GLACIERS and GLACIOLOGY

by

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Glaciology

The study of existing or modern glaciers in their entirety, involving all related scientific disciplines. As glaciology embraces so many interconnecting facets, it is considered a master science. It is largely concerned with present glacial characteristics and processes, as opposed to studies in glacial geology which relate to the nature and effects of former glaciation.

A glacier is a naturally accumulating mass of ice that moves in the process of discharging from head or center to its margins or a terminal dissipation zone. Glaciers are nourished in areas of snow accumulation that lie above the mean climatological or orographical snow line, which on the glacier surface is referred to as the névé line or firn line. The most active glaciers are generally found in regions receiving the heaviest snowfall, such as the maritime flanks of high coastal mountain ranges. Exemplifying this are the great westerly facing glaciers of Mount Saint Elias and the Saint Elias Mountains in south coastal Alaska which are nourished by the heavy precipitation brought in by warm, cyclonic air masses moving across the warm waters of the Gulf of Alaska. Similarly, in New Zealand the vigorous glaciers of the southern Alps lie in the storm tracks of the prevailing westerly wind which brings much greater accumulation to the western than to the eastern slopes. Other such maritime glacial bodies are the Patagonian ice field in the southern Andes; the small ice sheet and ice caps of Iceland; the mountain ice fields of Norway and the Kebnekaise region of Sweden; and the mountain glaciers of eastern Siberia. There are many other glaciers in the high mountains of the middle and equatorial latitudes (as in Peru and East Africa); however, 96% of the world's glacial ice is represented by the vast continental ice sheets of Antarctica and Greenland. Together, these regions contain at least 5,600,000 mi^2 (14,300,000 km^2) of ice cover, or nearly 10% of the total land area of the globe. *See* SNOW LINE.

The total volume of the world's glaciers, ice fields, and ice sheets can only be estimated, but is at least 24,000,000 km^3. This mass of frozen water, if melted and returned to the sea, would raise the sea level 160–200 ft (±60 m). R. F. Flint in 1957 estimated that, during the Ice Age, this volume was probably 300–400% greater. *See* GLACIAL EPOCH; TERRESTRIAL FROZEN WATER.

Glaciation and deglaciation. In many regions glacier activity has formed a continuous series of events throughout the Pleistocene glacial-pluvial epoch, and in postglacial (post–Ice Age) and modern times. In order to simplify the terminology of past and present glaciation, regions which are presently glaciated are also sometimes referred to as glacierized, glacier-covered, or ice-covered. These terms are used instead of the ambiguous and unqualified term glaciated for reference to all areas formerly or currently ice-covered. In this context, for any one region, the glaciation limit is represented by the lowest elevation mountain summit carrying an existing glacier.

A glacier at any one time presents only partial patterns of its long-term regime. The complex study of glacial regimen, the processes and consequences of growth and decay, is being actively pursued in glacial regions throughout the world today. Other glaciological studies embrace research in many allied disciplines, such as geomorphology, meteorology and climatology, physics of ice deformation (glacier mechanics or continuum mechanics), thermodynamics, survey and mapping, glacier geophysics, lichenometry, palynology, and plant ecology.

Materials. Glaciers are composed of three substances: snow, firn, and ice. The main material of glaciers is bubbly glacier ice of a specific gravity approximating 0.88–0.90 g/cm^3. This glacier ice is composed of myriads of interlocking crystals, hence it is a polycrystalline material containing air pockets and entrapped water bubbles. Because of the proportion of bubbly glacier ice, a mean bulk specific gravity may be taken as 0.90 g/cm^3, as opposed to a specific gravity of 0.917 g/cm^3 in dense, solid, and unaerated ice. Below the late-summer glacier snow line (névé line, as defined below), only bubbly glacier ice is exposed. Above the névé line, the other categories of snow and firn (and firn ice) exist to depths of a few to hundreds of feet. It is deeper in polar (colder) firn packs. (Firn is a consolidated granular transition of snow not yet changed to glacier ice. The process of transformation of snow to firn is termed firnification.) The density gradation between firn and ice is usually asymptotic, but with a relatively sharp line of demarcation between a seasonal snowpack and underlying firn. Arbitrary and approximate densities of new snow are 0.1–0.3; old snow, 0.3–0.45; firn, 0.45–0.75; and firn ice, 0.75–0.88. Firn ice is not considered to be a separate stage of metamorphosis but rather the result of a mixture of partially altered firn and bubbly glacier ice. The processes by which new snow is transformed into bubbly and dense glacier ice are complex and varied. In middle-latitude regions the refreezing of percolated meltwater, compaction, and flow deformation play substantial roles in the metamorphosis. In polar regions wind packing, mechanical compaction, and flow recrystallization are the prime factors producing glacier ice.

Terminology. The geographical term névé refers to the area covered by perennial snow or firn, that is, the area lying entirely within the zone of accumulation. The word firn refers only to the substance of the material itself. The terms snowpack and firn pack refer to the volume of snow or firn, respectively, at any one point on the névé of a glacier's surface and connote thickness or depth characteristic rather than area. Snow cover and firn cover refer to the blanket of snow or firn over a névé, a glacier, or a bedrock surface, and have areal rather than depth connotations.

The term névé line (firn line) represents the elevation of the periphery of a névé at any point on a glacier's surface. More specifically, the elevation of the névé line's most stable position over a period of several years is referred to as the semipermanent névé line. The term glacier snow line or transient snow line describes the transient outer limit of retained winter snow cover on a glacier. Its elevation gradually rises until the end of the annual ablation season, by which time the old snow has become firn and the glacier snow line becomes the seasonal névé line. In regions such as coastal Alaska, where glaciers descend to sea level, the summer snow line on intervening rock ridges and peaks is often much higher than the snow line on the valley glacier, and in most instances is more irregular and indefinite. If it has not disappeared completely from the bedrock surfaces by the end of summer, the lowest limits of retained snow may be connected as an irregular limit and be termed the orographical snow line since it is primarily controlled by local conditions and topography. Variations in the position of the orographical snow line are so great from year to year that only from records over a long period can a meaningful trend be discerned. This snow line is important in the development of nivation hollows and protalus ramparts in deglaciated cirque beds. Of greater present-day significance is the average regional level of the orographical snow line rather than its lower limit. This mean position, based on observations over a number of years, is called the regional snow line or climatological snow line. On adjacent ice sheets and glaciers this coincides with the mean névé line. In regions of extensive present glaciation the regional snow line thus equates to the mean névé line. Specifically, the mean névé line is the statistical average of consecutive annual positions of semipermanent névé lines over a period of at least 10 years.

In glaciers of the polar regions a different term is used to delineate areas of net accumulation from those of net wastage, in consequence of the refreezing of meltwater and drainage on surfaces down-glacier from the transient snow line. The position where this refreezing ceases is referred to as the equilibrium line, a theoretical line separating the area of net gain from the area of net loss. On temperate glaciers this coincides with the névé line. It is an important concept in glacier velocity considerations because it represents the position or zone least subject to seasonal variations in flow resulting from excessive accumulation or ablation.

