



Glacier Geophysics

Barclay Kamb

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Glacier Geophysics

Dynamic response of glaciers to changing climate may shed light on processes in the earth's interior.

Barclay Kamb

In ordinary experience ice is a strong, rigid substance, and to casual observation glaciers appear to be solidly permanent features of the landscapes where they occur. The finding that these great masses of ice are actually in motion, flowing like fluids, has long attracted attention and has provoked much scientific controversy (1). Only within the past 15 years or so have we arrived at an understanding of how and why glaciers flow, through concepts of solid-state physics and of the new materials sciences and through development of new and better means of making physical measurements on glaciers.

Ten percent of the earth's land area is at present covered by ice; during the recent ice ages the ice-covered area was almost three times as large. On at least four occasions ice invaded most of Canada, the northern part of the United States, northern Europe, and many mountain regions of the world. There is no general agreement yet on the cause of this phenomenon (2). In the search for an explanation, glacier geophysics aims to provide a quantitative connection between climatic change and glacier fluctuation, so that glaciers can be interpreted confidently as long-term

monitors of climate. To provide such a connection requires an intimate understanding of the dynamics of glacier motion. It has only recently been realized, for example, that there are inherent instabilities in glacier response, so that a small climatic change may produce a large glacial advance or retreat.

The flow of glaciers also serves as a reminder that, on a time scale of millions of years, "solid" rocks themselves flow like fluids, in the complex and little understood processes by which the earth builds great mountain ranges and other structures of continental dimensions, the processes of tectonophysics. In this similarity, glacier geophysics and tectonophysics have an important meeting ground. Glaciers constitute great outdoor laboratories in which concepts and theories derived from indoor laboratory experimentation can be tested on time scales and distance scales more nearly appropriate to the phenomena of solid-earth deformation, and yet still accessible to human observation and measurement. Flow in glaciers produces striking internal structures (see cover) that are analogs of structures in certain metamorphic rocks from which great deformations of the earth's crust have been inferred. By studying how and why these structures originate in glaciers we can hope to get a better under-

standing of deep and fundamental earth processes that will probably forever be hidden from direct investigation.

This article sketches recent ideas and measurements bearing on the glacier flow process, its expression in dynamic response of glaciers, and the possibility of using these concepts to increase our understanding of solid deformation processes in the earth generally.

Glacier Flow

Flow velocities of most valley glaciers are in the range of 0.1 to 2 meters per day, and are in general an increasing function of glacier size and valley slope. In ice falls, like those seen in the cover photograph, velocities of up to about 6 meters per day occur commonly. Exceptional motions of as much as 30 meters per day have been reported for the great glaciers flowing from the ice sheet into fjords on the west coast of Greenland, such as Jakobs-havn.

These velocities are actual downslope ice motions under gravity. The apparent advancing or retreating movement of the glacier snout (terminus) represents, of course, a balance between ice wastage (ablation) and forward motion, and is usually much slower than the ice motion itself. A striking exception is the movement of certain glaciers, such as the famous Black Rapids Glacier of Alaska, which at times advance suddenly and catastrophically at speeds of up to 50 or even 100 meters per day (3).

Although velocity measurements are made by long-established methods of precise surveying, only recently have the motions of individual glaciers been measured in sufficient detail to define completely the velocity field at the surface of the ice. Data for a portion of Saskatchewan Glacier (Canada), one of the most completely studied so far (4), are shown in Fig. 1. Figure 1b

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shows strain rates calculated from the velocity data in Fig. 1a and demonstrates the close connection between fracturing of the ice (crevassing) and the strain pattern, crevasses tending to open up perpendicular to the directions of greatest extension of the ice.

Of importance for understanding gla-

acier flow are the velocities at depth in the ice and the internal deformations taking place. To measure these a hole is drilled through the glacier and a pipe is placed in it; subsequent tilting of the pipe as a result of glacier deformation is measured by instruments lowered into the hole (5). The velocity profiles in

Fig. 2 are the best data of this kind so far published. The vertical component of velocity cannot be measured by this method, so the results are not completely three-dimensional. By planting ultrasonic transmitter-detector devices in a network inside the glacier it is possible by "three-dimensional trilatera-

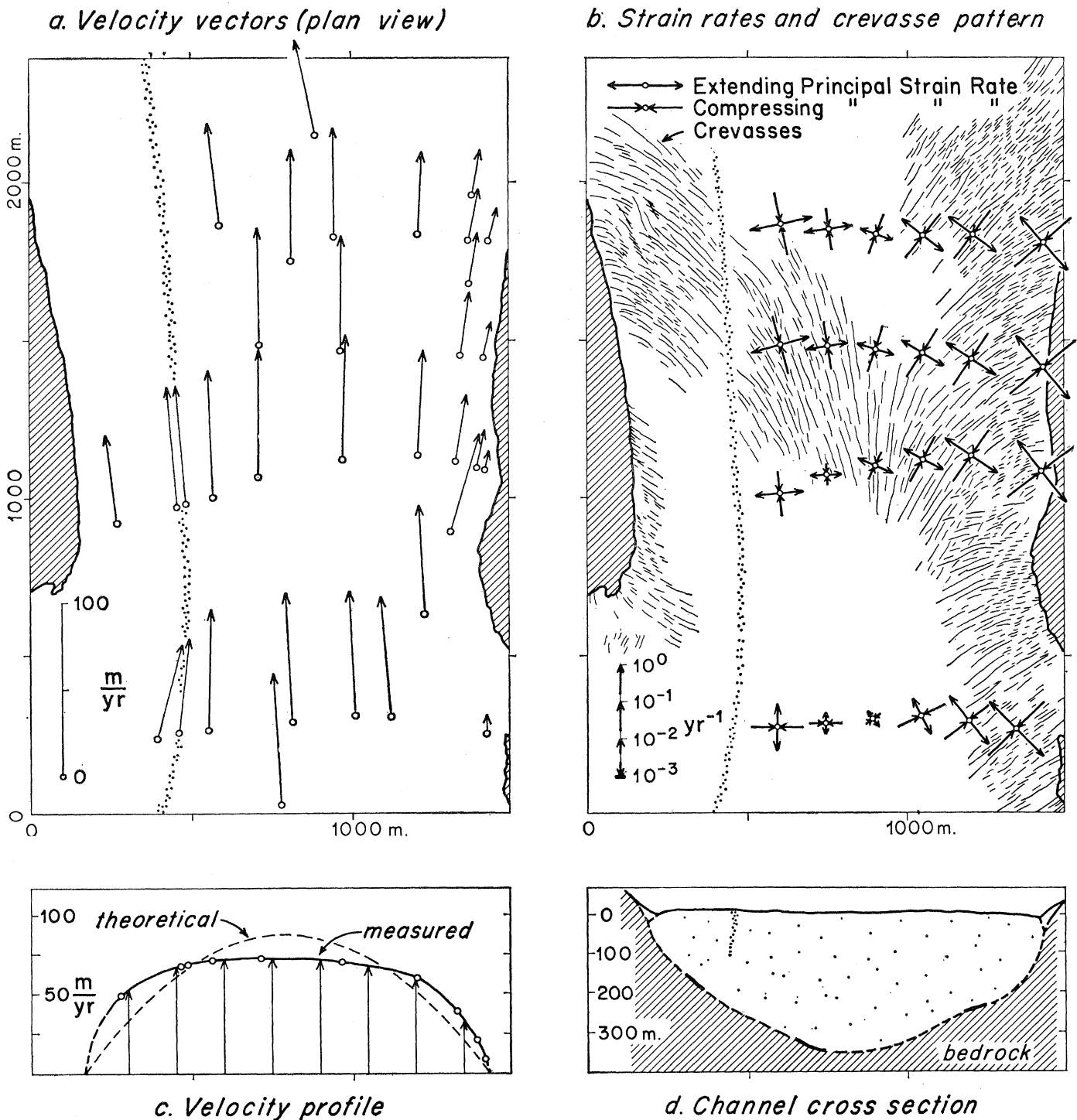


Fig. 1. Flow velocity and strain-rate data from a portion of Saskatchewan Glacier, Canada (4). (a) Measured flow velocities at the surface of the glacier. (b) The corresponding strain field, shown by plotting the principal strain-rate magnitudes and directions (directions of maximum extension or compression); the magnitudes are plotted on a logarithmic scale (note that the smallest strain rate that can be represented on the scale used is 0.001 per year); the crevasse pattern on the glacier surface is superimposed for comparison. (c) Surface velocity profile across the glacier at the 1000-meter coordinate. (d) The corresponding channel cross section as determined by seismic measurements. The theoretical curve in c is calculated from an elliptical cross section approximating d, for flow of a viscous fluid of viscosity 0.8×10^{14} poise.

tion" to measure the complete velocity field, but the method is so new that results comparable to those of Fig. 2 have not yet been obtained (6).

