of 10-30 meters; in some places, however, they were measured to be thicker than 100 meters.[7] They are also found in the plains of the lowland (Mittelland). In the region of Lake Constance, for instance, the moraine has thicknesses between 100 and 150 meters.[7] Another extensive gound moraine landscape spans from Lake Geneva to the Jura Mountains (see Maps 1 and 2).

Drumlins (Rundhügel) are another common landform associated with glacial deposition. They are streamlined hills, 30 to 100 meters high (as measured in Switzerland), have an elongated shape with a steep slope facing in the upglacier direction and stand parallel to glacier flow. Drumlins usually occur in large groups arranged in long stretches. They are formed mainly of till similar to that found in ground moraines, but sometimes they contain stratified material or a core of rock and gravel. The mode of formation of these drumlins is not known with certainty, but it is assumed that such hills could have formed along radial rifts at the margin of glaciers where ice thickness was low. Most likely glacial till could accumulate along such longitudinal crevasses.[7,p.135]

Eskers, Kames, and Kettles

Some deposition features are distinctly different from end moraines although they seem to be related to stagnant ice in the terminal zone of a glacier. The most easily identified are those consisting of stratified drift with flowtill as secondary constituents.[2] They were formed under running water conditions in the channels and other openings between and beneath the separating ice blocks of the disintegrating glaciers. Most of these features occur in regions of low relief and are therefore not common landforms in mountainous areas.

Eskers are elongated, sinuous ridges with occasional widening and other irregularities. Some eskers, like those in central Sweden, attain lengths of hundreds of kilometers.[5] The material within them is always well worn and sorted and may consist of pebbles, stones, or bedded sand. They are thought to be formed by subglacial rivers occupying tunnels in the stagnant ice.

Kames originated in different ways, but in general they resulted from deposition of drift within a depression (hole) on the surface of the glaciers. After or during the melting away of the ice, the stratified deposits would unmold and slump into conical hills. Kames are therefore more or less isolated mounds, while kame terraces are continuous features along a valley side. They are deposited by melt-water between the edge of a glacier and the valley wall. After ice disintegration they slump down to form the characteristic slopes of kame terraces.

Kettles are depressions or basins in glacial drift material. They were formed by the ablation of an isolated mass of ice burried in the drift. Thus kettles are, in a sense, the counterparts of kames. They occur singly or in groups of many so that the resulting landscape is sometimes described as "kame-and kettle topography". Kettles can have a diameter of a few meters to several kilometers, and they may have any shape in plan although most of them seem to be circular.

Proglacial Features

Two important landform types result from the deposition of outwash material at the end of glaciers. When the ice margin is wide, terminating on a broad lowland area, the melt-streams build up an extensive surface called outwash plain. These plains are associated with ice sheets. If the glacier is confined to steep valley walls, however, the resulting deposit is called a valley train.

The drainage system of a glacier consists of many small streams upon, beneath, and along the margins of the ice, and the head of the drainage basin usually goes as far back as the firn limit. Valley trains (and outwash plains) have braided stream patterns while deposition is in progress. This is due to the heavy bed loads which usually reach the limit of carrying capacity of the streams. In the warm season of the year, the emerging meltwater at the snout of a glacier can be very powerful and carry large ice blocks with it. Some of these large lumps of ice may be buried in the outwash sediments and eventually melting out to form kettle-holes.

The debris provided by the ice is normally very poorly sorted, so that the coarser material is deposited near the ice margin, and the flowing meltwater carries the finer sandy sediments farther down the valley. Since main valleys fill up with outwash material much faster than side valleys, outwash is built into the mouths of tributaries in fanlike or deltalike forms. Thus many tributary streams with small discharge are dammed and ponded by the fills at their mouths and form lakes. Two good examples of valley trains are provided by the alpine valleys of the Rhine and the Rhone in Switzerland. Both have extremely flat floors in their upper reaches, the result, perhaps, of deltaic sedimentation in lakes.[5,p.519]

HAZARDS

Some of the glacial processes taking place in the mountains can be very hazardous in that they threaten life and living space of men and animals. There are well-kept records from at least the 13th century on about the periodic natural catastrophies that worried the inhabitants of the European Alps up to the recent climatic amelioration of the 19th century. With the predicted and already recorded cooling trend of the present,[4] those alpine catastrophies of past generations might soon become reality again.

Glacial Surges

Glaciers, in general, respond very sensitively to climatic changes, advancing when cold and retreating when warm. Some glaciers, however, experience sudden, spectacular movements induced by dynamic conditions within the ice mass. After such surges the glaciers fall into a long interval of quiescence or even stagnation.

During a surge the middle segment farther upstream becomes thinner, while the terminal part thickens. This massive transfer of ice, without any net change in the mass of the glacier, results in frontal advances of several kilometers and changes in thickness of up to 100 meters. Rates of movement have been measured all over the world and some, e.g. in the Himalayas, were found to be as high as 10 km in two and a half months.[5,p.69]

In the past, glacial surges had consumed many uplands close to the glaciers in the Alps. Until the end of the 18th century the people of those marginal settlements lived in periodic fear to lose house and land by these sudden glacial advances. In 1642, for example, the glacier Des Bois near Chamonix advanced by "...over a musket shot every day.[1,p.170] And the glacier of Argentière, in the same year, advanced so fast that it threatened to carry away the whole village of La Rosière, after it had already covered cultivable fields and meadows.[1] Some surges have been minor and brief, but others have been major events in the history of a glacier. Surges seem to occur periodically but the cause for this apparently cyclic movement is unknown. One speculation is that during stagnations ice accumulates unproportionally in the upper parts, building up stresses which are released after the critical point has been reached. During this phase of instability the surge occurs, thinning the ice at the source and forming crevasses in very large numbers. An important factor for such movements is thought to be basal melting which permits the glacier (or a substantial segment of it) to slide on its bed.[2]

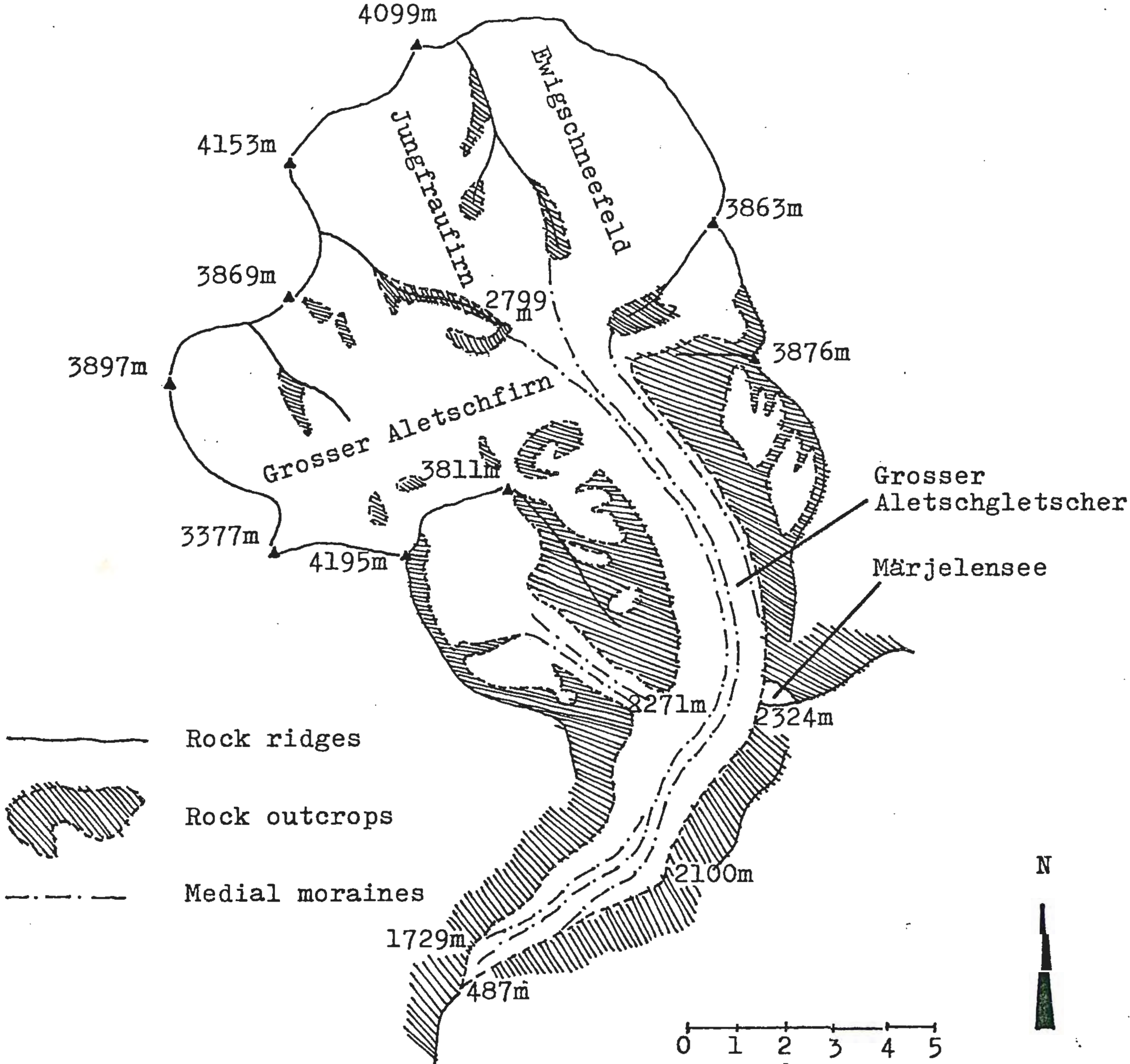
Most glaciers are known to have surged repeatedly in the recent past.[2,p.44] A remarkable year for surges in Alaska, for instance, was the year of 1966,[5] when several glaciers surged forward at impressive rates. The most reliable indicator for past surges is the presence of folded moraines within the glacier, especially the loops and folds in medial moraines.[5] It is therefore possible to recognize the cyclic nature of glacial surges which occur over short time intervals (2 or 3 years) followed by a surge-free period that can last more than 100 years.