Morphological categories. Glaciers develop numerous forms. Basically, they are of the mountain (alpine) type and of the plateau (polar) type. Mountain glaciers generally are moderate to small

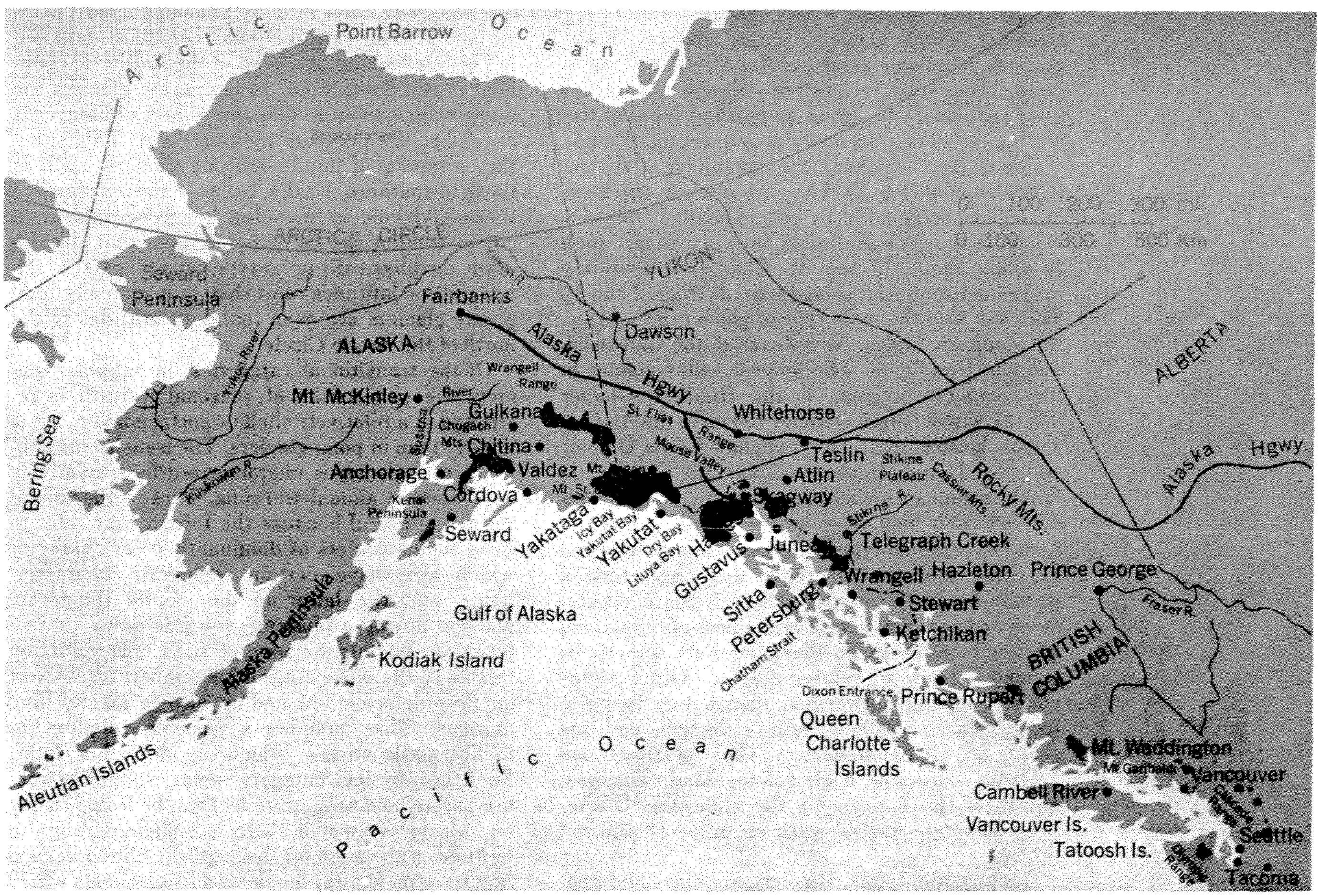

Fig. 1. Map of Alaska showing main centers of existing glaciers. Note that these centers lie along the south and the southeastern coasts, where there is a ready source of moisture and high ranges of mountains, resulting in heavy annual accumulations of snow. In northern Alaska it is too cold and dry to produce the requisite snowfall for glaciation under present climatic conditions. (*Foundation for Glacier and Environmental Research*)

Fig. 2. Oblique air photograph of 4-mi wide tidewater terminus of the Hubbard Glacier, a prototypical valley glacier which reaches tidewater from a source area in the St. Elias Mountains of Alaska and the Yukon. This glacier has advanced several miles since 1894. *(Photograph by H. B. Washburn, August, 1938)*

in size and include valley glaciers (main ice streams), icefall glaciers, cirque glaciers, basin glaciers, hanging glaciers, cliff glaciers, and glacierets. These terms are self-descriptive and for the most part relate to strong and varied relief of the kind found along the mountainous southern coast of Alaska (Fig. 1). Valley (ice stream) types are the most common (Fig. 2). They are often in the form of glacier systems fed by cirque-headed tributary valleys and serve as outlets from ice fields, such as those found in the St. Elias and Boundary ranges between Alaska and Canada (Figs. 2 and 3). They are also the main type of glacier in the Alps, the southern Andes, New Zealand, the Caucasus, and the Himalayas. The longest valley glacier in the temperate regions is the Hubbard Glacier (Fig. 2), with a length of about 100 mi in the Alaska-Yukon border area. The Vaughan Lewis Glacier and the Upper Herbert (Camp 16) Glacier on Alaska's Juneau Icefield are typical icefall glaciers derived from high névé basins or plateaus (Figs. 3 and 4). Plateau-type glaciers, dominantly of the ice-sheet and ice-cap form, are characterized usually by vast size with relatively flattened surfaces or low relief. These are typical of the Greenland and Antarctic ice sheets, and are sometimes termed inland or continental ice. Often valley glaciers extend outward as distributary tongues. Intermediate between valley glaciers and ice sheets are piedmont glaciers. These occupy broad lowlands bordering a glacial highland. The best known of this category is the Malaspina Glacier near Yakutat, Alaska, with an area of 1400 mi^2 (Fig. 1).

Geophysical types. Glaciers are classified geophysically into two major groups, polar and temperate, and two transitional groups, subpolar and subtemperate (Fig. 5). The temperature of a polar glacier is perennially subfreezing, except for a shallow surface zone which may be warmed for a few weeks of each year by seasonal atmospheric variations. The extreme polar condition depicted in Fig. 5 is found in the heart of the Antarctic continent at the South Pole. In temperate glaciers the temperature below a recurring winter chill layer is always at the pressure melting point. This situation is typical of middle-latitude glaciers, such as those in southern Alaska. Because these terms are thermodynamic in meaning but geographical in connotation, it should be pointed out that glaciers of the geophysically polar type can still exist at relatively low latitudes, and that geophysically temperate glaciers are even found at latitudes to the north of the Arctic Circle.

Of the transitional categories, in subpolar glaciers the penetration of seasonal warmth is restricted to a relatively shallow surface layer, but is greater than in polar glaciers. The transitional subtemperate glacier is characterized by a relatively deep zone of annual warming. These subordinate terms are useful because the former may refer to transitional glaciers of dominantly polar character which still have certain temperate characteristics, and the latter to dominantly temperate glaciers having a tendency towards polar characteristics. The significance of such differentiation relates to the close control ice temperatures exert on flow deformation, and so also to glacial fluctuations. This includes a possible relationship to kinematic surges, which are described later. Each geophysical category (polar, subpolar, subtemperate, and temperate ice) can be found in any one glacier system if there is sufficient range in latitude or elevation for the requisite climatological factors. Present-day ice fields and ice sheets which are thermally temperate in some sectors and grade through to thermally polar geophysical conditions in other sectors are referred to as polythermal. Such probably was the geophysical character of the continental ice sheets during the waxing and

Fig. 3. Part of Vaughan Lewis Glacier, Juneau Icefield, Alaska, showing surface bulges and wave ogives in apron area just below icefall. Similar to view looking downvalley in Fig. 4. Series of medial moraines visible upper right. (*Photograph by M. M. Miller, August, 1968*)

waning phases of the Pleistocene Epoch. A glacier which is geophysically temperate (0°C) throughout is usually referred to as isothermal.