Drilling is also a way of determining the bottom topography or channel configuration of the glacier, which is essential to any dynamical analysis. Seismic methods (echo sounding) are much more efficient and have been used in connection with the best, most detailed studies of glacier motion. Gravity measurements can also be used (7).

As a problem in hydrodynamics, glacier flow ought to be simple, because the velocities are so slow that inertial forces are negligible in comparison to the gravity and "viscous" forces operating. An analysis in terms of slow flow of a viscous fluid is unsuccessful, however. It predicts, for example, a parabolic velocity profile (Fig. 1c) across glaciers flowing in channels of elliptical cross section, like the typical U-shaped glacial valleys (Fig. 1d). Instead, the actual velocity is nearly constant over a wide central zone and the drop-off in velocity is concentrated toward the sides, forming a kind of "boundary layer" (Figs. 1c and 2).

The failure of a viscous theory is not surprising, in view of the fact that ice is truly a solid. The brittle type of solid behavior represented by the crevassing so prominent in glaciers is, however, confined to a relatively thin layer (10 to 30 m thick) near the surface; as in other solids, confining pressure inhibits brittle fracture and allows the inherent plasticity of the material to manifest itself. Modern understanding of glacier flow thus begins with the recognition that, since it is a solid, glacier ice ought to behave in accordance with the theory of plasticity (8) rather than with that of viscosity.

The "perfect" material envisioned by the theory of plasticity remains undeformed up to a yield stress τ_0 , at which it is capable of limitless solid deformation. A "perfectly plastic" glacier would be able to flow only if the shear stress at the base of the ice, resisting its downward motion, were everywhere equal to τ_0 . The shear stresses at the bottom of glaciers are in fact roughly constant, being almost always in the range 0.5 to 1.5 kg/cm² in spite of the fact that climatic and topographic conditions vary greatly among glaciers (8). Hence the yield stress can be identified as $\tau_0 \approx 1.0 \text{ kg/cm}^2$.

The basal shear stresses have not, however, been measured directly, and in fact no direct measurement of an

internal stress in a glacier has yet been made. The difficulties involved in making a measurement of this kind become apparent if one tries to think how it might be done. The stresses cited are averages estimated from the requirement of gross mechanical equilibrium—balance of the gravity force and the resisting stress integrated over the glacier cross section.

Properties of Ice

Idealized assumptions about ice properties are subject to experimental test, and the attractiveness of glaciers for physical study is due in no small part to the relative simplicity of their material (nearly pure solid H₂O) and the relative ease with which it can be studied experimentally.

Most of the properties of glacier ice are, ultimately, manifestations of the structure of the individual ice crystal. Single crystals of ice, while brittle by ordinary standards, are remarkably plastic (9, 10) when tested in the laboratory (Fig. 3). The deformation takes place through the gliding over one another of layers parallel to the "basal" plane (plane of the snowflake). Although the structure of the crystal restricts the gliding rigorously to this plane, it appears to exert no control on the gliding within the plane, so that the layers can slide readily over one another in any direction, as though the planes themselves were structurally amorphous. This peculiar behavior is quite unlike that of "normal" plastic crystals (11).

The plastic property is in fact not plasticity in the strict sense (8), but is instead a type of creep, in which, after an initial transient response, the deformation under fixed applied stress proceeds at a steady rate (steady-state creep). The response to stress is markedly nonlinear and can be represented by the empirical equation

$$\dot{\gamma} = a\tau_s^n$$

where $\dot{\gamma}$ is shear strain rate, τ_s is shear stress across the basal plane, and a and n are empirical constants with $n \approx 3$. This type of creep is also found for metals at high temperature (12) and has been explained theoretically in terms of the motion of crystal dislocations (13). Although the dislocations themselves are under strict crystallographic control, their combined motion turns out to produce for $n = 3$ exactly the observed rotational symme-

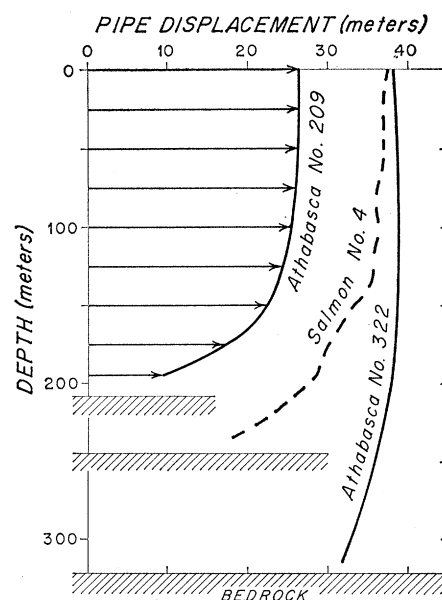


Fig. 2. Internal deformation as measured by tilting of pipes placed in glaciers (5). The displacement of each pipe from its original vertical position in 1 year is plotted, and the resulting curve shows directly the profile of flow velocity (in meters per year) with depth. The curve from Salmon Glacier (Canada) represents less accurate measurements than the other two, and the irregularities in the curve are probably due to measurement error; this curve is plotted at half the scale (both horizontally and vertically) of the other two.

try of response about the hexagonal axis of the crystal.

Properties of single crystals provide a start toward understanding glacier ice; however, glacier ice is always polycrystalline (Fig. 4). Polycrystalline specimens show (14, 15), when tested in shear, a creep response much like that of single crystals but with the proportionality constant a reduced by a factor of about 100. The reduction is caused by unfavorable orientation and mutual interference among the crystals, but the detailed mechanism of this interference is not well understood. The problem is complicated by the occurrence of solid recrystallization (15), which modifies the crystalline texture progressively during deformation (Fig. 4) and is an important element in the creep mechanism.

The creep behavior of polycrystalline specimens in compression or tension is similar to that in shear, except that the strain rate under a given longitudinal stress is only about one-tenth that under an equal shear stress. The relationship here is basic to analysis of stress distributions and internal motions in glaciers, and is also of fundamental interest. It concerns the proper formulation of relations between stress and

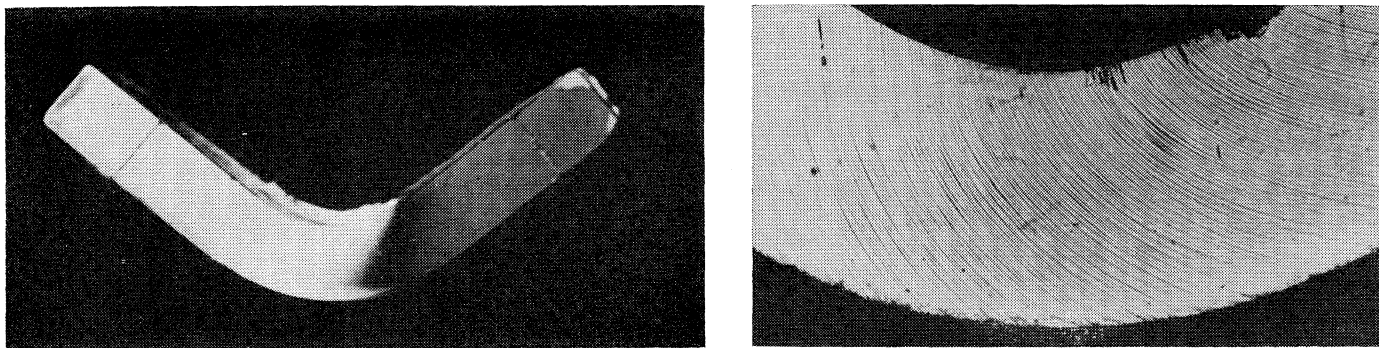


Fig. 3. Deformation of a single crystal of ice. (Left) Prismatic beam cut from a single crystal and bent by loading the center while supporting the ends of the beam (photographed under polarized light). (Right) Shadow photograph of the central part of the bent beam, in which "glide lamellae" parallel to the basal plane of the crystal may be seen (10).

strain rate for creep of solids under general states of stress. The simplest proposed representation for ice (16) has the form of a relation

$$\dot{\epsilon}_{ij} = f(\tau) \tau'_{ij}$$

between the strain-rate tensor components $\dot{\epsilon}_{ij}$ and the components τ'_{ij} of the stress-deviator tensor (stress tensor minus mean hydrostatic stress). The function $f(\tau)$ depends on the "effective shear stress" quantity

$$r = [1/2 \sum_i \sum_j (\tau'_{ij})^2]^{1/2}$$

which is a measure of the overall stress level pertinent to solid deformation. The function $f(\tau)$ is arbitrary and is to be determined by experiment. For the observed "power-law" response to stress

it has the form $1/2 a \tau^{n-1}$, which accounts approximately for the observed difference in behavior of ice in shear and in compression or extension tests (18).