Glacial Lakes and Catastrophic Bursts

Meltwater lakes are by no means rare today, but they are more restricted in their occurrence than in the Pleistocene glacial periods when the extent of ice was greater. In mountain valleys the majority of lake basins were formed near the margin of a glacier advancing over ice-free ground that sloped down toward the glacier. Others apparently were created when glaciers of a main valley blocked off those coming in from tributary valleys. Part of the side valley was then filled with meltwater for which there was temporarily no outlet. This situation created a large number of small and inconspicuous lakes in almost every glaciated region of high relief.[2,p.193] Most of these lakes are marginal in various ways, but some are supraglacial and others subglacial, and the majority of them periodically empty, often catastrophically, by subglacial or englacial routes. But as a result of the recent glacier stagnation and recession, the number of these lakes has been slowly diminishing.[5,p.533]

One of the most classical examples of this kind of lake is the Märjelensee in the Swiss Alps. This ice-dammed lake occupies a tributary valley held up by the Aletsch glacier (see Map 4). In the summer of 1878 the lake was 1600 meters long and about 80 meters deep, reaching its greatest dimensions.[5] Its normal mode of emptying has been into crevasses in the impounding ice wall, or beneath the ice. Between 1813 and 1900 the lake is reported to have emptied subglacially nineteen times, causing a flood-wave on the Rhone when in 1873 it released 10 million m^3 of water in 8 hours.[5,p.535]

A similar example is that of the Mattmarksee impounded by the Allalin glacier which crosses the Saaser Visp valley in Switzerland. In the past it has had a history of violent outbursts. In the middle of the 17th century the lake broke through the ice barrier several times and ravaged the valley repeatedly. It was so bad that the people vowed "...to abstain for forty years from banquets, festivities, balls, and cards."[1,p.176] But God's wrath apparently could not be mollified: The Allalin glacier kept blocking off the Saas glacier so that the ice-dammed lake mounted right up to the level of the huts on the mountain pastures of Mattmark. Between 1859 and 1909 26 outbursts of the lake were recorded, so that the valley was at least flooded every other year.[5,p.536] At present, the Saaser people enjoy "times of feast" because the recent climatic amelioration made the Allalin glacier retreat into its own valley, for the time being!



Map 4. Aletsch Glacier in Switzerland, showing the blocking of a tributary valley by the glacier, thereby forming the ice-dammed lake Märjelensee.

From Reference 5, p. 97.

MASS MOVEMENT PROCESSES

The altitudinal rising of the 0°-isotherm and the lowering of glacier surfaces result in rockfalls and heavy wastage of coarse material from the valley flanks above the glaciers. Through shrinking of the total glacier surface in the mountains, the areas of glacial erosion, gravel transport, and deposition become smaller, and the glacial processes are progressively replaced by others.

Exposed rock material is subject to different forms of weathering and in the process loses its physical and chemical stability. Weathering can be mechanical (temperature variations, frost, impact) chemical (oxidation, hydration, solution), or biogenic (root prying) and its intensity is highly selective and variable. But in any event, weathering and the resulting loosening of the rock masses are the preparatory phases of mass wasting, which is particularly dramatic and often catastrophic in high-relief terrain.

There are two principle classes of mass wasting to be considered, namely, mass movements per event and mass movements per unit time.[8,p.28] The first class relates to sudden transpositions of immense quantities of matter, such as rockslides, floods, and snow avalanches. The second class, mass movements per unit time, deals with the slow but continuous processes, such as the transport by rivers of chemically dissolved substances or the momentarily imperceptible creep of slopes.

The literature on mass wasting classification is rather confusing, for the nomenclature of specific types of mass movement is everything but uniform. The notion "landslide", for instance, is a very imprecise designation because it incorporates a wide variety of slides and flows. In the following sections mass movements of materials activated directly by the force of gravity will be discussed, whereby the name "landslide" shall not be used in a classifying sense.

Rapid Movements

Without topographic relief, mass wasting by force of gravity would not be possible. The significant landslides, therefore, take place on slopes, and rapid movement of the material is in most cases facilitated with increased water content since water counteracts the internal friction.[9,p.290] Huge masses of fast moving rock behave like a fluid, spreading out in the valley, and sometimes flowing some distance uphill on the opposite side of the valley.[10] Most landslides of rapid movement are immediately destructive and many of them cause damage to highways, railroads, buildings, power lines, and life.

Rockfalls

The typical denudation process in the unvegetated mountains in the Swiss Alps is the rockfall. It is the breaking loose and falling down of individual rocks of smaller or larger size. The major agent of this kind of disintegration is the frost that acts in the joints between bedding and cleavage planes. This frost action removes cubic rocks and huge blocks from the steep cliffs and causes them eventually to fall (See Fig. 9).

The extent of rockfall (and other mass-reducing processes) in granite and similar massive crystalline rocks is low. On the other hand, the brittle and heavily jointed dolomites and limestones are extremely subject to rockfall. It is below these cliffs where the most impressive talus accumulations (Schutthalden) are found.[8,p.29] These taluses have been accumulating since the retreat of the glaciers, and water erosion was not very effective in removing them because the dry colluvium of limestone and dolomite is very permeable.[8]

Even more effective is rockfall in shale material. An extensively studied example is the "Bündnerschiefer" in the upper Rhine valley in Switzerland. Here, contrary to the limestone and dolomite wasting, talus formation at the foot of the cliffs is almost impossible because of the effects of running water. First of all, shale is practically impermeable, and secondly, the fine platelets of the broken material are easily washed away by rain. The fact that limestone and dolomite cliffs form taluses, and those made of clay-rich shale do not, documents the important effect of permeability on talus formation.[8]

When a wall (cliff) borders on a river, rockfall scree is continuously washed away and the formation of a colluvial talus is made impossible. In general, the quantitative effect of the rockfall phenomenon is often (for psychological reasons, perhaps) slightly overestimated.[8] Even the great taluses of the high mountains (Hochgebirge) consist of freshly fallen material only on their very surface. In addition, the shapes of these taluses are morphologically misleading in that they suggest scree quantities beyond their actual volume. Nevertheless, some of the mountain passes and railroads in the Swiss Alps are exposed to rockfall and must be protected and repaired the year round.

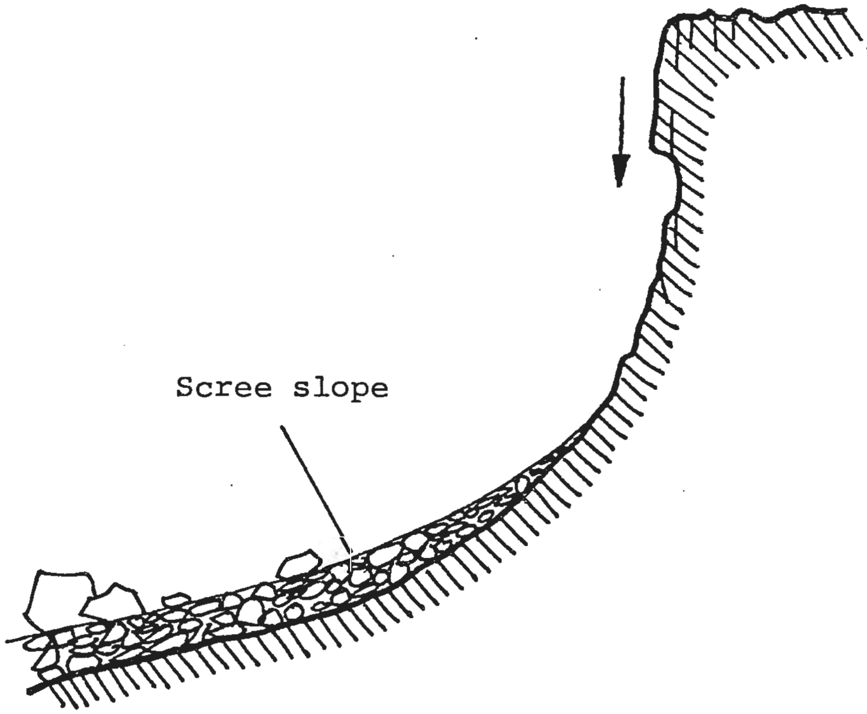


Figure 9. Rockfall: Cross section through mountain cliff showing wall and scree slope.

Rockslides

Rockslides develop when a mass of bedrock breaks loose and slides down the slope. The processes leading to this kind of landslide include increase of the pore-water pressure in the ground due to water seepage, undercutting by river erosion, man-made excavations, change in the ground water regime, earthquake shocks, and progressive structural changes in the material adjoining the slopes.[11,p.88] As soon as the average shearing stress on the potential surface of sliding becomes equal to the average shearing resistance, a landslide occurs.

A great number of rockslides take place during or shortly after heavy rainstorms and, consequently, most of them are ascribed to a decrease of the shearing resistance of the ground due to the "lubricating action" of the water that seeped into the ground. However, it usually requires only an extremely thin film of any lubricant to produce the full static effect characteristic of the lubricant. Furthermore, water can act as an anti-lubricant when in contact with certain minerals; for instance, the coefficient of static friction of smooth, wet quartz surfaces has a value of more than twice that of dry ones.[11,p.91] For these reasons, perhaps, the explanation "lubricating action" should be avoided.

Nevertheless, water does affect the stability of slopes in many ways, one of which is the rise of the piezometric surface. If the potential surface of sliding is located in sand or silt, the shearing resistance s per unit of area at any given point is given by the equation

$$s = (p - h\omega) \tan \phi$$

But if a material has cohesion, c per unit of area, the above equation is expanded to

$$s = c + (p-h\omega) \tan \phi$$

where s = shearing resistance per unit area.

c = cohesion factor per unit area.

p = pressure at the sliding surface due to weight of solids and water above the surface.

h = piezometric head (height of water above sliding surface at a given point).