Thermodynamics. The internal thermodynamic and heat-transfer character of glaciers is more complex than suggested by the foregoing geophysical differentiations. This is because individual glaciers vary much in their structural makeup and related regime histories. Some of the complication is shown by the table of thermal constants for snow, firn, and ice at 0°C.

The problem of thermal changes within glaciers thus requires the approach of the physicist. One supplemental effect, however, must be kept in mind. This relates to the fact that, although normal thermodynamical processes pertain in temperate glaciers, infiltrating meltwater also plays an important role. In polar glaciers the fundamental diffusivity relationship shown in this table dominates, as there is only minimal meltwater effect restricted to the surface zone.

Irregularities in thermal dissipation occasioned by the presence of mobile water and by various physical inhomogeneities in glaciers usually preclude close quantitative agreement with hypothetical temperature curves calculated from assumed bulk diffusivity. Computed values are near to observed conditions only in purely polar glaciers, or in those sections of temperate glaciers in which during the winter months there is no liquid water to prevent changes in internal temperature from being controlled by conduction. The detailed mathematical analysis of the heat transfer is not considered here, beyond mention of the fundamental thermodynamic properties pertaining to understanding and appreciation of field observations.

In snow and firn, internal temperature changes may be considered as occurring in a semi-infinite, homogeneous, isotropic solid (although individual crystals are structurally anisotropic) influenced by fluctuating external temperatures which are a harmonic function of time. The basic physical factors controlling development of the sinusoidal cold wave are the thermal conductivity k of the medium, its specific heat c, and its density ρ. It can be shown that the transmission of heat actually depends on the diffusivity K, defined by a combination of these quantities in the relationship, $K = k/c\rho$. Here the specific heat is taken as constant at the value for ice. The density and conductivity, however, are variable, depending on the age of the medium and related genetic factors. The resulting diffusivity is therefore a function of these factors.

The relative diffusivity between dense firn and ice is not so large that mass heat transfer and temperature changes in firn are greatly altered by the

Thermal constants for snow, firn, and ice at 0°C*

Material	Conductivity, cal °C^{-1} cm^{-1} sec^{-1}	Specific heat, cal °C^{-1} g^{-1}	Density, g cm^{-3}	Thermal diffusivity, cm^2 sec^{-1}	Relative diffusivity to ice, approx ratio
New snow	0.0003	0.5	0.20	.0030	0.27
Old snow	0.0006	0.5	0.30	.0040	0.36
Average firn	0.0019	0.5	0.55	.0070	0.64
Firn ice	0.0038	0.5	0.75	.0100	0.91
Ice	0.0050	0.5	0.92	.0110	1
Water, 0°C	0.0014	1.0	1.00	.0014	0.13
Rubber	0.0005	0.40	0.92	.0014	0.13
Steel, mild	0.1100	0.12	7.85	.12	11
Aluminum	0.4800	0.21	2.70	.86	78
Copper	0.9300	0.09	8.94	1.14	104

*The given conductivities for snow and average firn are based on data from U.S. Army Corps of Engineers, Snow, Ice and Permafrost Research Establishment, *Review of the Properties of Snow and Ice*, rep. no. 4, 1951. The other data for snow and ice are from M. M. Miller, *Glaciothermal Studies on the Taku Glacier, Southeastern Alaska*, Tome 4, Publ. no. 39 de l'Association Internationale d'Hydrologie, Union Internationale de Geophysique et Geologie, Assemblée generale de Rome, 1954.

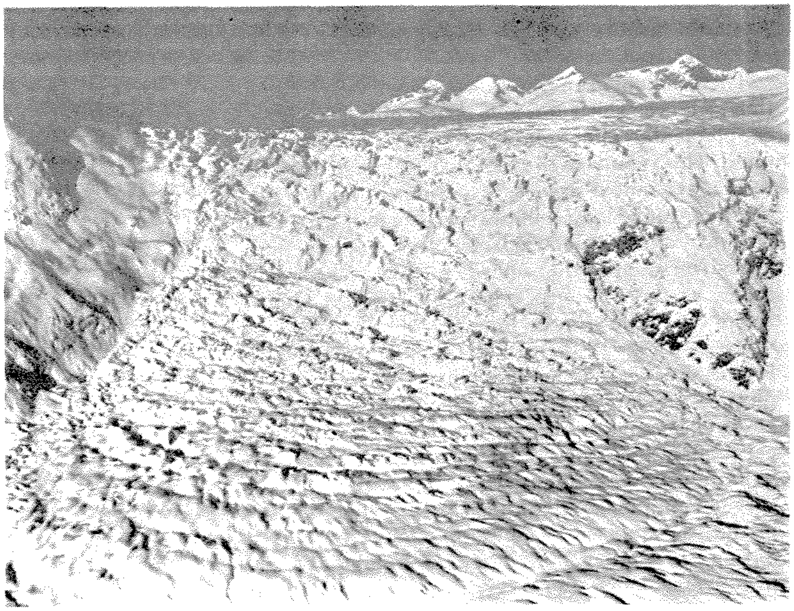

Fig. 4. Oblique air photograph of Upper Herbert (Camp 16) Glacier on the Juneau Icefield, Alaska, showing its plateau névé or accumulation zone at 5000-ft elevation, as well as the ice cascade, seracs, and wave bulges on the apron at the base of the icefall. (*U.S. Forest Service photograph, October, 1962*)

presence of a few ice strata. On the other hand, a glacier with a substantial covering of snow will display important differences in surface heat transfer compared to one with only solid ice exposed. The general relationships are illustrated in the table. The acute sensitivity of glacier flow to changes in internal temperature, as discussed under structure and movement below, underscores the importance of appreciating the role of these thermal constants in circumstances of pronounced and long-term climatic change.

Regime. Upon the regime, or the annual state of health of a glacier system, depends the glacier's eventual growth or decay, that is, its mass balance. The controlling factors are accumulation (gain) and ablation (loss) over the whole system. On temperate glaciers the critical sector is the névé line, or more properly a névé-line zone which at the end of the annual melt season separates the névé area (accumulator) from the bare-ice area

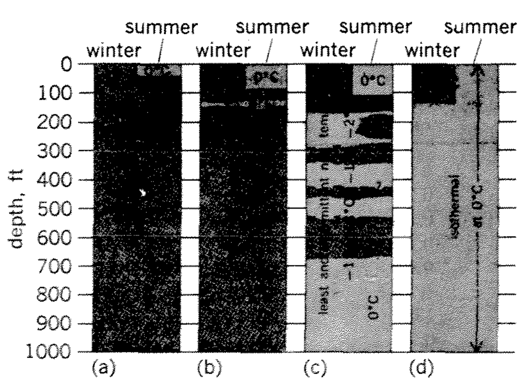

Fig. 5. Thermophysical classification of glaciers or ice sheets. (*a*) Polar. (*b*) Subpolar. (*c*) Subtemperate. (*d*) Temperate. Arbitrary depth-temperature values at surface illustrate seasonal variations. Dark tone, "cold" subfreezing state; light tone, "warm" (0°).