A substance with the foregoing properties behaves like a "viscous" material in which the "effective viscosity coefficient" is a strongly decreasing function of the stress level. Under a shear stress of 1 kg/cm² the effective viscosity of ice (at -5°C) is about 10¹⁴ poise (the value for water is 10⁻² poise). For a doubling of the stresses, the viscosity decreases by a factor of about 5 and the strain rates increase about tenfold (19).

The strain rates are also sensitive to temperature, decreasing tenfold for a temperature drop of about 15°C. Measurements of the activation energy for

creep (9, 14, 15) show a wide scatter, for reasons not yet clear, but the value 14 kcal/mole tends to be favored at present, and probably corresponds to the activation energy for self-diffusion of protons in ice. This correspondence reflects the important influence of proton mobility and molecular reorientation on the properties of ice (20).

Internal Deformation of Glaciers

The manner in which ice approximates perfect plasticity may be seen in Fig. 5. A shear stress of 1 kg/cm² acting for a year produces a strain of about 30 percent. Hence, on a time scale of years, stresses much larger than 1 kg/cm² would produce extremely large de-

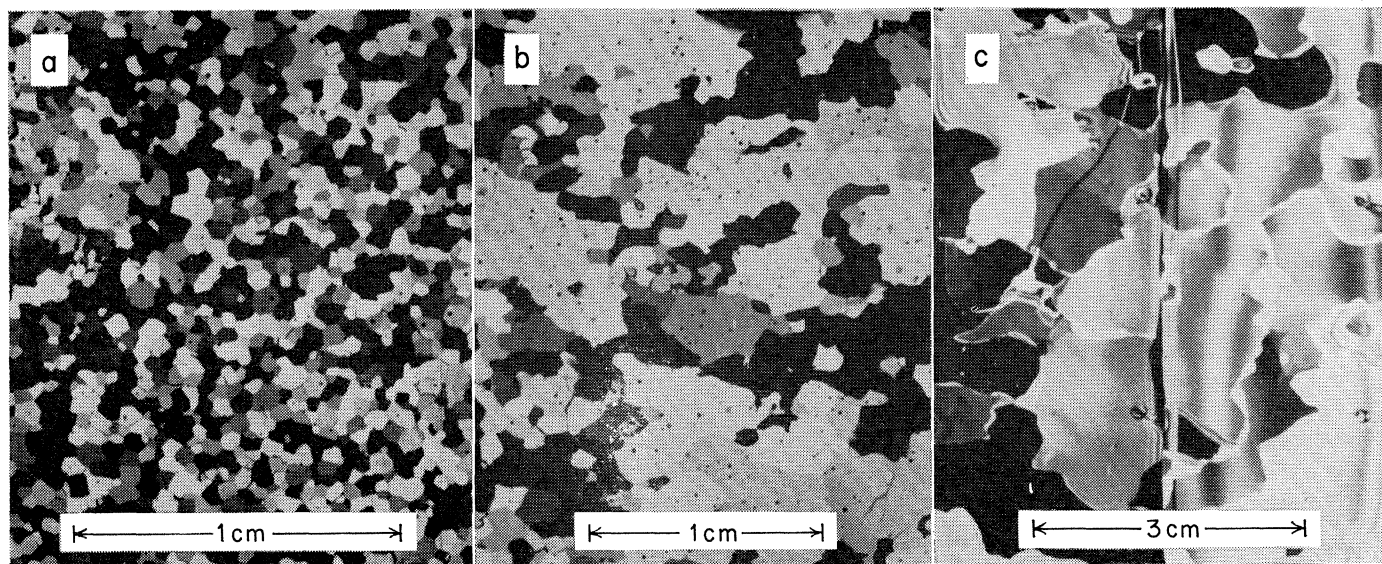


Fig. 4. Effect of deformation on the crystalline texture of experimental and natural ice samples, photographed in thin section under polarized light. (a) The starting material in an experimental study (41) of ice deformation and recrystallization; it resembles natural glacier ice at an early stage of its evolution from the original snow (46). (b) The texture that evolves from a after 3 months of deformation under a shear stress of 1 kg/cm² at -1.5°C. The crystals have grown greatly in size, and the original equidimensional, polyhedral shapes have evolved into the irregular, branching forms typical of glacier ice. (c) Glacier ice from a depth of 128 meters in Blue Glacier, Washington (43, 44) (the photograph shows two pieces of a core sample, side by side). The crystals in c are, on the average, about 50 times larger than those in a.

formations (because of the nonlinear response), and stresses much smaller would produce practically no deformation. This is why glacier ice has, effectively, a yield stress of about 1 kg/cm².

While the assumption of perfect plasticity thus gives a good rough approximation to glacier behavior, the more accurate flow law is necessary for detailed consideration of stress distribution and deformation within glaciers. The internal deformation is revealed by velocity profiles such as those of Fig. 2. Up to now the main effort has been to interpret profile measurements of this kind in such a way as to determine the flow behavior of glacier ice *in situ* and to compare this with the experimental results (5, 16, 21). Some discrepancy between the two may be anticipated, first, on account of differences in the texture of laboratory and natural ice (Fig. 4), and second, because stresses practical for experimentation lie generally above 1 kg/cm², whereas most stresses in glaciers are smaller than this. In fact the agreement is fairly good (Fig. 5).

The flow-law interpretation of velocity profiles is based on the fact that, for glaciers of suitable shape, the shear stress causing the observed internal shearing flow can be assumed proportional to depth beneath the surface, so that a plot of measured pipe-tilting (Fig. 2) against depth gives directly the relationship of strain rate to stress. However, the analysis of nonlinear creep (22) reveals an important difficulty in this interpretation: longitudinal strain in the glacier has a pronounced influence on the vertical profile of velocity (Fig. 6). The effect is most dramatic in the idealized case of perfect plasticity (Fig. 6a) but is also evident for nonlinear creep (Fig. 6b). Longitudinal strain (compression or extension of the glacier along its length) is caused by an interplay of two factors, ice accumulation (or ablation) and longitudinal variation in glacier slope, the latter requiring longitudinal variations in ice thickness, and the former, in ice flux. Such variations are universal in glaciers, and their influence on measurements of internal deformation is considerable (Fig. 5).

This peculiar effect, not found for ordinary fluids, is caused by the way in which effective viscosity varies with overall stress level in nonlinear creep. Under longitudinal compression or extension, the overall stress near the ice surface remains at a level needed to

maintain the longitudinal deformation, hence the ice there behaves approximately as though it were an ordinary fluid of constant viscosity, whereas in the absence of longitudinal deformation the effective viscosity increases upward, becoming infinite at the glacier surface.

This effect is clearly observed in ice tunnels, which gradually close up under the pressure of the overlying ice. Tunnels in ice that is deforming as a result of glacier flow, as at the base of the ice fall shown in the cover photograph, are found to close much more rapidly than tunnels at the same depth in ice that is free of flow stresses, and the closure rate varies linearly with the overburden pressure in the first case and nonlinearly in the second (16, 21).

The nonlinearity of the flow response is a major obstacle to detailed analysis of glacier motion. To date,

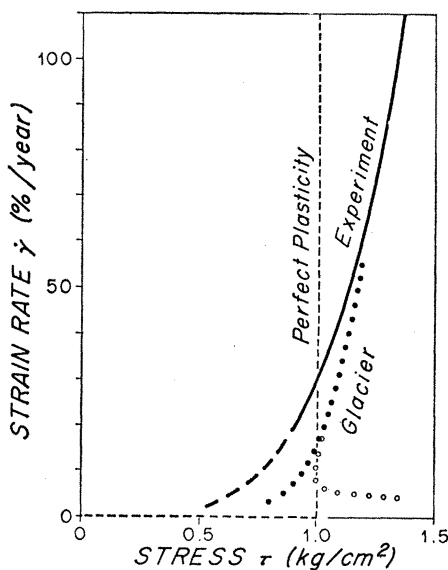


Fig. 5. Comparison of the flow behavior of ice as measured experimentally (solid curve) (see 14, 15) and as deduced from internal deformation in a glacier (solid dots), expressed as shear strain rate $\dot{\gamma}$ plotted against shear stress τ . The data for glacier flow are from boreholes in Athabasca Glacier, Canada (5) (see Fig. 2). Open circles show the flow curve that would be required by the data for measured internal deformation for borehole No. 322 if the longitudinal strain rate were constant with depth and equal to the measured surface value; such flow behavior would conflict decisively with the concept that the function $f(\tau)$ discussed in the text is a unique function of τ . A perfectly plastic substance would have the flow curve shown by the dashed line, whereas, for an ordinary viscous fluid, the curve would be a straight line passing through the origin of coordinates.

only one exact solution to a general problem in glacier flow has been found: a slab of uniform thickness and infinite lateral extent flowing down an inclined plane, when the state of stress does not vary laterally in the slab (22). It is thought that this solution should be applicable approximately to actual glaciers if the thickness and slope vary only slowly with position, and if, in valley glaciers, an approximate correction is introduced for the effect of the valley walls. The solution has therefore provided the basic theoretical framework for interpreting measurements of internal deformation in glaciers.