ω = unit weight of water

ϕ = angle of sliding friction for the surface of sliding.

These equations show that with a rising piezometric surface, h increases, and the shearing resistance s decreases. Furthermore, when the water pressure $h\omega$ becomes equal to p the overburden "floats" and is held only by the cohesion, if that is strong enough.[11,p.92]

Rockslides (Bergstürze) are the most devastating of all rapid mass movements, and indeed their history in the Alps is a very tragic one, with thousands of lost human lives. Most rockslides in Switzerland have occurred in the months between March and October, and no preference for specific lithologies could be observed.[8,p.33]

On September 2, 1806, the town of Goldau in central Switzerland was wiped out by a rockslide which killed 457 people. A rock slab 5,000 ft long, 1,000 ft wide, and about 100 ft thick rested on a stratified mass of Tertiary Nagelflue (conglomerate with calcareous binder) which rose at an angle of 30° to the horizontal (see Fig. 10). The slab had occupied its position since prehistoric times, but during the heavy rainstorms of 1806 it moved down the slope. This classical landslide has been explained in three different ways:

(1) The angle of inclination of the slope had gradually increased on account of tectonic movements, until the driving force which acted on the slab became equal to the resistance against sliding.

(2) The resistance of the slab against sliding was due not only to friction, but also to a cohesive bond between the mineral constituents in the contact layer. Progressive weathering, however, reduced the total shearing resistance due to the bond. The cementing material might have been gradually removed either in solution or by the erosive action of water veins.

(3) The piezometric head h in the above equations could have assumed an unprecedented value during the rainstorm, whereas the cohesion c remained unchanged, provided it existed at all.

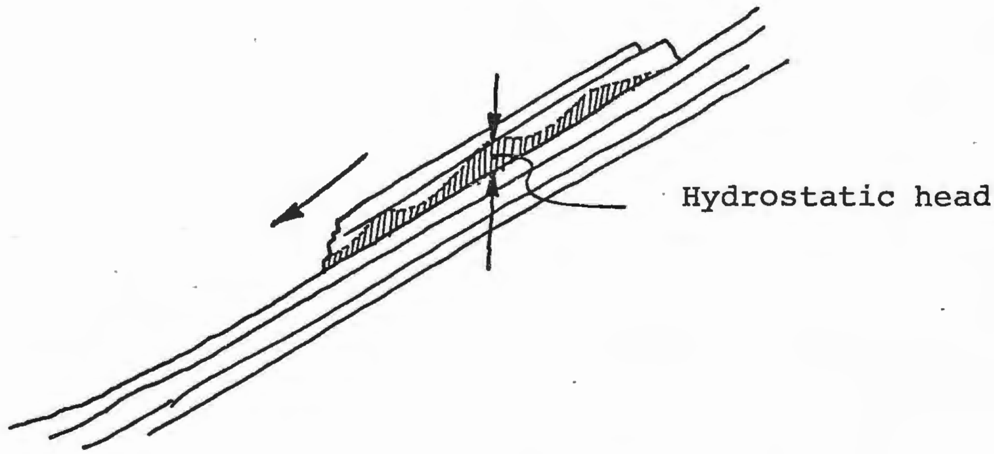


Figure 10. Diagrammatic section through site of rockslide in Goldau, Switzerland.

After Reference 11, p. 93.

Another classical slide was that of 1881 in Elm, Switzerland, also claiming many lives. In this case the rock mass was located on steeply dipping layered materials in which failure occurred rapidly. During excavation of a quarry for slate the slope had been undermined and the material started sliding, becoming a high-velocity flow with accompnaying high winds. The rock mass flowed across the valley floor and up the far side to a height of 100 meters.[12]

An indirect effect of a landslide is the flooding of the Vaiont Valley in the southern Alps (Italy) in the fall of 1963. The south bank of the valley above the Vaiont Dam (a 267 m high concrete wall) was not only very steep, but consisted of highly fractured oolitic limestone.[12,p.196] Failure took place along shear surfaces which were determined by the geological structure of the bank. The resulting landslide pushed 250,000,000 m^3 of soil and rock into the reservoir and displaced a tremendous volume of water which overtopped the dam by 100 meters. The water swept down the valley and killed almost 3,000 people.

Other Landslides

The most obvious difference between various kinds of landslide is the nature of the participating material. Some slides are entirely composed of rock, others of soils only, and a few are mixtures of ice, rock and soil.[12] Snow slides are called avalanches and will be dealt with in a separate section. But slide-material alone is not sufficient for classification; velocity, displacement, and mechanisms are additional, very important criteria.

For considerations of engineering works and effects on people, the speed at which a landslide moves is probably the most important factor, and the velocity of an event is usually related to its predictability. Future landslides are generally indicated by cracks which widen over a period of time. However, in very unstable areas the preliminary cracks may occur so rapidly as to go unnoticed before sliding ensues. The other significant factor related to the extent of damage a slide can cause is the displacement of material. If there is sufficient water present to transform the sliding mass into a fluid, and given the proper topographic conditions (e.g., high relief), even small slides may travel hundreds of meters. Thus material make-up, velocity, and displacement are closely related factors and none would be very potent in the absence of the others.

In general, the sliding material can be clearly distinguished from an underlying stationary bedrock or stable soil layer which does not take part in the motion. In most cases of rapid mass movement there is a shearing surface across which displacement takes place. Occasionally, however, it may be very difficult to detect a distinct transition layer, for the motion may die out gradually with depth. This situation emerges when the moving mass has the character of a viscous liquid. So if there is a clear shearing surface, the movement is a slide; otherwise it is a flow. The following paragraphs are a brief review of some common types of landslides or mass movement features.

Debris avalanches are relatively shallow mass movements that follow long and narrow tracks.[13,p.76] They usually develop by slippage along seepage lines on steep slopes and are later converted into incipient gullies. But debris avalanches can also occur on unchanneled slopes where they strip the surface (regolith) and allow weathering on bedrock to occur. In most cases these avalanches are triggered by heavy rains which increase the weight of the material and aid in its movement. When melting snow saturates the underlying regolith and a mixture of snow, water and debris flows rapidly downslope, the feature is called a slush avalanche.[13]

Debris slides are much like debris avalanches, but the water content within them is lower.[9,p.293] Thus they are less fluid, proceed rapidly downhill and eventually come to a stop and form tongue-shaped deposits.[14,p.117] Many debris slides are triggered on unstable scree slopes, especially when rockfalls occur from the cliffs above.

Mudflows move at perceptable rates on the order of 10 m/hr to flow rates comparable to that of water.[13,p.76] These flows are thus capable of destroying buildings and other engineering works. The distinguishing characteristics of mudflows are a basin-shaped source area, a long and relatively narrow flow track, and an expanded depositional toe zone. They occur most frequently on steep slopes with little soil cover (vegetation) and are therefore mainly associated with semi-arid and montane climates. They usually follow former stream courses and commonly recur in the same channels. Although short slides and small slumps may occur along their margins, they are strictly a flow phenomenon. Favorable conditions for the development of mudflows are:[2]

(1) abundant but intermittent water supply,

(2) absence of substantial vegetal cover,

(3) unconsolidated or deeply weathered materials containing enough clay and silt to aid in slipping of the mass, and

(4) moderately steep slopes.

It has been suggested (Terzaghi, 1950) that rapid flow could be initiated by spontaneous liquifaction if, at the time of movement, sandy materials lie below the water table. Furthermore, mudflows (especially the slow ones) have certain similarities with glaciers in that they develop a tongue. The flowing mass has sharp boundaries at its sides and base and moves almost "en bloc" along these boundaries.[14,p.143]

<u>Slumps</u>, also known as shear slides, tend to develop in situations where strong, resistant rock material overlies weaker rock. Movement, therefore, takes place along a definite shearing surface which is usually deep-reaching and spoon-shaped. Generally, two or more parallel slip surfaces are present, and the displaced materials form a series of blocks with their surfaces tilted back into the slope from which the slide descended (see Fig. 11). Slumps also occur in unconsolidated materials and are quite common in

artificial fills and cuts. Theoretically, assuming homogeneous material, shearing failure is dependent solely on height, slope, and shear strength of the material; in practice, however, geologic structure is also very important.[9]

Competent rock
overlying weak rock

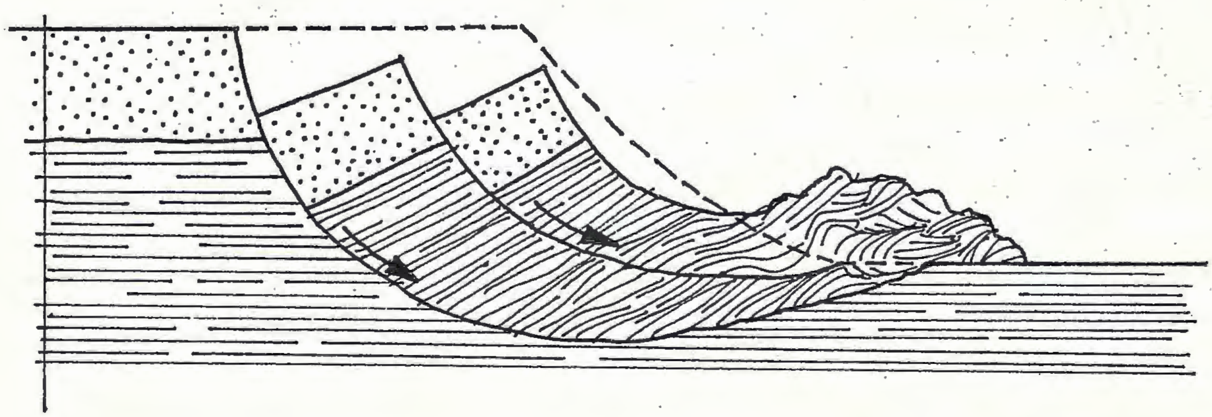


Figure 11. Slumping earth failure, showing curved slippage plane and reverse tilt of the slump blocks.