(dissipator). As already noted, it equates to the lower elevation limit of retained winter snowpack as observed at the end of summer. On polar glaciers the equivalent critical sector is, of course, the equilibrium line.

The height of the seasonal névé line, or yearly equilibrium line, shifts greatly according to the regional climatological situation and local conditions. With respect to the Juneau Icefield in southeastern Alaska, this critical line (Fig. 6) varied over a 20-year period between 2400 and 3800 ft (mean 3100 ft). In high-latitude polar regions the equilibrium line is usually at, or close to, sea level. On alpine glacier systems a higher than average névé line means a tendency toward a negative regime (more loss than gain), whereas a lower than average névé line is associated with an increase in the accumulation area and hence a positive regime. This kind of interrelationship is illustrated by the accumulation trend in Fig. 6 and by the typical mass-balance relationship depicted for a healthy glacier situation in Fig. 7. The specific net budget curve in Fig. 7 shows the difference between accumulation and ablation per unit area at every specific elevation. To obtain the actual or total budget (mass balance), this figure is multiplied by the total glacier area at that specific elevation. In alpine glaciers the area ratio of accumulator to dissipator is on the order of 4:1, whereas in the Greenland ice sheet it is in excess of 100:1, and in the Antarctic as much as 1000:1. Also, in the accumulation zone there is an increase of retained accumulation in any one year with an increase in elevation until the height of maximum snowfall is attained. It should be noted that this is not always at the highest elevation of land, because maximum snowfall is controlled by the mean freezing level, that is, the greatest snowfall occurs at or near 0°C. Thus, in mountain regions or on large ice sheets

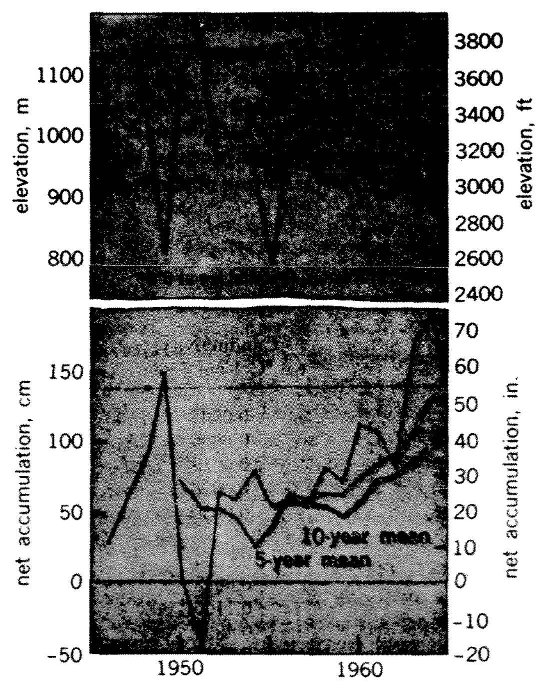

Fig. 6. Comparative neve line and net accumulation trends on the Taku Glacier, Alaska, during 1946–1965. (*Juneau Icefield Research Program*)

with a great elevation range at very high elevations, there is commonly less snowfall, the conditions there being generally too cold for much snow.

The rise and fall of the névé line over a period of years also parallels vertical shifts in the zone of maximum accumulation. This vertical shift may be measured by means of soundings, borings, test-pit studies, and observations on crevasse walls. In a healthy glacier the average level of maximum snowfall lies at the elevation of greatest glacier area (Fig. 7). A glacier in a poor state of health has usually experienced a rise in its level of maximum snowfall to a height above the area of the main névé. Also to be considered as an additional increment of positive net accumulation in a glacier system is that portion of summer meltwater in geophysically subpolar or polar (cold) névés recaptured by freezing as it percolates to depth. In contrast, meltwater percolation in temperate glaciers drains away almost entirely in subglacial drainage channels, eventually flowing out at the snout of the glacier as an increment of net loss.

Structure and movement. Because of the many structures in ice comparable to those in sedimentary and metamorphic rocks, a glacier is an ideal field laboratory for the structural geologist. A few such structures are: primary stratification (bedding strata), secondary fracture structures, discontinuous marginal or basal tectonic foliation, (Fig. 8), ablation surfaces overthrust surfaces, faults, folded structures, and a varied group of deformed sedimentary and structural bands, including the wave-ogive bands and surface bulges illustrated in Figs. 3 and 4. There are also subsurface diagenetic structures of both stratiform and transverse ice which result from refreezing of downward percolating meltwater in subfreezing firn. Cross-cutting transverse types are manifest at the surface as ice columns and dikes. Tension and shear fractures are also common as crevasses and bergschrunds (the latter exhibiting an overhanging upper lip), as well as moulins (glacier mills, or deep rounded holes caused by water action on embedded stones), cryoconite holes (thermal pits produced by the inmelting of organic or rock fragments), sastrugi (wind-scoured features), and other surface features. All of these are representations of the combination of processes which affect and control the surface regime and the dynamics of the internal structure and movement of glaciers.

Glacier deformation is a composite of internal and external movement. The internal movement is dominated by a continuous plastic creep (flow deformation), and the external in some cases by a fracture type of discontinuous movement at the bed or near the glacier margins. As a general rule, flow deformation, hence the rate of movement, is greater in the upper portion of a glacier than in its basal section (Fig. 9). In glaciers with a healthy or strongly positive regime in the accumulation zone, a dominant proportion of the total mass transfer may be expected to be via actual sliding of the glacier over its bed. Such glaciers are characterized by rectilinear, pluglike, or Block-Schollen velocity profiles. In plan view the greatest amount of movement is expressed in a broad central area. The surface velocity profiles in Fig. 10 were surveyed at stated intervals up-glacier from the terminus and represent the general flow characteristics of this strongly advancing glacier during the 1950s.

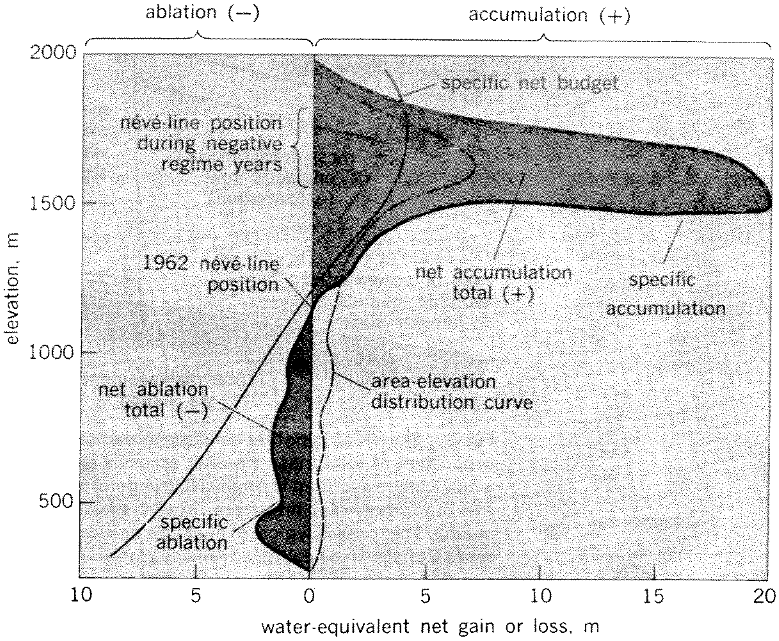

Fig. 7. Mass balance relationship on Nigardsbreen Glacier, Norway, in 1961–1962 budget year. This glacier normally has a negative regime, but it was surplus in this year, calculated to be about 95,000,000 m³ water equivalent. (*After G. Ostrem, Nigardsbreen Hydrologi, 1962, Norsk. Geog. Tidskr., vol. 18, pp. 156–202, 1963*)