The analytical solution requires that the rate of longitudinal strain in the glacier should be independent of depth. Some recent measurements provide a test of this feature of the theory in a favorable situation, and it is found to fail rather seriously (23). In the first place, the rate of longitudinal strain averaged over any vertical line through the glacier can be determined from velocity measurements at the surface, provided one knows the slope of the bed from seismic data and the flow velocity near the bed from borehole measurements. The average rate of longitudinal strain is found to differ from the surface value by up to about 50 percent. Secondly, a comparable vertical variation in rate of longitudinal strain is found necessary in order to interpret the velocity profiles in a way compatible with the theory of nonlinear creep (Fig. 5).

These measurements are probably the expression of a basic difficulty in the analysis of internal deformation, a difficulty which can best be understood by comparing the velocity profiles for perfect plasticity with those for nonlinear creep (Fig. 6). With perfect plasticity there is a natural "decoupling" of the glacier from its bed, because the basal ice, which is everywhere at the yield stress τ_0 , can deform very rapidly and hence can effectively permit the occurrence of a basal velocity discontinuity (a slipping of the ice over its bed); such a discontinuity is in fact often observed (Fig. 2). For the actual flow law, however, there is no natural "decoupling." The velocity profiles of Fig. 6b must arbitrarily be terminated at whatever depth corresponds to the actual glacier thickness. A theory requiring homogeneous longitudinal strain must then lead to a longitudinally varying basal velocity discontinuity that does not conform to the creep law

used in the analysis. Furthermore, it requires that the magnitude of the velocity discontinuity shall have no effect on the resisting shear stress at the bed, an assumption that seems improbable (24). These difficulties bring us face to face with the problem of the basal "decoupling" in glaciers.

Basal Slip

That glaciers slide over their beds is well known from the grooves, scratches, and glacial polish that they leave on bedrock. Borehole measurements (Fig. 2) show that this basal slip is highly variable but contributes on the average about 50 percent of the total surface motion of valley glaciers (6), hence

no satisfactory analysis of glacier flow can be made without taking it into account.

Only recently has it become clearly understood how basal slip takes place (25). If the glacier is not frozen to bedrock, the resisting stress on the glacier bottom must be borne by pressure against the upstream faces of protuberances in the bed (Fig. 7). Since the melting point of ice is lowered by pressure, ice melts on the upstream faces and refreezes on the downstream faces, allowing the glacier to slip. The velocity of slipping by this "regelation" mechanism is limited by the rate at which heat can be conducted from the downstream faces, where it is liberated by refreezing, to the upstream faces, where it is absorbed in melting. Analy-

sis shows that, for a given bed roughness and slope, the slip rate is inversely proportional to the "wavelength" λ of the obstacles (Fig. 7). Since there are always some obstacles of very large size, the slipping would be essentially prevented were it not for the fact that deformation of the ice allows the slipping glacier to evade the larger obstacles. The slip rate, as governed only by ice deformation, is an increasing function of λ , hence there is an intermediate wavelength λ_0 which effectively controls the slip rate; it is the wavelength for which regelation and ice flow make equal contributions to the slipping, as illustrated in Fig. 7.

The basal slip rate under given shear stress should be determined primarily by the roughness of the bed at or near the controlling wavelength λ_0 , which has been found to be about half a meter (26). However, obstacles of about this size are often absent or inconspicuous on glaciated bedrock surfaces, probably because glacial abrasion is most effective at this wavelength (and below). This situation tends to invalidate the existing theoretical treatments (25) which assume a bed having a "white" roughness spectrum and predict a relation

$$v_B = K\tau_B^{3(n+1)}$$

between slip velocity v_B and basal shear stress τ_B , the constant K depending on bed roughness.

For $n = 3$, a value appropriate for glacier ice, a quadratic dependence of slip velocity on shear stress is predicted. This quadratic dependence does not conform to the mathematical flow solution requiring constant basal shear stress, and the lack of agreement may well be a source of conflict between analysis and observation.

Indication of another complicating feature in the basal slip phenomenon is the seasonal variation observed in flow velocities of valley glaciers (1, 27). Velocities in summer are higher than in winter, the difference increasing progressively downstream and amounting to a variation of 25 percent or more in the lower reaches. It is attributed to "lubrication" of the bed by summer meltwater, but the mechanism of this "lubrication" remains a subject of debate.

These complications are responsible for the fact that a detailed theory of glacier motion which can fully replace the simple but useful model of perfect plasticity has not yet been developed.

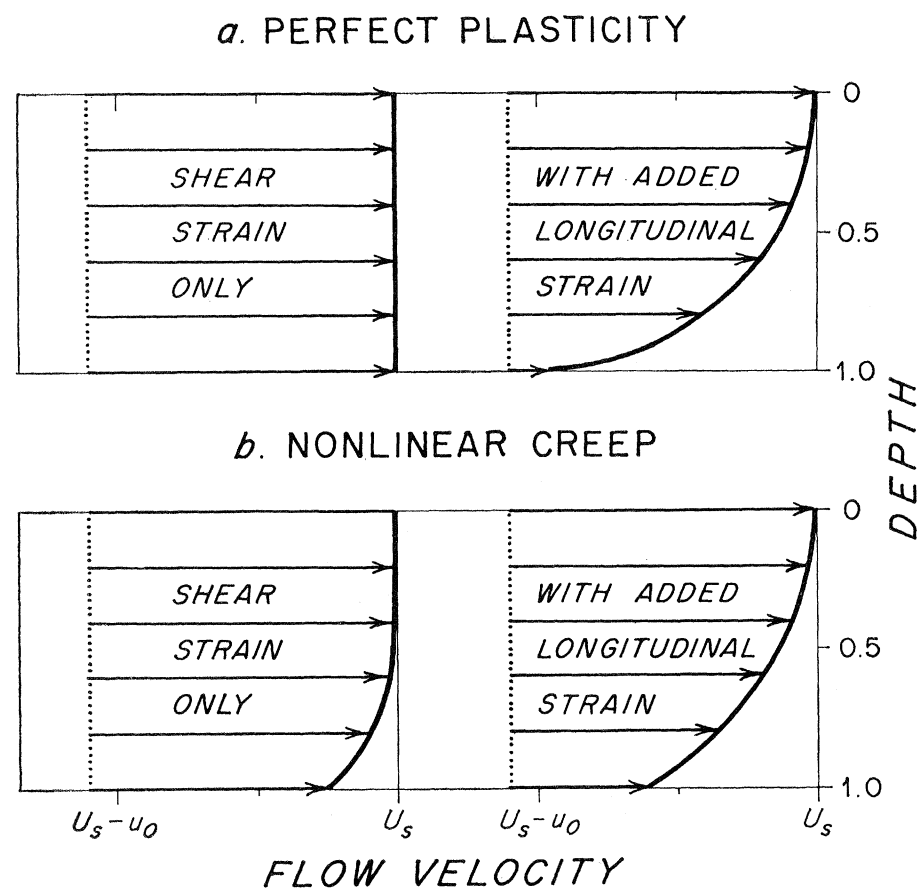


Fig. 6. Theoretical profiles of velocity plotted against depth (22), showing the predicted effect of longitudinal strain (extension or compression parallel to the flow direction) on internal deformation in glaciers, according to the theories of perfect plasticity (a) and nonlinear creep (b), calculated for $n = 3$. Velocities are plotted relative to the surface flow velocity U_s ; the zero of velocity (dotted lines) is arbitrary, since the theoretical solutions contain an arbitrary additive constant. The velocity variations are measured in units u_0 that depend on the dimensions of the glacier in question. For a typical case of a glacier 300 meters thick having a basal shear stress of 1 kg/cm^2 , the unit u_0 has a value of 45 m/yr . Depths are plotted in fractions of glacier thickness. The plots at right show the effect of superimposing a longitudinal tensile or compressive stress equal to twice the basal shear stress. For a basal shear stress of 1.0 kg/cm^2 , the corresponding longitudinal strain rate would be about 0.075 per year, a value which lies near the upper limit of those commonly found for valley glaciers, except near ice falls, where strain rates are considerably higher.

Thermal Regimes and the Flow of Ice Sheets

The great ice sheets of Greenland and Antarctica differ from typical valley glaciers not only in dimensions but also in thermal characteristics. A primary distinction must be made between glaciers in which the melting point of ice is reached throughout their mass and those in which the ice is entirely or largely below freezing. The former are called "temperate" glaciers because they occur typically in lower latitudes, as well as in some coastal areas of the Arctic, whereas the latter include the Greenland ice sheet and the glaciers of Antarctica and hence are called "polar." Glaciers are, of course, often polar in their upper parts, at higher elevation, and temperate at lower elevations, in particular below the snow line.