Adapted from Reference 10, p. 169.

<u>Earthflows</u> are the slowest among the rapid types of mass movement. They normally flow so slow that movement is imperceptible to the eye, and the turf cover may remain intact.[13] Often, however, earthflows are found to be

characterized by slumping at the head and bulging at the toe of the flow, in which case the turf would be considerably disturbed. Earthflows are most common in tropical regions, which have more porous surface layers underlain by a zone of imperveous clay. But these flows may frequently occur in the mountains of the northern hemisphere as well, and the final result of their movements can be catastrophic in terms of destruction to man-made works.

Stream Erosion and Landslides

Undercutting of steep slopes by river action is a major cause of all kinds of mass movement in the mountains. Yet erosion along the river banks is not a continuous process as far as intensity is concerned. Indeed the potent erosion events with extreme high water peaks are concentrated to very few days or even hours. All other fluvial erosion in between the few extremes is of strikingly little significance.[8,p.73] As long as the land is covered with snow, high water peaks do not occur and erosion during this time is minimal; actually, in the winter many high-altitude river beds are completely dry.

A first erosion phase arrives each spring during snow melting and lasts until early summer. A second erosion phase is brought on by glacier ice melting in high summer. The discharge curves of both the snow and ice melting phase show conspicuous diurnal periods, whereby the ascending leg of the period is shorter and steeper than the descending one. These daily periods are most distinct in streams with small watersheds, whereas they are no longer noticeable in larger rivers of main valleys. Contrary to thunderstorm discharges, pure meltwater discharges show never extreme peaks; they carry much water though, but insignificant load. Figure 12 below shows two typical meltwater curves of a mountain stream in Switzerland.

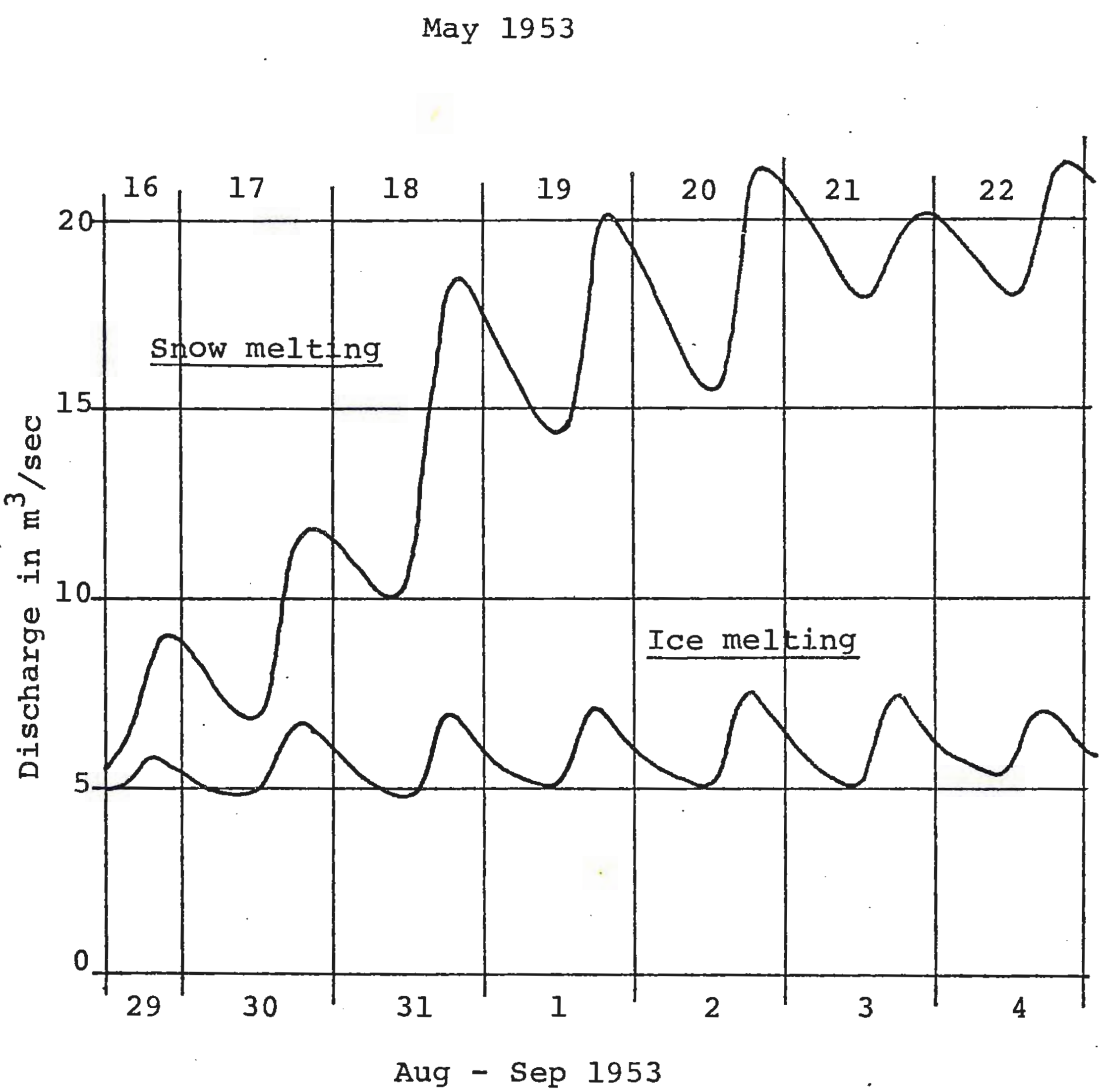


Figure 12. Discharge curves of a typical tributary river in the Alps (Landquart River, Switzerland), indicating conspicuous diurnal periods. The upper curve (spring) shows a strongly increasing discharge, whereas the lower curve (late summer) shows a somewhat constant discharge.

After H. Jäckli, Ref. 8, p. 74.

Thunderstorm discharges, on the other hand, are of much shorter duration but deliver at times very extreme high water peaks. This type of discharge is responsible for the second most effective erosion work which, along limited sections of a river, occasionally excavates more material in one hour than otherwise in a hundred years.[8,p.75] In rivers with glaciated watersheds, the relatively frequent glacier bursts (the sudden emptying of huge water-filled basins on, in, or under a glacier) surpass the erosion capacity of thunderstorm high water peaks by far, naturally. The glacier bursts (Gletscherausbrüche) give rise to an extreme one-time high water peak of very short duration which, depending upon the degree of glaciation in the watershed, can be devasting to the valley. In Switzerland, again, the Rhine Valley has little trouble with glacial bursts, but in the Rhone Valley they are the cause of periodically recurring catastrophies.[8]

Both vertical and horizontal stream erosion affect the stability of the sides of the valleys. When a river erodes the foot (base) of a stable slope, it increases the danger of a landslide which otherwise might never have occured. And when a river touches the foot of an unstable slope, it continuously removes at least that part of the slope material which is being wasted. Thus river erosion at the foot of unstable slopes increases creep velocity and promotes any kind of ground movement.[8] Enormous volumes of wasted material are removed from the slopes and redistributed by the rivers in this fashion every year. The inner structure of such unstable slopes is therefore very loose and ruptured by cracks and crevasses, so that an effective vegetation cover will never be able to establish itself. In an effort to retard this kind of surface degradation, many of the more important mountain streams in the Alps have been corrected or modified by expensive control structures (Wildbachverbauungen).

The primary function of these control structures is to reduce, or hopefully to prevent, both vertical and horizontal channel erosion and thereby to minimize the landslide potential of the valley flanks. The construction of lateral retaining walls at crucial points along a stream corrects and stabilizes the channel profile, reduces the gradient and thus aides the deposition of gravel to counteract the mass movements in the valley sides (see Fig. 13).

The construction of these stream control systems, of course, costs millions of dollars, but seems to warrant the security of montane villages and the protection of forests and other land resources. These engineering systems are therefore also important in terms of national economics. However, the geological consequences are considerable and cannot be disregarded.