Sliding glaciers often contain sheared basal ice heavily entrained with rock fragments and serving as effective agents of erosion. In glaciers of equilibrium or negative regime in the névé, the dominant mode of movement is expressed as internal streaming flow. In plan view such velocity profiles have a smooth, arched, parabolic form with associated streamlines. All gradations of movement from continuous laminar (streaming) flow to discontinuous Block-Schollen, or plug flow, may occur in any one glacier system. The relative proportions depend upon the névé regime pattern in recent decades, on internal temperature characteristics of the ice, and on configuration of the bedrock channel.

During the 1950s laboratory and field experiments by M. Perutz, J. Glen, M. Miller, J. Nye, S. Steinemann, and others ushered in a new set of concepts in glaciology, which related rheological and mathematical models of glacier mechanics (deformation and flow) to the phenomena of glacier movement and mass transfer observed in nature. The field measurements were largely based on the deformation of glacier tunnels and elongated metal

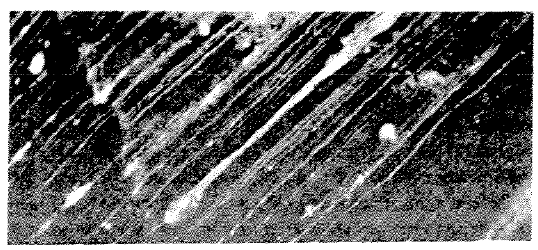

Fig. 8. Basal tectonic foliation (discontinuous laminar shear bands) with entrained fragments of pebble-size rock and fines in sole of Moltke Glacier, a geophysically polar glacier east of Thule. (*Photograph by M. M. Miller*)

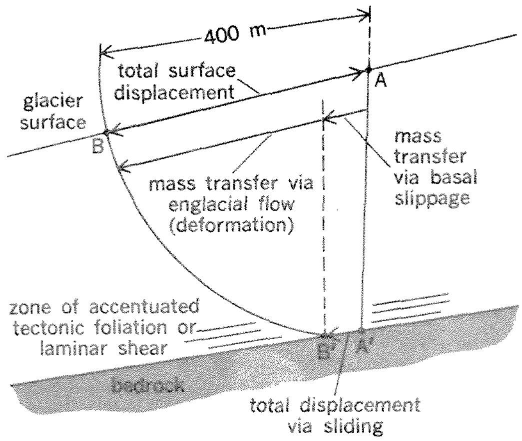

Fig. 9. Sketch of hypothetical case to demonstrate the proportion of total mass transfer across a given profile which can be ascribed to englacial flow deformation plus the proportion resulting from erosive slippage or bed sliding. This case is typical of plug flow or Block-Schollen mass transfer in a healthy advancing glacier.

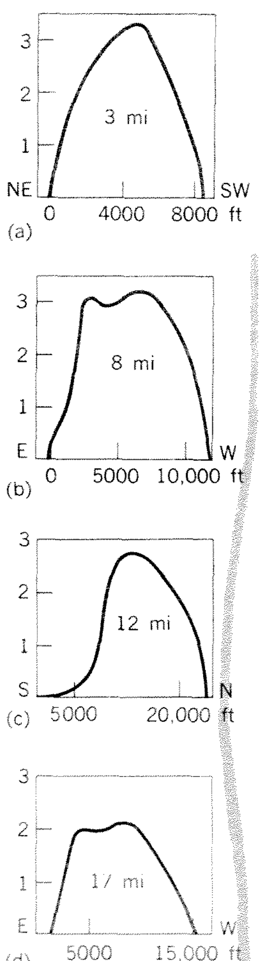

Fig. 10. Horizontal across glacier surface velocity profiles in ft/day from terminus up into névé zone of Taku Glacier, Alaska. (*a, c*) Partial parabolic streaming. (*b, d*) Block-Schollen (plug flow).

pipes drilled into the glacier perpendicular to its surface. As a result of this pioneering work, the nature of internal deformation of glaciers is treated as a problem in the physics of shear, with reference to plasticity theory. Thus it is expressed by an exponential relationship between gravitational stress and deformation per unit time (creep velocity) with a simplified power law equation, Eq. (1).

$$\frac{d\gamma}{dt} = k\tau^n \tag{1}$$

Here γ is the shear strain, τ is the shear stress in bars, k is a constant for any given temperature, and n is an empirical constant, depending in large measure on the physical character of the ice. Also, the exponent n probably depends to some degree on the magnitude of stress, a factor judged as probably significant only in very deep or otherwise highly stressed ice.

Under temperate valley glacier conditions, the factor n in nature is found to be close to a value of 3. In the flow law n represents the slope of the double logarithmic plot of the fundamental shear stress – shear strain relationship. The constant k is obtained from the stress-strain formula at the stress of 1 bar. Since the logarithm of 1 is zero, k is, in fact, the ordinate at the point where the abscissa of any log stress – log strain rate line is zero.

Also $d\gamma/dt$ or $\dot{\gamma}$ is the strain rate per year, usually calculated in radians of angular deformation per year, because it is found to be numerically equivalent to the tangent of the changing tilt angle in slowly deforming englacial pipes from which vertical velocity profiles have been measured. Several examples of field measurements are given in Fig. 11.

It should be noted that the power law used with a universal constant n represents only an average for low-gradient temperate glaciers, and therefore can give no more than a good approximation of the true strain rate in a glacier at depth. Nonetheless, by application of this law, at least for glaciers of simple configuration, a reasonable determination can be made of the movement within a glacier. And by combining the englacial movement data with surface velocities obtained by periodic surveys of across-glacier stakes and with geophysically determined depth records, an assessment can be made of the relative proportion of sliding or slippage on the bed (Fig. 9).

The shear stress at any depth with a glacier is calculated from relation (2) in dynes cm^{-2}, where D is

$$\tau = D\rho g \sin\alpha \tag{2}$$

the depth in centimeters, ρ the bulk specific gravity of the overlying mass in grams per cubic centimeter, α the surface gradient, and g the acceleration of gravity (980 cm sec^{-2}). In relation (2) the very critical control exercised by slight variations in gradient is well revealed.

Thus in the flow law with all factors determinable in nature, and with n and k calculated for any particular velocity profile, it is possible to extrapolate the differential flow all the way to the base of a glacier. In this way determination is made of the proportion of total mass transfer within a glacier. For this calculation the difference is found between the quasi-plastic flow velocity at any given depth U_D and the surface velocity U_0 at the glacier surface. The relation is expressed by combining the empirical constants k and n with the stress at the indicated depth D as in Eq. (3), which was formulated by Nye.

$$U_0 - U_D = \frac{k}{n+1} D\tau^n \tag{3}$$

In this manner the approximate value of basal sliding is determined. A hypothetical case is shown in Fig. 9. The actual extent to which bottom sliding takes place depends not only on the surface gradient but on the slope and roughness of the underlying topography, on the width and thickness of the glacier, and on any other blocking factors involved. All these factors vary in their effect on the total stress distribution in different sectors of a glacier, and they are expressed at any particular point in the ice by the creep law, which integrates the gravitational load and englacial temperature. It must be kept in mind that such analyses oversimplify the situation in nature, because ice is not actually a homogeneous plastic material. Instead it should be considered as a quasi-elastoplastic substance, whose behavior under stress is still not fully understood.