The discussion of glacier flow given above pertains primarily to temperate glaciers. The flow of polar glaciers is likely to be rather different, first because of the sensitivity of the mechanical properties to temperature, and second because basal slip does not take place if the bottom of the glacier is below freezing. Since the temperature distribution is itself modified by the ice motion, the flow problem is evidently a complex one.

Although some excellent thermal measurements have been made recently in Greenland and Antarctica (28), our knowledge of the temperature distribution in the ice sheets is still very incomplete; for example, it is not known with certainty whether the melting point is reached at the base of the ice. Hence, theoretical calculations of ice sheet flow (29) are of necessity still rather hypothetical, although they provide some important general concepts. There is an almost total lack of flow and deformation measurements against which to check these calculations, hence the discussion here is confined primarily to temperate glaciers, although many of the principles apply also to polar ice sheets.

Dynamic Response

The response of glaciers to changing climate has two aspects: (i) manifestation of climatic change in terms of alteration in the balance or imbalance between annual ice accumulation and ablation—the "mass budget" of glaciers

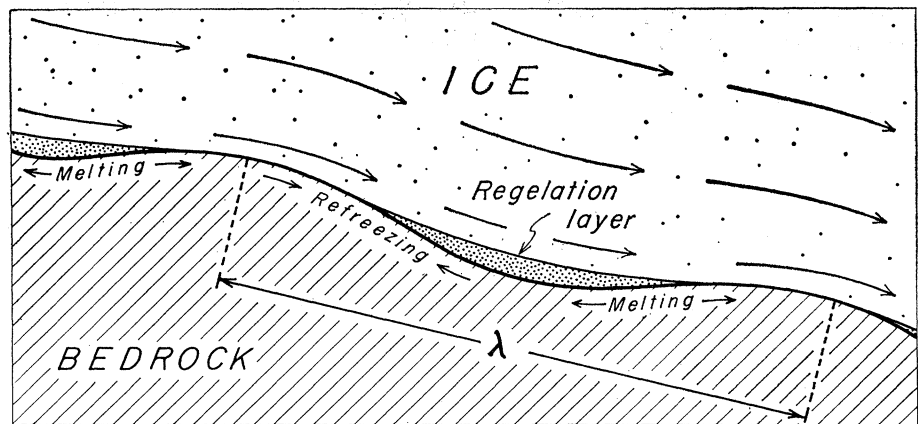


Fig. 7. Diagram illustrating the process of basal slip over bedrock topography of roughness-wavelength λ . The regelation layer consists of ice that melts in front of bedrock obstacles and refreezes behind them. When the maximum thickness of the regelation layer is half the height of the bedrock obstacles, as in this illustration, then regelation and ice deformation are making equal contributions to the basal slip, and the obstacle wavelength is in this case the "controlling" wavelength λ_0 at which the resistance to slip is a maximum.

—and (ii) response of glaciers to alterations in mass budget, which is seen as advance or retreat and as changes in ice thickness. The first aspect is basically a problem of meteorology, whereas the second is the expression of glacier dynamics.

Intensive study (30) of the energy budgets and mass budgets of glaciers in relation to meteorological variables shows that radiation balance (solar and sky radiation) is the dominating energy factor but that no single climatic variable is responsible for the changing mass balance, even though a general correlation has been established between increasing mean summer temperature and glacier retreat (31). The meteorology of glacier nourishment is, however, not yet amenable to quantitative treatment of the type now possible for the resulting glacier response, which depends on principles of glacier mechanics that are becoming well understood.

Glaciers are usually to a greater or lesser extent in the process of readjustment to changing conditions. Year-to-year fluctuations in weather conditions are so large that it is very difficult to discern long-term climatic trends, and annual mass budgets of glaciers show correspondingly large fluctuations. Glacier response represents a dynamic averaging of these fluctuations, and glaciers therefore can act as useful monitors of long-term climatic trends provided their dynamic response characteristics are understood. An understanding of dynamic response is also needed for predicting future glacier behavior from

meteorological data, a problem of much practical importance in some areas, particularly Switzerland, where during the past half century many dams and power plants have been built in places that would soon be threatened were a major glacier advance to set in. A successful start in the attack on these problems is a very recent achievement of glacier geophysics (32).

The response of glaciers to changes in mass budget is controlled through the requirement of mass conservation or continuity, which expresses in mathematical form the fact that the rate of growth at any point is determined by the balance struck locally among influx of ice from upstream, outflux downstream, and local ice accumulation or ablation at the glacier surface. Application of the continuity condition requires knowledge of a relationship between the configuration of the glacier surface and the flow rate, or, more precisely, the flux q of ice transported by the glacier at each point. According to glacier flow theory this should be expressible to a good approximation as a functional relation, of the type

$$q = q\left(h - \frac{\partial h}{\partial x}, x\right),$$

between the flux q and the local ice thickness h and surface slope $\partial h/\partial x$ relative to the bed. The dependence on position x allows for the fact that the slope of the bed and other factors may vary from place to place in the glacier. If the dependence on $\partial h/\partial x$ is neglected, which tends to be valid for steeper glaciers, the continuity condi-

tion then leads to a class of motions known as kinematic waves (33), which arise in many flow systems, such as rivers (flood waves), road traffic, and chromatographic columns. The flux values are found to propagate at a kinematic wave velocity $c = \partial q / \partial h$, which for glaciers should be about three to five times the actual ice-flow velocity.

The waves can propagate only in the downstream direction on glaciers, being in this respect unlike the more familiar dynamic types of wave motion.

The way in which kinematic waves control glacier response (31) is illustrated by a simple but appropriate case in which the accumulation area is

a region of uniform longitudinal extension and the ablation area one of uniform longitudinal compression. If the glacier is initially at equilibrium, and if an extra layer of ice of uniform thickness is suddenly added to the surface (as might be the situation after a winter of extra-heavy snowfall), the subsequent evolution of this perturbation in the glacier surface is as shown in Fig. 8a. The accumulation area reacts stably, and the extra thickness decreases exponentially, whereas the ablation area reacts at first unstably, with exponentially growing thickness. Stability is restored by the propagation downstream of a kinematic wave from the accumulation area. Near the glacier snout a considerable perturbation can build up before the stabilizing wave arrives; this should lead to erratic, "whip crack" behavior of the snout, which is sometimes observed. If, instead, the rate of accumulation changes abruptly to a higher value, the subsequent glacier reaction is as shown in Fig. 8b. In the accumulation area the surface rises and approaches the new equilibrium profile (dashed curve) at an exponentially damped rate, whereas the rise in the ablation area is at first "negatively damped" and exponentially accelerating. It becomes positively damped through control by a kinematic wave propagated down the glacier from the accumulation area.

The erratic, unstable behavior of the lower part of the glacier, as indicated in Fig. 8, tends to be mitigated by the influence of $\partial h / \partial x$ on the flow, which leads to a diffusion effect mathematically analogous to thermal diffusion (heat conduction) (32). This will cause spreading out of the kinematic waves and rounding of the sharp corners in Fig. 8.

Inclusion of the "diffusion" effect makes the elegant theory of kinematic waves no longer strictly applicable, and no comparably general solution of the equations for complete glacier response has yet been found. The only specific example for which the response equations have been solved so far (32) shows a response rather different from the nondiffusive case, particularly with respect to the time constant associated with the response at the snout. In the nondiffusive case this time constant is the time required for a kinematic wave to propagate from the accumulation area to the snout, and for typical glaciers it would be on the order of 25 years. In the diffusive case such a time constant is found for an