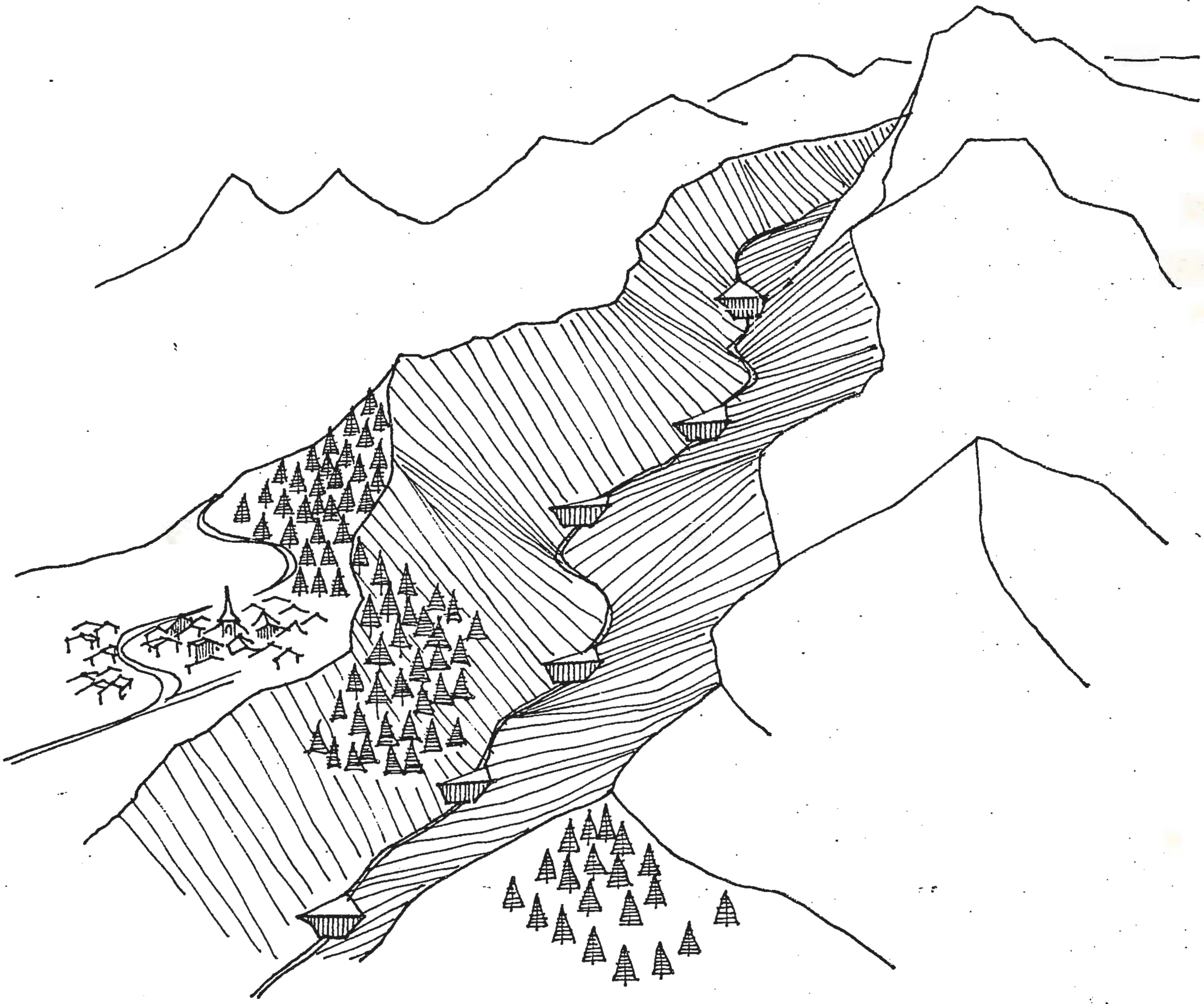


Figure 13. Steep valley with stream control system consisting of lateral retaining walls, designed to reduce erosion.

cannot be disregarded.

The prevention of vertical erosion in the corrected segments of a stream means that the most intensive degradation processes of the high mountains are locally neutralized. As a result, the collector rivers downstream carry less load and their equilibrium between erosion and deposition is significantly disturbed. For example, from the town of Reichenau on down, the Rhine River has deepened its channel by several meters in the last few decades.(8) This unprecedented erosion has been attributed to the construction of stream control systems all over the canton of Grisons (Graubünden).

SLOW MOVEMENTS

The rumbling rockslides and other sudden displacements of earth material often turn out to be a catastrophe. But besides these disasterous landslides there are surface slip features which move so slowly as to be imperceptible to the eye. However, these creeping movements are not as harmless as they are slow, for they are persistent in their progress down the hillsides and they can cause destruction and hardship. Soil creep cripples trees, tilts telegraph poles, displaces retaining walls, highways, and railway alignments. The low creep velocity though allows for protective measures before life could be endangered. Nevertheless, under special circumstances, as in heavy rain of long duration, a creeping slope might greatly increase its rate of movement, which may lead to a sudden and even catastrophic displacement of the soil mantle.

Solifluction

This type of slow mass movement is almost universally present on slopes in high latitudes and montane climates. The bulk of the material involved in solifluction is fine debris, but blocks of considerable size may be carried along in suspension. In the Alps, most of the solifluction is caused by the agencies of (1) snow melting, (2) ground frost, (3) ice lenses in scree slopes, and (4) snow pressure. The process of solifluction is the slow slowing from higher to lower ground of waste material saturated with water (definition by J.G. Anderson, 1906).

Meltwater is probably the most effective solifluction agent in the high mountains. On steep slopes, under the influence of gravity, water-saturated scree can travel considerable distances downslope, and areas covered with grass, above 2,000 m, become densly striped with compression ripples (Rutschwulste).[8]

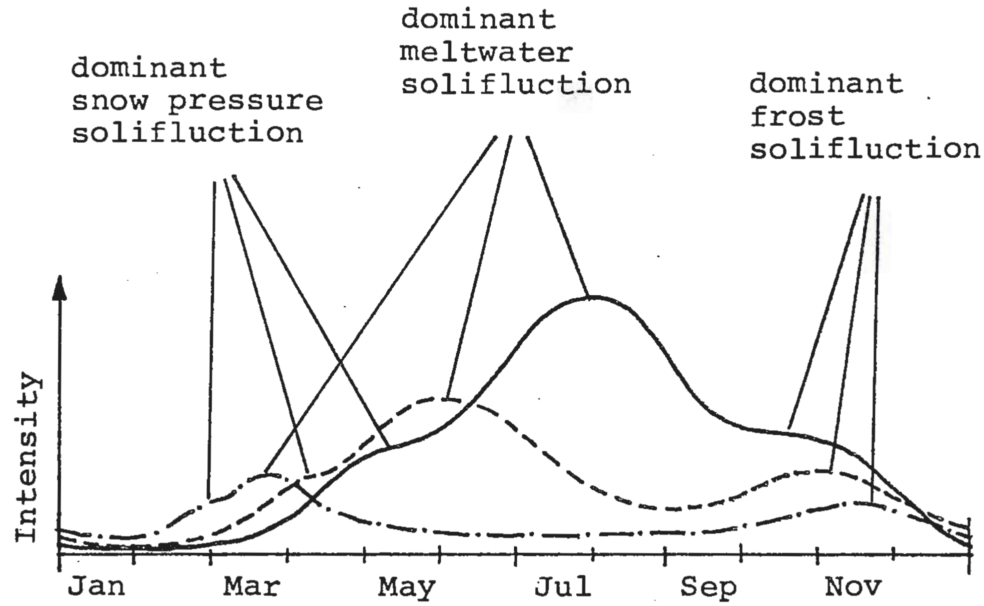
Ground frost is another factor in slow surface flow. In autumn, when the temperatures fall below 0°C, freezing begins at the surface and proceeds into the ground. As this process goes on, water in the capillary openings is pulled toward the growing ice crystals on the surface. The moisture from a large volume of soil is thus concentrated in the upper layers in the form of ice. During spring melting, this concentrated ice causes an over-saturation of water in the surface layers and results in a moving scree pulp (Schuttbrei).

At this point one should also mention the process of frost heaving, the expansion of certain soils that takes place during freezing. The mechanism of frost heaving is perhaps not fully understood but, in any event, larger particles in the ground are pushed upward by the expansive action of the growing ice crystals and finally end up at the surface from where they start creeping downslope. Important factors in frost heaving are particle size, permeability and excess moisture. In uniform materials with low range of particle sizes, it is unlikely for frost heaving to occur.

Another agent of solifluction is the formation of ice lenses in scree during cold nights in the fall before the slopes are covered with an insulating sheet of snow. The expanding effect of the growing ice lenses initiates movement in the scree accumulations.

Creep attributed to pressure by the snow cover is yet another form of solifluction. Due to its plasticity, the snow cover moves slowyly downhill and exerts shearing stresses to the ground. These plastic deformations reach their maximum in spring and cause considerable solifluction in loose ground.

These different types of solifluction features follow a temporal succession and, although they strongly overlapp, reach their maximum at different times of the year (see Fig. 14).



——— high mountains (more than 2,500 m)

- - - - middle altitudes (1,500 - 2,000 m)

—·— low altitudes (600 - 1,000 m)

Figure 14. Solifluction intensities at different altitudes.

From Reference 8, p. 36.

Rock Glacier Creep

Loose masses of coarse block scree mixed with fine-grained material which, under the influence of gravity, advance at very low velocity are known as rock glaciers (german: Blockströme; french: coulées de blocs) They normally have distinct lateral boundaries and show the form of valley glaciers with an articulate tongue. Active block streams of this kind move from a few decimeters to a few meters in a year, and vegetation on them is scanty or is missing all together. They are most common in mountainous regions that have been glaciated, and indeed they occur in great numbers in the Swiss Alps. In the region of the Upper Rhine (Bundnerisches Rheingebiet) alone, seventeen block streams (direct german translation) have been located.[8,p.41] They all occur in altitudes between 2,840 m and 2,260 m and have gradients of 26% up to 47%. Their lengths range between 220 m and 1,000 m and they are as wide as 200 m. An interesting observation in connection with these block streams or rock glaciers is that none of them is oriented to the south. Those oriented straight north are the ones that reach farthest down the valley and many of them originate in actual glaciers.[8] Their morphological characteristics are:

(1) Surface is convex upward and forward.

(2) Typical flow structures on the surface with bulging folds, convex forward.

(3) Steep and very distinct boundaries along front and flanks. The front scarp is extremely steep and therefore unstable and apt for rockfalls.

Rock glacier research is still in its beginnings and many questions abojt their origin have remained unanswered.[8.p.46]

Scree Slides and Talus Creep

The material involved in scree slides comes from moraines, weathering debris, rockfall scree and other sources. The major cause for the instability is the temporary water saturation during the snow melting season or during rainstorms, and the decrease in shearing resistance is ultimately responsible for the sliding movements in the slope. Another major cause of scree slope sliding is fluvial erosion, both deep and lateral, and cultivated land seized by scree slides in this manner is in most incidences lost.

Talus creep is supposed to be somewhat slower, although in cold regions it can be rather rapid where the major cause is the expansive force of alternate freeze and thaw of ice in the interstices. Thus adequate drainage is most important for talus stability.