In glacier ice when the stress conditions are great and substantial bottom sliding takes place, there is also much marginal shearing along the edges of the glacier. This is suggested by the form of transverse surface-velocity curves (Fig. 10) and the field measurements of vertical velocity profiles which have been noted (Fig. 11). The accentuation of laminar shearing or tectonic foliation in the sole of a glacier, observed as ribbon structures along the glacier's margins and bed (Fig. 8), indicates that slippage more usually takes place in a zone of highly sheared basal ice carrying entrained rock fragments and debris, which in turn serve as the key erosive tools of the glacier.

Erosion and transportation. Like rivers, glaciers have distinct regions of erosion and transportation. For example, in the high cirque basins of alpine glaciers, rock fragments fall from cliffs or slide into bergschrunds or marginal moats and become entrained in the ice for transport down the valley. These transported fragments thus become

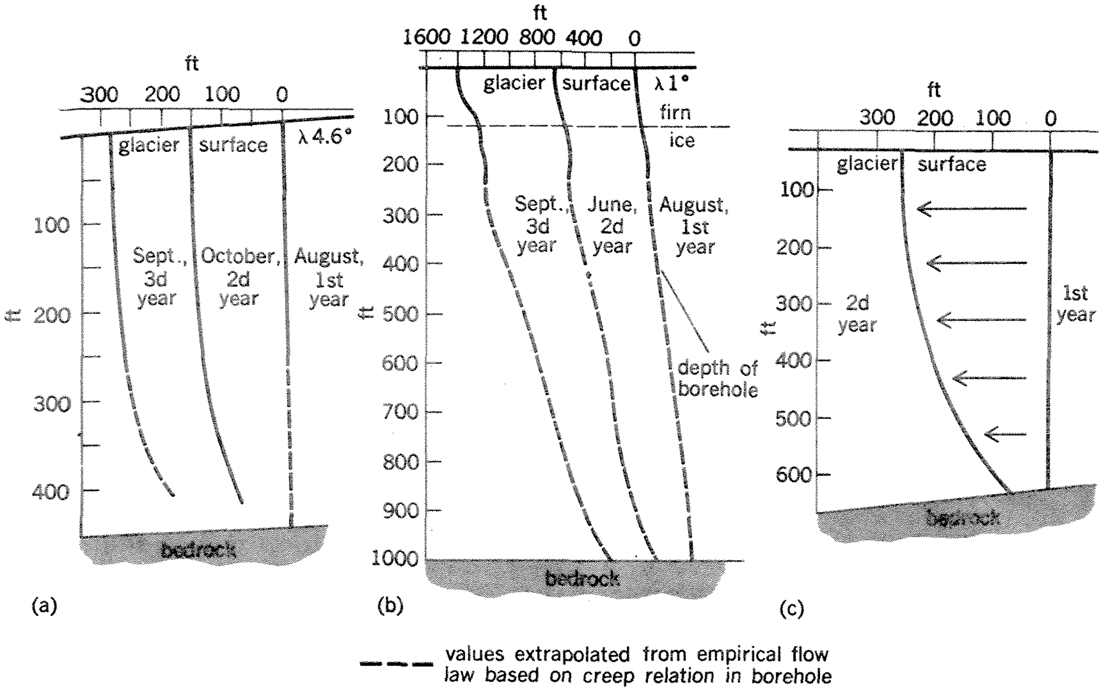

Fig. 11. Vertical velocity profiles, based on measurements made in the field, showing annual changes on the (*a*) Aletsch Glacier, Switzerland (*after M. Perutz*); (*b*) Taku Glacier, Alaska (*after M. Miller*); and (*c*) Saskatchewan Glacier, Canada (*after M. Meier*).

the critical tools of erosion along the ice-bedrock interface (Fig. 8). The larger angular rocks scratch the bedrock and form grooves and striae (Fig. 14), while the finer clastic detritus serves as the abrasive to smooth and polish. In such areas the presence of impact structures, striae, grooves, and crescentic features are not only proof of former glaciation but also reveal the former direction of flow.

Other geomorphological characteristics support the role of bottom slippage in erosion, such as deeply entrenched and U-shaped valleys reflecting the parabolic form of englacial creep in an actively advancing glacier, and the vast quantity of rock flour (clay size) carried out from beneath many glaciers by subglacial and proglacial drainage streams. Neither of these conditions could possibly have developed without considerable abrasion from basal sliding of a debris-entrained sole of the glacier.

Plucking is also an important and related agent of erosion. This process involves the penetration of ice or rock wedges into subglacial niches, crevices, and joints in the bedrock. As the glacier moves, it plucks off pieces of jointed rock and incorporates them as supplemental agents of abrasion and further plucking. Down-valley ends of jointed hummocks in the bedrock are produced in this manner and are known as roches moutonnées. Figure 14 shows the glacial grooves and oriented crescentic erosional features on top of a roche moutonnée. A sequence of such bedrock bosses produces a steplike longitudinal profile on a glacial valley floor. The steps often coincide with particularly resistant lithologies or selective low-angled joint surfaces.

Surging glaciers and kinematic waves. Research has suggested that glacier surges, expressed as sudden catastrophic advances or raising and lowering of the ice surface, are relatively common phenomena in some regions. Such abnormal surges are characterized by marked and seemingly anomalous increases in flow velocity and often by a rapid transfer of ice from the névé to the terminus, with much increased pinnacling and crevassing. The upper glacier surface may sink or be substantially lowered with this volume loss, expressed as a comparable thickening in the lower valley sector. Such surges are believed to be kinematic in nature, in that the wave moves through the glacier at a substantially faster rate than the actual discharge of ice. Often much crevassing and geometric folding of in-ice structures and medial moraines occur, as well as strong buckling, shearing, and surface lowering along the

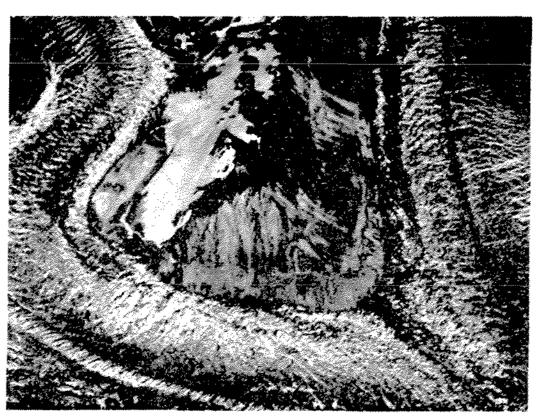

Fig. 12. Massive marginal shearing and lowering of ice along the valley walls of surging Walsh Glacier, in the Yukon Territory of Canada. Note the strongly related development of conjugate shear crevasses. This glacier moves forward tens of feet per day. (*Photograph by H. B. Washburn, August, 1966*)