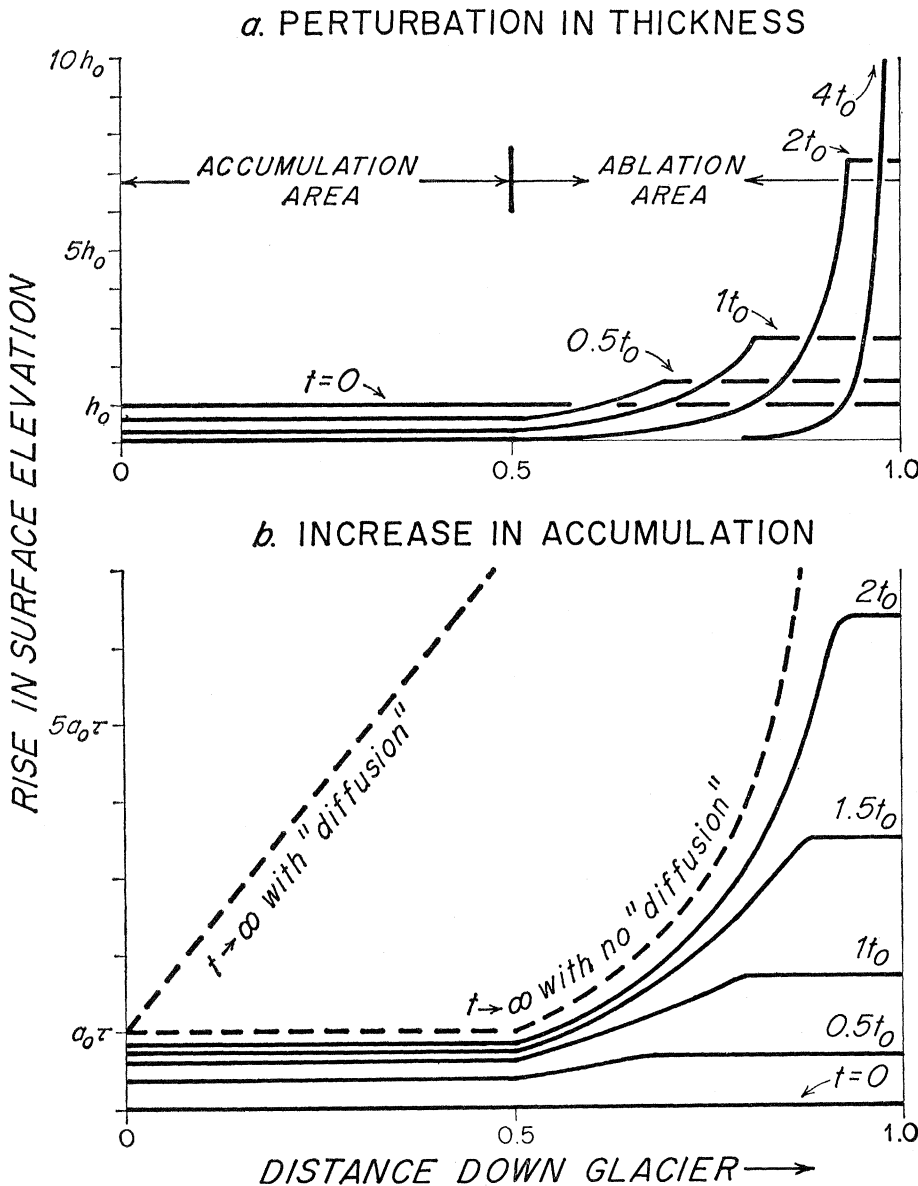


Fig. 8. Theoretical calculation of the response of a glacier to changes in ice accumulation, for an idealized case (32). The diagrams show how perturbations in the glacier profile (surface elevation plotted against distance downglacier in units of the glacier length) vary with time. The actual surface profile is not shown, but, rather, the deviations from an initial profile assumed to be in equilibrium with prevailing climatic conditions. In a the climatic conditions remain unchanged, but at time $t = 0$ an extra thickness h_0 of ice is added everywhere to the glacier surface. The subsequent profiles are given at times labeled in units t_0 equal to the time constant for decrease of the thickness perturbation in the accumulation area (left). The sharp corner propagating downstream represents a kinematic wave that originates at the accumulation-ablation area boundary (at distance coordinate 0.5). In b there is no initial perturbation in the profile, but at $t = 0$ the rate of ice accumulation increases by a constant amount a_0 over the whole glacier surface. The profile perturbation that will be attained when equilibrium is reestablished is shown by a dashed line, both for the pure kinematic wave treatment and for inclusion of the "diffusion" effect. (The unit of rise in surface elevation in b should be labeled $a_0 t_0$.)



Fig. 9. Glacier waves. This train of waves on the Austerdal Glacier in western Norway (see cover) illustrates a phenomenon found on many glaciers. The glacier is flowing from right to left, and the wavelength of the waves is 100 meters. The fine light and dark bands running parallel to the wave crests are the surface outcrop of internal structure (foliation) in the ice, which is seen here in greater detail than in the cover photograph.

initial, unstable part of the response, suggesting that kinematic waves continue to play a part, but the response is mainly governed by a much longer time constant, on the order of 800 years for a typical glacier. The discovery of this long time constant raises serious questions in the interpretation of glacier behavior.

A related discovery of the analysis (32) is a curious asymmetry between

(i) the problem of predicting glacier response from climatic data and (ii) the problem of deducing climatic trends from observations of glacier fluctuation. Predicting glacier response requires mass budget data over a time of the order of the long time constant, whereas deducing climatic trends requires observations of glacier fluctuation only over a time of the order of the short time constant. It thus appears that

problem ii is tractable whereas problem (i) is not, in the present state of our knowledge.

The long time constant originates not so much in the damping effect of diffusion on the kinematic waves as in the effect of $\partial h/\partial x$ on the equilibrium profile. The governing time constant for response to an increase in accumulation rate turns out, essentially, to be just the time required to fill up, at

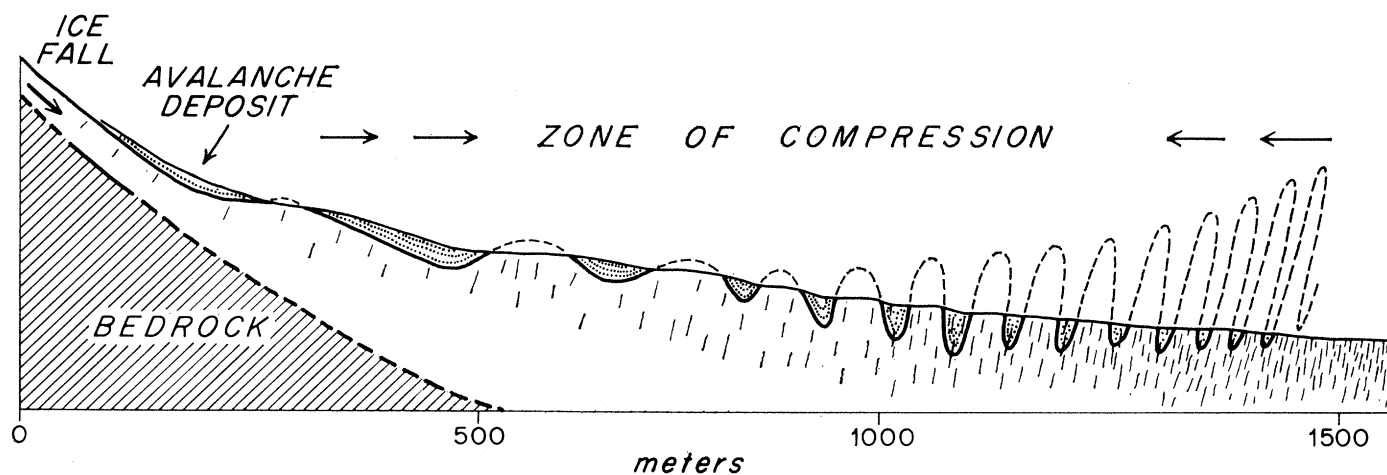


Fig. 10. Longitudinal cross section through Austerdal Glacier near the base of the ice falls (see cover). The snow and avalanche deposit that forms near the base of the ice fall preserves a record of an initial depression in the ice surface there and thus makes it possible to follow the subsequent evolution of this surface, the dashed curve in midair being reconstructed from the observed remnants on the basis of the required conservation of (ablated) ice volume. The actual surface waves (Fig. 9), which show faintly in the cross section, are minute compared to the wave train so reconstructed. The dashed pattern shows the progressive development of foliated ice structure cutting into the sedimentary layering of the avalanche deposit (dotted pattern).