Rock Creep

Movement in rock creep consists mainly of sliding rather than flowing. As a result of frost action and root prying, blocks of hard rock often creep gradually downslope. Other reasons for this type of creep are thin layers of argillaceous material beneath the blocks. Laminated rocks, such as shales, thinnly bedded sandstones and limestones either sag between strata or they show a reversal of true dip (see Fig. 15). Both types of creep are very common in the shales of the Upper Rhine region in Switzerland and are known as "Schieferrutschungen" and/or "Felsabsackungen."(8)

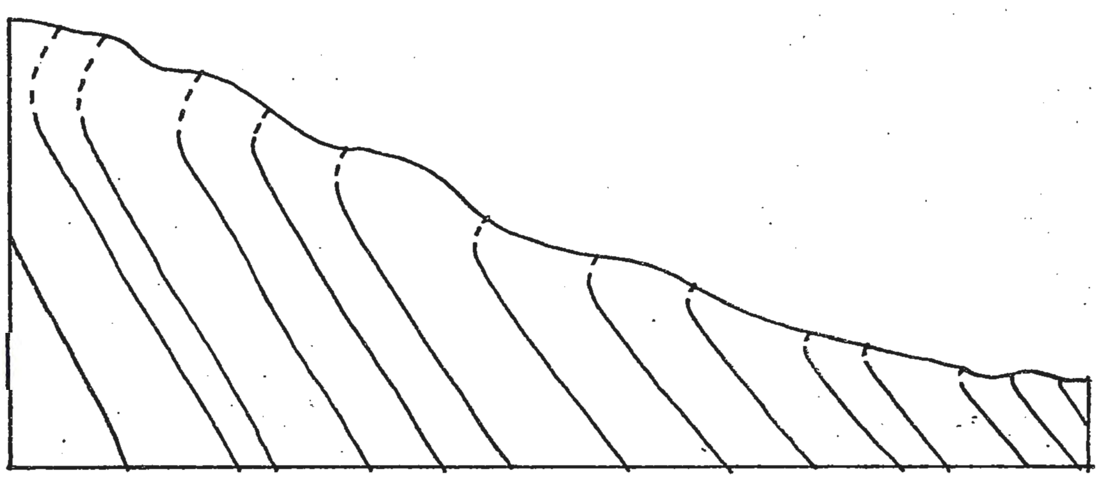


Figure 15. Outcrop curvature (reversal of true dip) resulting from dwonhill creep.
From Reference 9, p. 285.

Soil Creep

This type of movement is found on almost every soil-covered slope. A common evidence of creep are detached fragments of bedrock forming imperfect bands downhill. Where developed, they appear as a broken line of stones. The rate of creep on a hillside depends on climatic conditions, angle of slope, soil type, and numerous other factors. Soil creep is caused by a great number of agencies among which are frost heaving, alternate heating and cooling during day and night, moisture changes, and action of plants and animals. Dammage done to engineering structures by this type of movement is considerable.

SNOW AVALANCHES

The following section is a brief survey of the history of avalanche catastrophes in the Swiss Alps and, in general, a summary on formation of avalanches and protective measures taken against them.

One of the most prominent dangers of living in the Alps is presented by avalanches. Each winter thousands of tons of snow roar down the mountainsides, wiping out everything in the way. The high mountains of Switzerland have been colonized (settled) for hundreds, maybe thousands of years, and the people have learned to adapt to the natural hazards of this rugged environment. In some locations avalanches occur regularly every year, and the residents of nearby villages are able to plan around them; but in other places they are much less predictable, occurring only once or twice a century. Thus avalanches create serious limitations in Alpine land use.

In the last 76 years (1900-1976) 1,388 people have been killed by avalanches in the Swiss Alps.[19] In the period from 1900 to 1945 the average number of yearly deaths was 12, but since the popularization of skiing, that number has gone up to 18; and these figures do not include the victims of catastrophes. The statsitics of avalanche accidents distinguish between casualties of catastrophes, workers (woodsmen and farmers), and tourists (skiers). It was found that the relative number of deaths between workers and skiers varies with the amount of snow in any given winter (see Fig. 16)

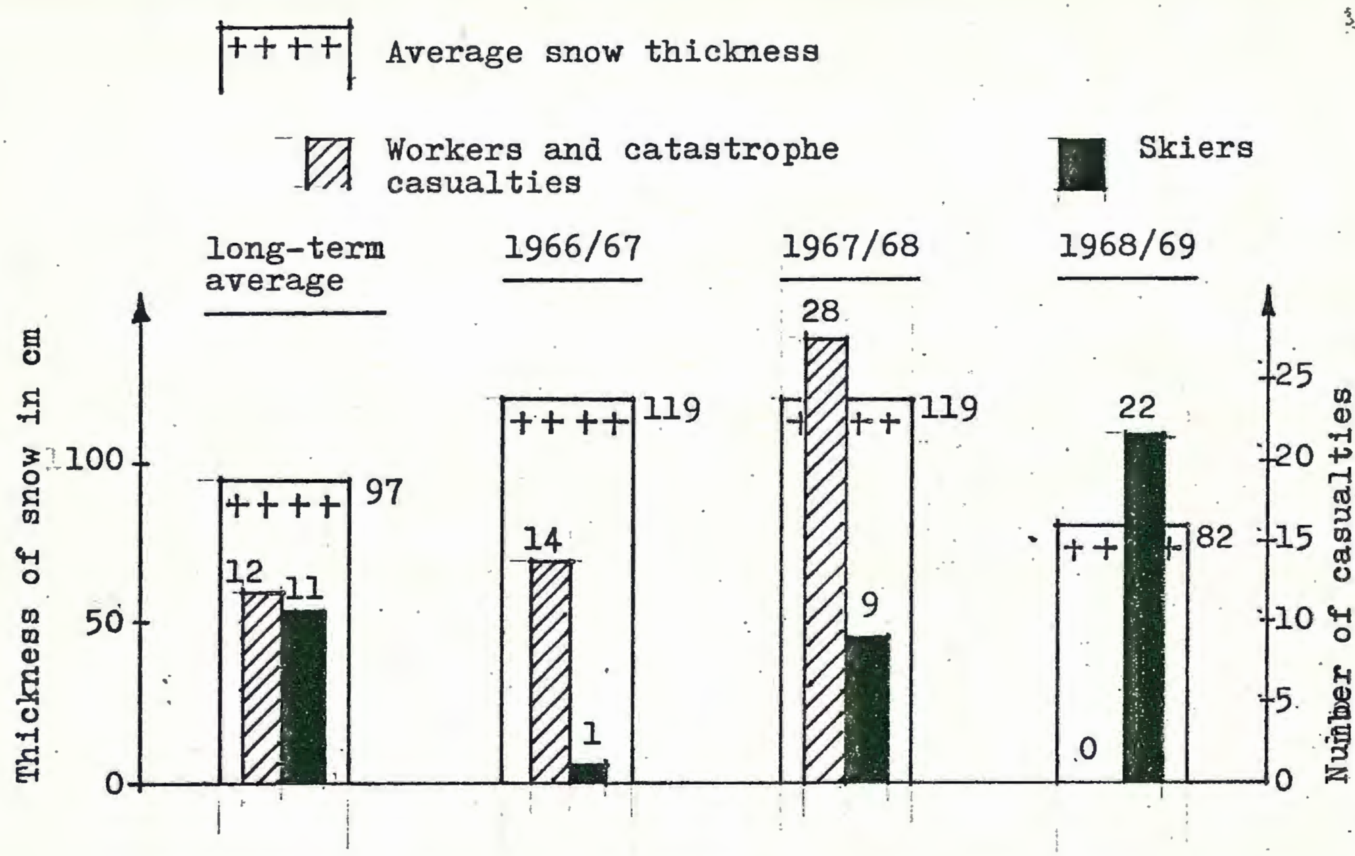


Figure 16. Distribution of avalanche casualties for high-snow and low-snow winters in Switzerland.

After Frutiger, Reference 19.

In the 31 winters since 1946 about 4,400 avalanche accidents had been described and analyzed.(19) The results of some of the great catastrophy winters in the last 100 years are summarized in Table 2. Catastrophy situations are primarily evoked by heavy snowfall. They are regionally confined but can occur one after another at different locations in the same winter, and thus affect the entire Alpine area. Also, region-specific patterns of cyclic recurrence have been noticed; the average catastrophy recurrence for all Switzerland being about seven years.(19)

Table 2

AVALANCHE CATASTROPHES IN THE SWISS ALPS*

Winter	Avalanches causing damage	Persons		Livestock killed	Buildings destroyed	Forest damage in m^3 wood
		buried	killed			
1887/88	1,094	84	49	665	859	82,100
1916/17	?	?	39	?	200	82,100
1934/35	?	?	36	?	?	25,000
1944/45	500	56	39	?	300	40,000
1950/51	1,300	234	98	884	1,489	710,000
1953/54	325	159	33	228	634	10,300
1967/68	421	106	37	23	404	25,400
1969/70	254	120	56	3	98	41,000
1974/75	1,022	127	27	172	305	146,000

*This table is incomplete. It does not list all catastrophes.

After Frutiger, Reference 19.

Avalanche Formation

Although catastrophic avalanche episodes are usually directly connected with heavy snowfalls, the condition of the old snow cover plays an important role. New snow falls under the most varied weather conditions. It may fall at temperatures ranging between -30°C to +50°C, and at wind velocities from 0 to around 200 km/hr. Wind turbulence leads to fragmentation of the original new snow crystals so that the layered new snow can display a highly variable consistency from the outset, ranging from the lightest, fluffy to the fine-grained, stiff snow, whereby the density may vary between 30 and 250 kg/m^3.[17]

Immediately after it deposits, new snow undergoes a decomposition transformation. It first becomes felty and then round-grained and at the same time progressively increases in density. Independent of this and often progressing simultaneously, there is a constructive transformation taking place. It is characterized by a general growth of selected snow grains with a change in their shape tending toward edged, planar forms. This transformation is called temperature-gradient metamorphism. If snow is warmed by warm air or by radiation to the point of melting and is subsequently cooled again, it undergoes the melting transformation during which it changes to a round-grained adhesive mass.