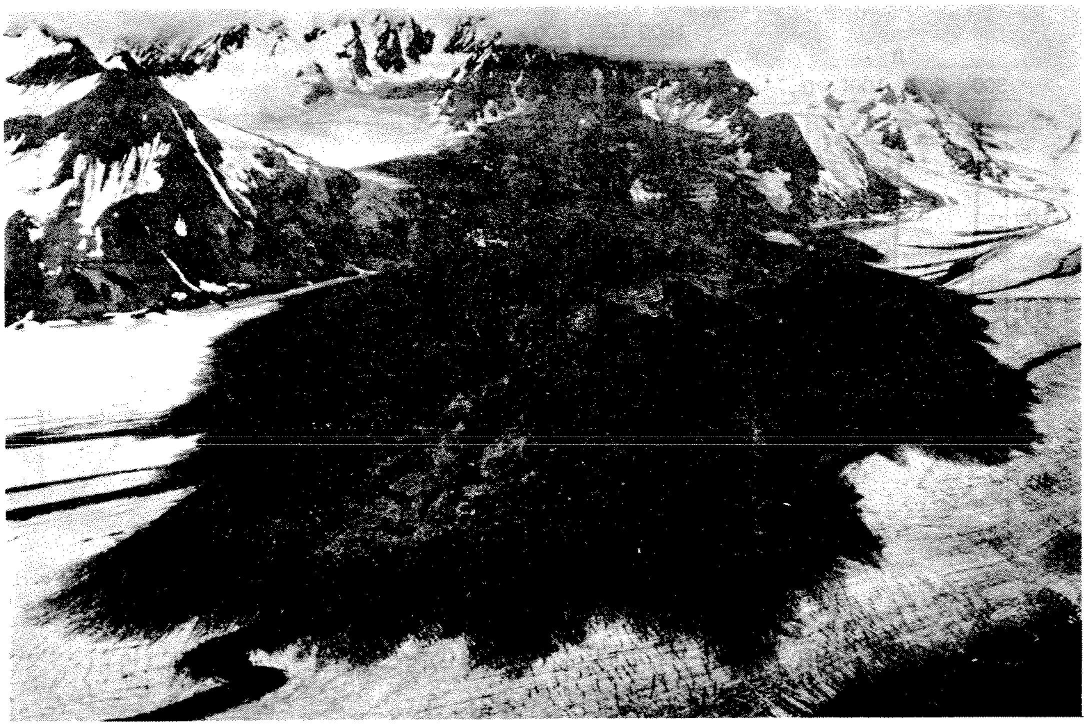

Fig. 13. Massive earthquake avalanche of rocks and debris covering approximately 10 mi^2 of the Schwann Glacier in the Eastern Chugach Range, Alaska, was caused by the 1964 Good Friday earthquake. *(Photograph by M. M. Miller, National Geographic Society, Sept. 15, 1964)*

valley walls (Fig. 12). Striking examples are the Dusty, Lowell, Walsh, and Steele glaciers in the Yukon, where in 1965–1969 surge velocities of up to 60 ft per day were reported.

Surging glaciers were first reported by R. F. Tarr and L. Martin in Yakutat Bay, Alaska, in 1910–1913, with later reports of vigorous surging in this same area in 1965–1968. Surges of spectacular nature have been reported on many glaciers, including the Black Rapids Glacier, Alaska, in 1937; Bruarjokul Glacier, Iceland, in 1965; Medvezhii Glacier in the Pamirs, Soviet Union, in 1963; and Muldrow Glacier on Mt. McKinley, Alaska, in 1964. Others have been reported in Spitsbergen in 1952, Ellesmere Land in 1964, and even in the Karakoram and Himalayas in 1953 and 1963.

Although this phenomenon is not adequately understood, much research is under way. It is at least known to relate to development of a dynamic instability that usually attenuates within 1 to 5 years, after which the lower glacier area affected by such an advance begins to stagnate. It is not the same phenomenon producing normal discharge wave-bulges and advance, as these may be presumed to relate to annual accumulation variations and to gradual climatic change. One possible origin of the catastrophic surge phenomenon is a very unusual and sudden change in load stress via thickening of the glacier in its upper névés or high-level nourishment basins, with a consequent energy release via a kinematic wave.

Of course such thickening may be associated with abnormal increases of snowfall by way of what appears to be sudden climatic change. In certain unique situations, the suddenness of such change can be accentuated by significant horizontal or vertical shifts in the zone of maximum snowfall across the critical névé level. But there are several other possibilities. One may relate to warming of the englacial ice itself, a factor which probably caused the Brasvalsbreen Glacier in Spitsbergen to slide forward catastrophically in the late 1940s (a 5-mi advance in less than 5 months). This may not have been a true surge but a response to substantial changes in englacial temperature, since it is known that the mean winter temperatures in the northeastern Arctic rose about 20°F during 1920–1960.

The third possibility is the sudden imbalance produced by earthquake-caused rock and ice slides. A number of huge slides were reported on certain Alaskan glaciers following the 1964 Good Friday earthquake. One of these is pictured in Fig. 13, showing apparent flow effects in a deformed medial moraine below a 10-mi^2 debris slide on Schwann Glacier in the Copper River region, Alaska. A fourth, and probably minor, possibility is some actual effect of the earthquake shock in the disarticulation of the glacier structure, which could abet the "sudden-slip" character of a subsequent advance. This would probably be true when the epicenter of the earthquake lies in the vicinity of the glaciers involved. A further possibility is a buildup of excessive pressure melting on a glacier's bed, or in some cases an intensification of geothermal heat at the sole of the glacier, resulting in abnormal quantities of lubricating water (slush) which could accentuate the effects of basal slip. In some cases the most dynamic surges could be a result of a combination of two or more of these factors. However, glaciologists must remain cautious, because many glaciers surge and even express seemingly less kinematic effects, such as surface buckling and locally increased crevassing, without any evidence of avalanche material being dumped on them. At least one correlation which is of interest is that some of the most spectacular surging glaciers have been reported in regions of the most active tectonic-earthquake activity.

The greatest bed erosion takes place when the regime of a glacier is healthy and its mass transfer substantial. A vigorously advancing glacier exhibiting Block-Schollen or plug flow is the most effective erosive body. This adds credence to the concept that glacial erosion is an indirect consequence of glacioclimatological oscillations affecting the growth and decay, thickening and thinning, and advance and retreat of glaciers.

As to the direct cause of high ratios of bed slippage and the occasional development of anomalous surges, scrutiny of the shear stress equation reveals that flow within a low gradient glacier or ice sheet (for example, at slopes of 1 – 3°) is so small that in many cases stresses other than those explained by pure gravitational shear must be in effect. Hydrostatic pressure is ruled out on the basis that it is as negligible in ice as in liquids. The most significant supplemental stress may be attributed to a strong down-glacier longitudinal force superimposed on the gravitational stress in quasi-elastoplastic ice whose yield limit is already exceeded. Such supplemental stressing can be the result of any one factor or a combination of the factors which come into play in advancing and surging glaciers. To such stress tensors, the strain effects of which are sometimes accentuated by an increase in temperature within the ice, the phenomenal glacial advances in Alaska (group IV, Fig. 15) may be ascribed.