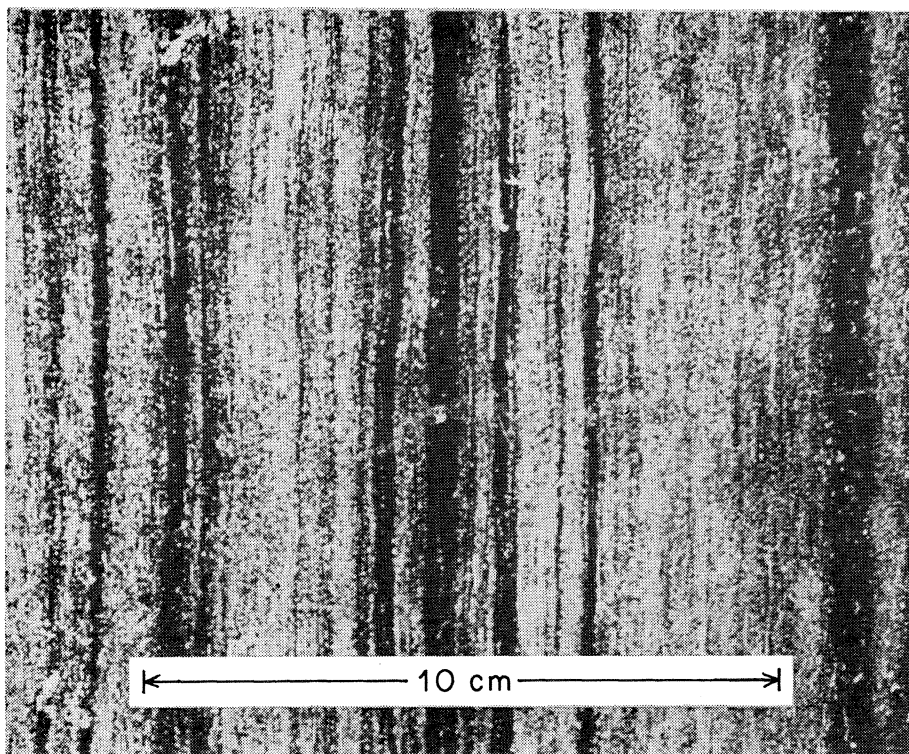


Fig. 11. Foliated structure in glacier ice from the Great Aletsch Glacier, Switzerland. The folia originate in the vertical attitude shown. The bright spots are air bubbles in the otherwise clear ice.

the extra accumulation rate, the additional ice volume needed to build the glacier up to its new equilibrium profile (the area under the pertinent dashed curve of Fig. 8*b*). This additional volume, which gives the "response inertia" of the glacier, is seen to be much greater if the "diffusion" effect is operative than if it is not.

Although analytical solution of the response equations presents difficulties, it has been possible to devise a numerical method for calculating the response characteristics of actual glaciers from field measurements, and the method has been successfully applied to South Cascade Glacier in Washington (34). It consists in calculating the frequency response of the glacier to sinusoidal climatic fluctuations and, from the results of this calculation, obtaining a set of constants λ_i that relate perturbations in ice thickness, h_i , at the snout to the current perturbation, a_i , in overall accumulation or ablation over the glacier, according to the series

$$a_i = \lambda_0 h_i + \lambda_1 \dot{h}_i + \lambda_2 \ddot{h}_i + \dots,$$

which requires knowledge of the current surface perturbation h_i and its time derivatives \dot{h}_i , \ddot{h}_i , and so on, from field records. Field data are not yet extensive enough to allow a thorough testing of this procedure, which opens

up for the first time the possibility of quantitative climatic interpretation of glacier variations.

Detailed application of these ideas has not yet been made to the well-known general retreat of mountain glaciers throughout the world, which has gone on over the past 50 years or so at a rate of 5 to 20 meters per year. Correlation with climatic trends (31) suggests that the retreat may be about to reverse itself, and recently there has been standstill or moderate-to-strong glacier advance in places, notably in the Pacific Northwest and on the island of Jan Mayen in the northern Atlantic (35). Although the theory for interpreting this phenomenon seems now to be at hand (34), field measurements detailed enough for its application have apparently been made on only one glacier in the world, and the necessary climatic observations have been made over a span of time much too short, so that an educated prediction of the future behavior of the world's glaciers is not yet possible.

Glacier Waves

Kinematic waves might also provide a mechanism whereby local surges in flow, caused, for example, by local avalanching of extra loads of snow

onto the glacier surface, could be propagated downstream at a speed much greater than that of the ice flow itself. Such local flow surges or waves have been inferred in several glaciers (1, 36), of which the best known are the glaciers of the Yakutat Bay region, Alaska. These glaciers advanced suddenly and spectacularly about 10 years after the great earthquake of 1899, which shook down much snow onto their upper reaches. Detailed field observations of surge propagation are as yet very scanty, the best-documented example being a 2-meter-high bulge that traveled down the Mer de Glace (France) from 1891 to 1899 at an apparent speed about five times that of the ice flow. The recent theoretical treatments that include diffusion (32) cast doubt on a simple kinematic-wave interpretation of such surges because they indicate that, for typical glaciers, the diffusion effect should dominate any wave propagation for surge wavelengths shorter than, roughly, half the length of the glacier. The recent Alaskan earthquake, which should produce effects much like those of the 1899 quake, offers an excellent opportunity for field study of this phenomenon and for careful evaluation of the theory.

The clearest wave form actually seen on glaciers (Fig. 9) does not move with a kinematic wave velocity but, instead, simply rides along with the ice. The wavelengths of these waves are of the order of the glacier thickness or less, and for such wavelengths the kinematic wave analysis breaks down. The waves are generated at the foot of ice falls (see cover), where there is a large longitudinal compression as the rapidly moving material in the ice fall is slowed down to the speed of the gently sloping glacier tongue below. It used to be thought that these are "pressure waves" associated with seasonal flow variations, since the crests are spaced a year's flow apart. However, there is a process that necessarily generates waves quite independently of any pressure-wave mechanism (37). Ice passing down the ice fall in summer is thinned more by ablation than ice that passes down in winter, and the resulting difference in thickness between "winter" and "summer" ice is amplified by the great longitudinal compression at the base of the ice fall, producing an annually repeating wave form.

This mechanism should, however, generate a wave amplitude that increases progressively through the compression zone, whereas a progressive

decrease actually takes place over most of this zone. The progressive decrease of the waves might be due to a plastic "settling" of the wave crests under their own weight, but it seems to be caused instead by more rapid ablation of ice from the crests than from the troughs of the waves. The behavior of this differential ablation is difficult to assess. It is therefore fortunate that, in the glacier where these phenomena have been carefully studied (see cover), there are internal structural features in the ice which serve as "tracers" for following the development of the wave pattern unaffected by ablation (38). They reveal a wave train closely resembling the one predicted by the theory (Fig. 10).

Internal Structure

Each "tracer" in Fig. 10 is a mass of snow and avalanche debris that accumulated in a corresponding wave trough at an early stage of amplification, at the head of the wave train. The subsequent evolution of these trough fillings not only defines the progressive deformation of the ice, but also provides object material for a study of the great structural change that takes place in glacier ice during deformation. Such change is designated "dynamic metamorphism" by geologists, and its recognition and interpretation in rocks constitutes one of the main types of evidence for great deep-seated deformations of the earth's crust in the processes of mountain building (39). From this point of view the trough fillings in Fig. 10 constitute a succession of nearly identical "sedimentary basins" whose progressive dynamic metamorphism can be followed in detail while it is taking place. This is unlike any observation possible in structural geology, and illustrates the potential usefulness of glaciers as model laboratories for studying crustal deformation in the earth (40). The flow of glaciers is in itself grossly similar to the important type of crustal deformation recorded in nappe structures, of which the classic examples are in the Alps, visible alongside the Alpine glaciers which they so strongly resemble.

Some of the structural changes that occur in deforming ice are illustrated in Figs. 4, 11, and 12. The most obvious is development of a distinctive layering or foliation (Fig. 11), consisting of inhomogeneous distribution of bubble-like inclusions of air trapped in

the original snow. Like the analogous structure common in deformed metamorphic rocks, foliation can be produced by passive deformation of original sedimentary layering, but the most intense foliation, like that of Fig. 11, is generated anew in the process of deformation itself, as shown by the fact that it cuts across the sedimentary layers (Fig. 10). A layered structure has been produced experimentally (41) by deformation of ice that was initially quite homogeneous, showing the opera-

tion of some inherent process of compositional segregation of the kind known in petrology as "metamorphic differentiation."

The foliation in Fig. 10 develops with simple, uniform orientation perpendicular to the direction of longitudinal compression. In three dimensions the pattern is more complicated. Figure 13 shows the structural pattern in Blue Glacier, Washington, revealed not only by surface observation (42) but also by subsurface data from core

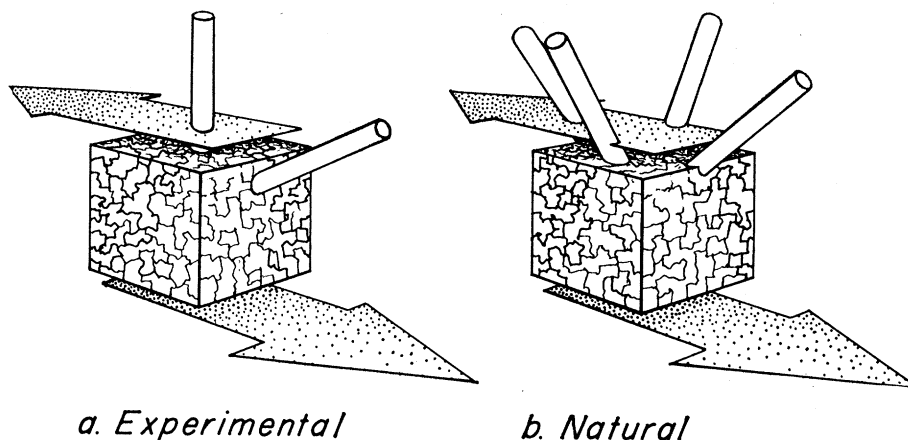


Fig. 12. Crystal orientations in deforming ice. For polycrystalline samples deforming in shear as indicated by the arrows, the orientations assumed by the hexagonal axis of individual crystals are depicted schematically by the cylindrical shafts in the diagram. Diagram *a* shows the orientations produced experimentally (41); *b* shows the orientations observed in samples of temperate glacier ice deforming under essentially the same stress conditions (43, 46).