In response to temperature changes, the surface layer takes on a powdery consistency, and after a further snowfall, the old loose surface layer remains in the interior of the snow cover and prevents binding between new and old snow. In great snowfalls, superimposed layers undergo only slight temperature-gradient metamorphism, remain fine-grained, and under the snow cover burden attain densities of over 500 kg/m^3.(17) The various types of snow as they are characterized by the nature and gedree of their metamorphism have highly variable mechanical properties, and the structure of the snow cover looks quite different from winter to winter. This condition results in a variable long-term basic disposition to avalanche formation.

Density, strength and deformability are the snow properties which are decisive for avalanche formation. Like in glacier ice, the shear stresses in a snow cover increase from the surface down. Thus, the ground everywhere supports the immediately superimposed snow weight in the form of normal pressure and shear stress. Under the prevailing stresses, the snow suffers a continuous deformation which consists of a volume reduction and a shear deformation. The latter is recognizable as a slow creep motion parallel to the slope which ranges from fractions of a millimeter to several centimeters per day, depending upon the nature of the snow, slope inclination, and temperature gradient.[17,p.8] In pure creep motion the snow adheres to the ground; either because it is frozen or prevented by the ground roughness from undergoing a gliding motion. If, on the other hand, the ground is unfrozen and smooth, then there is superimposed upon the creep a gliding motion.

Creep and sliding motion can locally be modified by differences in the slope inclination, in the slope characteristics, or by direct obstructions. If this is the case, there arise marked dislocations of stress through the emergence of both tensile and compressive forces. Where the stress first attains the strength limit, the primary fracture occurs in the snow cover across the slope, and a coherent slab avalanche is about to go down.

In addition to the slab fracture type, another type is observed which cannot be interpreted by means of the above-described mechanism. In this case, snow motion develops from a point at the surface and acquires breadth and depth with the snow moving from the outset in the form of individual crystals or clumps. A small impulse suffices to set a clump of snow in motion when it does not start spontaneously. It collides with particles lying farther down the slope and these in turn attach themselves to the motion and propagate the so-called loose-snow-avalanche.

This type of avalanche originates, as the name indicates, with snow of slight cohesion. A loose-snow avalanche developes more slowly than a slab avalanche, and since in its upper course it usually includes only a thin layer, there it is relatively harmless. Also of essential interest is the fact that an avalanche like this arises only on very steep slopes (over 40°) and only beneath some disturbing agency such as a skier.[17]

Avalanche winters have always been closely related to periods of intense snowfall. Therefore, it is important to know where and with what frequency great snowfalls are to be expected. In the Swiss Alps, catastrophic events must be reckoned with whenever the new snow totals (over several successive days) amount to about 120 cm.[17,p.12] On the other hand, snowfall pauses at high temperatures permit the new snow to consolidate and delay or prevent the development of an avalanche. In general, there exist distribution in the Alps, and every region must consider its unique environmental conditions.

Protective Measures

During the last fifty years the former mountain hamlets have grown into sophisticated summer and winter holiday resorts, and more and more people live part of the time today in the mountains. The enlargement of the mountain hamlets and the associated expansion of transportation routes in and through the mountains confront the responsible authorities of entire regions with almost insoluble problems. As a result of this rapid development, avalanche control has become more important than ever before.

It is assumed that structural control of avalanches had its inception with the settlement of the mountain valleys. Haystacks, herdsmen's huts, and stables had to be protected from the outset by means of inclined terraces at the back and by splitting wedges. Such types of avalanche protection may be found in most alpine valleys. Early in the 19th century, organized attempts by entire communities had been made to protect roads, mountain passes, and hamlets. Galleries were introduced into the most avalanche-exposed parts of the highways. In other instances, the people had fixed posts of larch into the ground on steep mountainsides, or they dug series of horizontal ditches in order to anchor the snow cover (see Fig. 17). A primary objective in connection with these structural control systems was afforestation of the steep slopes, and today it is still a difficult struggle for the federal forest service to maintain and reconstruct the protective forests.

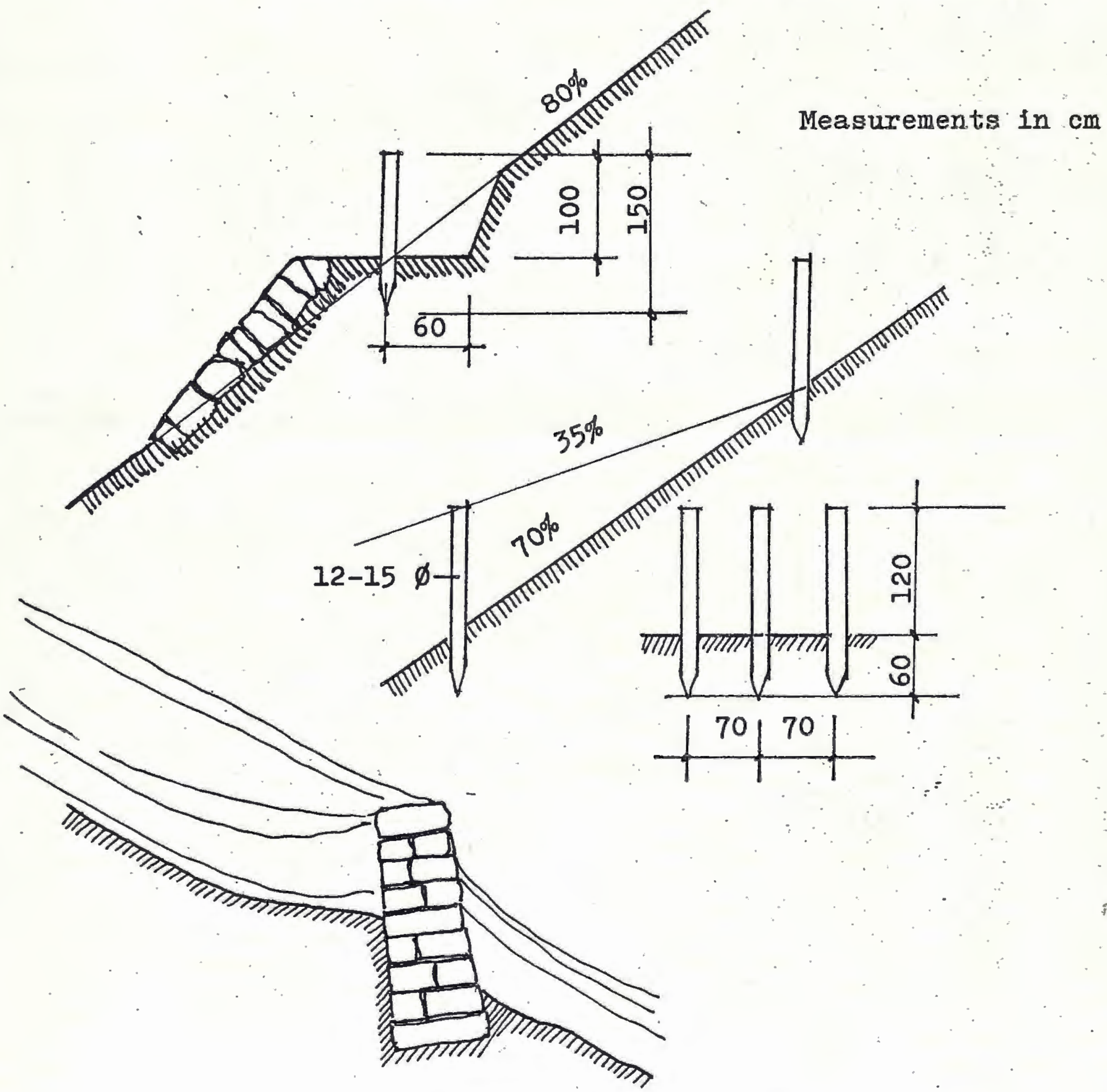


Figure 17. Early avalanche control structures in the Swiss Alps.

From Reference 17, p. 41-42.

Todays control systems are supporting structures capable to hold the snow cover in such a way that there can arise either very little movement or none at all. Avalanches are now prevented in their starting zone so that the steep walls of the bare high mountains are systematically secured with continuous arrangements of the typical snow bridge elements, made of steel, aluminum, or concrete. Figure 18, below, illustrates the principle.

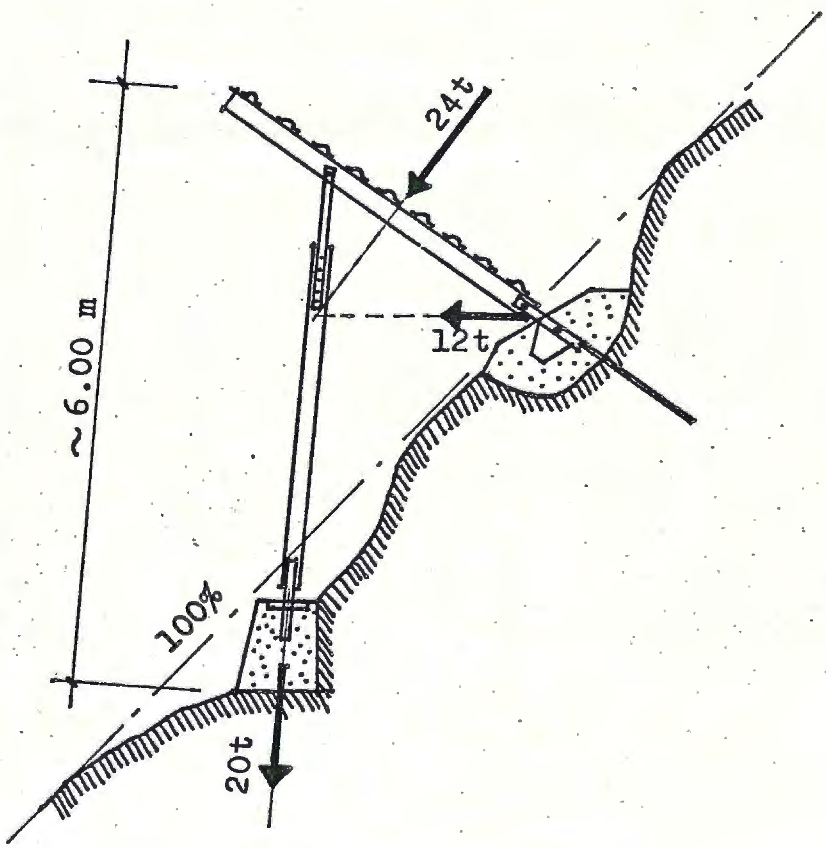


Figure 18. Section through a supporting structure (Schneebrücke), indicating snow pressure forces calculated for a snow thickness of 5 meters.