The repeated oscillation of glaciers in mountain cirques and over the floors of outlet valleys results in effective scouring, transportation, and removal of material. Such continuous sequences of process produce the wide-strath highland glacial basins and deep U-shaped outlet valleys so common in the Alps, the Cascades, and the Rocky Mountains. Turbid glacier rivers and moraines (lateral, medial, and terminal) are living examples of the immense eroding and transporting power of glaciers. Festooned arrays of terminal moraines also testify to continuous glacier activity, because they express repeated oscillations most usually related to cyclic changes of accumulation and ablation in the distant névés many years before.

Recent and current fluctuations. Following the post-Wisconsinan maximum about 10,000 years ago, the Ice Age entered its latest waning phase. The result was an almost complete retreat and disappearance of glaciers in the mid-latitudes and tropical regions of the Earth. Coincident with this disappearance was the Thermal Maximum (Hypsithermal Interval or Climatic Optimum) culminating about 5000 years ago. Between 1000 B.C. and A.D. 1 a worldwide recrudescence of glaciers began to take place. This is termed the Neoglaciation and is associated with a return to harsher climatological conditions, which still prevail today. This temperature fluctuation culminated in a colder condition at the beginning of the Christian era. The evidence suggests that in the 5th and 7th centuries the polar seas were freer of ice than they are today, as far north as the pole. Peripheral waters thus remained relatively clear through the 10th century. The records of Norse settlers in Greenland indicate that the climate remained relatively mild until the 14th century. Then, about 5 centuries ago, there began another worldwide expansion of glaciers and a thickening of polar ice, generally referred to as the Little Ice Age. Technically, the Little Ice Age refers to this latest major phase of reglaciation in the latter quarter of the Neoglacial Age.

Fig. 14. Glacial pavement on grantic bedrock surface in recently deglaciated mountain region of Alaska, showing orientated abrasion structures resulting from basal sliding of the glacier over its bed. In this view can be seen crescentic gouges (concave in up-glacier direction), lunate furrows (concave in down-glacier direction), grooves (elongate linear furrows), and striae (fine lines of scratches). The direction of the former ice movement is from foreground toward valley in distance. (*Foundation for Glacier and Environmental Research*)

Little Ice Age fluctuations are double-phased in nature, having produced a worldwide growth of temperate glaciers which reached their culminations in the early to mid-18th century, and again in the late 19th to mid-20th centuries. The Alaska Little Ice Age pattern is illustrated in Fig. 15. This reveals that at any one time the terminal regime of glaciers in adjoining areas may be quite out of phase; that is, one may advance while another simultaneously retreats. The latest advances on a small percentage of high-level trunk glaciers have continued into the 20th century, in spite of a general diminution of ice cover around the periphery of some of the lower Alaskan ice fields. Recent fluctuations in Scandinavia and Patagonia are quite similar to this Little Ice Age pattern and have been shown to have a teleconnectional similarity via upwards of a dozen recessional moraines over the past 200 years. Such evidence supports the global nature of the causal factor, and the acute sensitivity of glaciers as excellent historians of secular climatic change.

In contrast to the behavior of temperate glaciers, the polar glaciers of Antarctica exhibit a

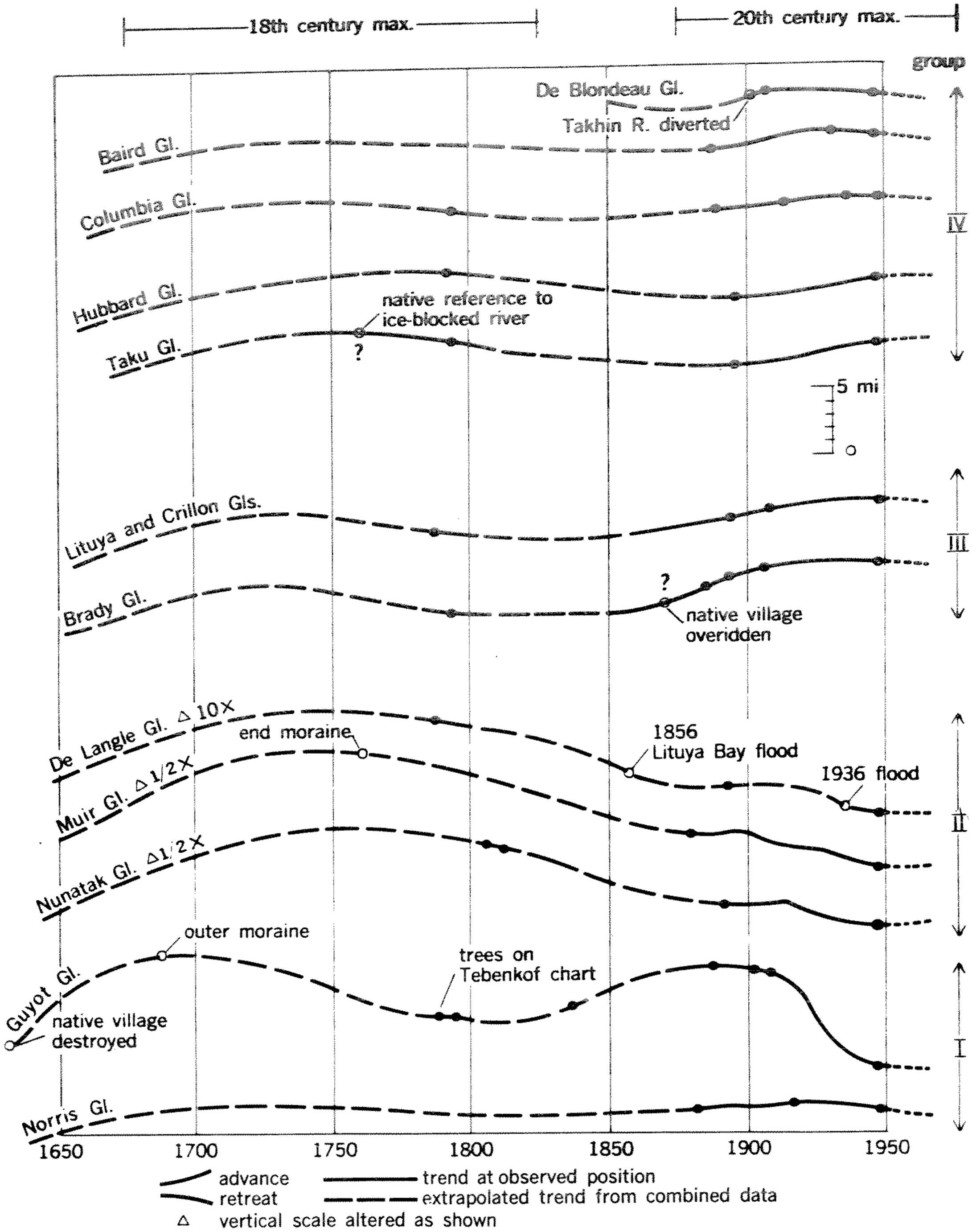

Fig. 15. Two-phased regional fluctuation pattern of the Alaskan Little Ice Age showing how, at any one time, glaciers in a region may reveal patterns of strong different advance simultaneous with significant retreat of others. This out-of-phase pattern reflects the role of differences in size, geographical position, and elevation or orientation of névés, as well as differences in flow lag via differences in size, configuration, and length of the outlet valley glaciers.

fairly stable regime. This suggests that significant volume changes in middle-latitude glaciers, including the sub-Arctic and the sub-Antarctic, and noticeable increases in their internal temperatures have been instrumental factors in accentuating the fluctuation patterns of Neoglacial time. *See* GLACIAL EPOCH; GLACIATED TERRAIN.

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