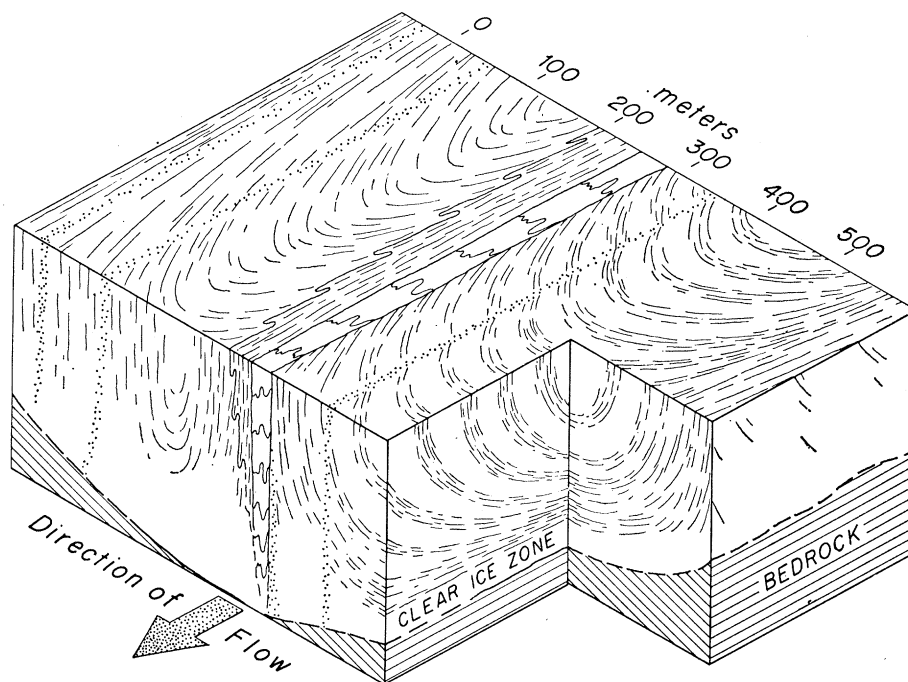


Fig. 13. Block diagram showing the internal structure of a portion of Blue Glacier, Washington. The dashed pattern represents ice foliation. Regions of uncertain structure are left blank. Core sampling (43) has revealed the detailed structure of the region exposed in the cut-out corner at front of diagram. The dark (heavily dashed) and light zones in that corner are structures of the same kind as the bands seen in the cover photograph and represent zones of contrasting foliation type.

sampling (43). Such patterns are typical of glaciers with ice falls, which cause pervasive, intense deformation throughout the ice mass. Their complicated features are due to the geometry of flow in ice falls (42). A complete interpretation cannot yet be made, partly because the mechanism of foliation generation is not fully understood. The problem is now being attacked in reverse through quantitative comparison of the observed foliation pattern (from core samples) with measurements of ice deformation at depth (43).

Accompanying the structural changes described is a pervasive alteration in the crystalline texture of the ice by recrystallization during deformation (Fig. 4). A similar phenomenon is observed in metals and rocks under conditions of high-temperature creep (44), and its dynamics are readily studied experimentally in ice (15). Deforming stresses promote crystal growth, but the ultimate crystal size is a decreasing function of the stress level. For this reason the crystals in glacier ice become smaller toward the bottom of a flowing glacier (43), even though the ice is presumably oldest there and the time available for crystal growth has been longest. Near the surface in the accumulation area, however, time is the dominating factor, and in fact a complex metamorphic process of grain growth and densification from porous snow to dense glacier ice first takes place before a texture like that shown in Fig. 4a is reached (45).

The most startling structural change occurs in the spatial orientation of the crystals in deforming ice (Fig. 12). Starting with initially random crystal orientation, a stress-induced recrystallization in the laboratory (Fig. 4, a and b) generates the strong preferred orientation of crystal hexagonal axes indicated in Fig. 12a (41). Under the same stress conditions in temperate glaciers a very different arrangement (involving 4 preferred c-axis orientations) is found (43, 46), shown in Fig. 12b. Until recently it could be hoped that this striking discrepancy simply indicated a qualitative difference in behavior between sub-freezing ice in the laboratory and temperate ice in nature, but the 4-orientation patterns have now been found in polar glaciers also (47). Another attractive explanation calls upon annealing recrystallization, such as occurs in metals after cold working (48): an initial orientation of the experimental type might be converted to the 4-orientation

pattern by annealing recrystallization near the glacier surface, where the deforming stresses are reduced. This possibility has been disposed of recently by the discovery of very strong 4-orientation patterns in the highly stressed ice at the base of a flowing glacier, by means of core sampling (43).

It seems, instead, probable that the discrepancy reflects a great difference between the time scales of laboratory experimentation and of the corresponding natural processes. If so, there is indicated here a fundamental difficulty in the study of tectonophysical phenomena by laboratory experimentation, a difficulty which confronts much excellent laboratory work on rocks going on at present, and which is all the more severe because of the very long time scale of millions of years for typical tectonophysical processes in the earth. For nonequilibrium processes like solid deformation, knowledge of the rate-controlling mechanisms appropriate to the different time scales involved is necessary for a safe extrapolation from experimental results to geological time scales (49).

Theoretical considerations on glacier flow have, up to now, ignored the strong orientation anisotropy indicated in Fig. 12, which is a general feature of glacier ice and which implies a strong flow anisotropy, because of the great plastic anisotropy of the individual crystals. Moreover there is now evidence from core sampling (43) that textural variations associated with glacier structure are correlated with variations of the flow properties, a possibility also overlooked hitherto. It seems that, until structural features of these kinds are understood and taken into consideration, theoretical discussions of glacier flow will continue to have a somewhat formal character. It also seems that, until we can account satisfactorily for the origin of internal structures in glaciers, where we can observe them in process of formation and can measure the deformations that produce them, the interpretation of corresponding structures in rocks deformed deep in the earth will remain somewhat speculative.

Summary

From physical measurements on glaciers and experimental studies of ice properties a framework of concept and

theory is being built which bids fair to place glaciers among the more quantitatively understandable phenomena in the earth sciences. Measurements of flow velocity, deformation and stress, ice thickness and channel configuration, temperature, internal structure of the ice, mass and energy balance, and response to meteorological variables all contribute to this understanding, as do still other measurements hardly discussed here, such as electrical properties, radioactive age measurements, and detailed studies of chemical and isotopic composition. The obvious goals of this work—the interpretation of past and present glacier fluctuations in terms of changes in world climate, and the prediction of glacier behavior—remain elusive, even though a good conceptual groundwork has been laid for dealing with the more tractable aspects of these problems. Intriguing recent discoveries have been made about such matters as the way in which glaciers react dynamically to changing conditions, the interrelations between thermal regime and ice motion, the structural mechanisms of glacier flow, and the changes produced in ice by flow. One can recognize in these developments the possibility that concepts derived from the study of glacier flow may be applicable to phenomena of solid deformation deep in the earth. In this way glacier geophysics may have a useful impact beyond the study of glaciers themselves.

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Middle Stone Age Culture in India and Pakistan

Recent investigations show a distinct culture
intermediate between the Early and Late stone ages.

H. D. Sankalia

The discovery of a distinct Stone Age period in the Indo-Pakistan sub-continent, with characteristic tools, faunal assemblages, river deposits, and inferred climatic conditions, is comparatively recent. Such a period was first recognized not more than a decade ago. The Age and its artifacts cover

the period between the Early and Late stone ages—that is, from approximately the end of the Middle Pleistocene to the beginning, or even into the early part, of the Holocene.

The discovery was in a sense the natural consequence of an organized search for man's early beginnings, and

subsequent development, in India. Until 1930, some two stone ages were discernible in India. These were the Paleolithic and the Neolithic. The latter included the microliths, though as early as 1906 Logan (1) had designated the period that intervened between these two stone ages the "Mesolithic." However, this classification into Paleolithic and Neolithic, proposed by Robert Bruce Foote (2, 3) who tried to arrange his discoveries of stone tools and pottery from several parts of India in some sort of chronological order, had become current. Foote's attempt at classification was soon followed by that of Coggin Brown (4), who dealt with a few Copper Age tools and weapons as well as Stone Age objects. However, no steps were taken to put this knowledge on a proper foundation, as Logan had urged and as Foote had tried to do. Even the discovery, in 1920, of the

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