After Reference 19.

Corresponding to their nature and mode of action, two different types of structures are distinguished, namely, structures in the starting zone which prevent avalanches, and structures in the track and runout zones which reduce the damaging effect of descending avalanches. In the starting zone, support structures (see Fig. 18) and drift structures are used, while deflection and retarding structures are introduced to protect objects in the track and runout zone.

In control-structure regions above treeline, permanent installations are set up which consist of durable materials. Within the tree zone, however, the requirements for durability do not seem to be so stringent; here it is temporary construction, for immediately after setting up the installation, usually consisting of impregnated wood, the area is afforested with suitable types of trees. The supporting structures, especially those above treeline, are built in continuous arrangement consisting of long, horizontal series of structures which extend over the entire breadth of the terrain to be controlled. These lines are interrupted only by parts of the terrain which are safe from avalanches. Actually the structures are erected in different arrangement patterns depending upon the types of structure elements, topographic conditions, and the significance of the objects to be protected. The other type of structure in the starting zone is the snow fences or drift structures. They are designed to prevent the deposition of wind-blown snow accumulations which could become the impulse for triggering an avalanche.

The deflection and retarding structures in the track and runout zone are constructed in cases where the onset of an avalanche cannot be prevented for either financial or physical reasons. Deflection structures are back-filled masonry walls or dikes designed to change the direction of a descending avalanche. Included in this category of structures are galleries, splitting wedges, and inclined terraces, all of which lie directly in front of the protected object. The splitting wedges, by the way, are prow-shaped headwalls of buildings facing an avalanche-prone slope. This mode of building reinforcement is as old as alpine civilization.

The retarding structures in the runout zone are designed to shorten the extent of an avalanche or to break it up in order to reduce its mighty power before it reaches a settlement. This objective is achieved by earth structures, such as trap dikes, trap walls, series of retarding mounds, and retarding wedges.

The artificial triggering of avalanches under controlled conditions is a further measure for the prevention of avalanche accidents. It is employed principly to protect tourist regions and transportation routes. The cheapest method of artificial triggering is the technique of stepping or jumping on the snow to release an avalanche. True preventive triggering, however, is done by blasting loose or shooting down avalanches by means of explosives. The first avalanches were shot down in the Bernese Alps during the winter of 1934/35. These were experiments carried out by the division of military engineers in which the infantry cannon, the 75 mm mountain gun, and the 81 mm mine thrower were used. Today, in Switzerland, during a normal winter, between 5,000 and 10,000 explosive trials are run for the purpose of triggering avalanches. The standard firing device is the 81 mm mine thrower (trench mortar).[17]

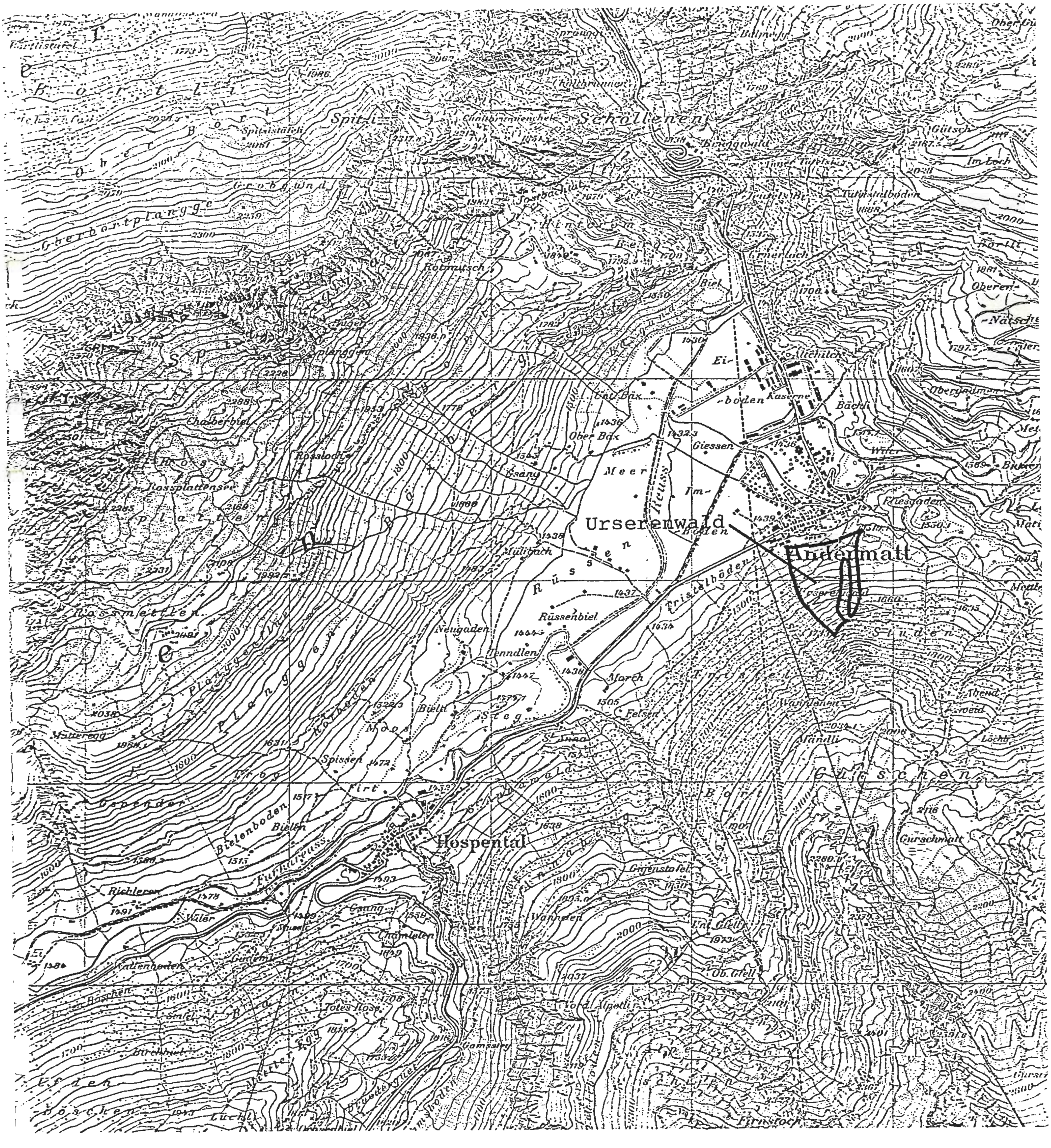
LAND USE CONSEQUENCES

The first prerequisit for settlement of a mountainous region is the protection of its communities and transportation facilities against more or less predictable and recurrent catastrophes, such as glacial burst and snow avalanches.

As the most important of these preventive and protective measures, avalanche defense acquires increasing significance. Both the maintenance of individual mountain communities and the development of entire valleys in the alpine area depend on it.

As a result of the rapid development of travel and winter sports activities, avalanche control has become even more important, because more people are entering into regions exposed to avalanche risk. Not the number of avalanches, but the number of risk-exposed people is increasing. Thus, it is man himself who increases the risks in his environment and endangers his own safety.

The original mountain dwellers had built their communities in close clusters in the most secure, or let us say in the least dangerous locations in the valleys. Many of the old hamlets and villages were built at the foot of a forested slope, for the forest would provide a natural protection from avalanches. A good example of this type of planning concept is the town of Andermatt in the Gotthard region. Andermatt throughout its history has been protected by the "Urserenwald," a small but well maintained forest on the north flank of the Gurschen mountain (see Map 5).



Map 5. Location of Andermatt (Hometown of Bernhard Russi) showing typical alpine community setting behind a forest, in this case the Urserenwald.

From Blatt 1231 of Landeskarte der Schweiz.
Scale: 1 : 25,000

N

The modern outsiders, who have been invading the Alps since the second world war, do not have this keen understanding for the hazardous environment of the mountains. Increased mobility and the "second home idea" brought people from the lowland into the mountains by the thousands. The real estate business started booming, the villages began to sprawl up the hillsides, hotels grew out of the ground, and sports centers expanded their activities. This sudden growth of the alpine communities resulted in an inefficient and dangerous development pattern which could not be allowed to go on unchecked. The coordination of regional planning efforts became inevitable, and an important basis for cooperation in this respect was the creation of the federal avalanche cadastral survey which has the following three goals:

(1) Scientific investigation of the entire avalanche phenomenon.

(2) The establishment and followup of comprehensive damage statistics.

(3) The determination and mapping of all zones endangered by avalanches.

On the basis of this survey, avalanche zone plans are now being drawn up for the individual communities and regions.

Like the avalanche survey, the avalanche zone plan is also a basic part of avalanche protection. It contains all danger zones in the settled and tourist regions and, as a part of local planning, serves the competent authorities in the evaluation of building applications. The avalanche zone plan provides three zones:

(1) The red zone is characterized by a general building prohibition.

(2) The blue zone has a certain degree of avalanche danger. Hence, dwelling houses are permitted in the blue zone only under the provision that certain safety specifications are met.

(3) The white zone has no avalanche danger.

The designation of avalanche zones can have drastic effects upon the property owner since the zone plan creates public law limitations which relate to the right to freely dispose of real estate. However, from this no compensation claims can be inferred since the agencies involved are forces of nature.

Although life in the mountains has become more comfortable, the basic relationship between man and geological processes in this high-intensity environment is still dramatic and hazardous. How permanent the conquest of the Alps really is, will be tested by the future glacial advances of a new ice age.

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