FLOW AND FLUCTUATIONS OF GLACIERS

Vladimir M. Kotlyakov
Institute of Geography, Russian Academy of Sciences, Moscow, Russia

Keywords: Corrie, fjord, glacier crevasses, glacier erosion (exaration), glacier fluctuations, icefall, kinematic waves, ogives, trough valley (trough)

Contents

Summary

A glacier's movement is determined by two different mechanisms: the viscous-plastic ice deformations which depend on temperature and a whole array of compressing and stretching stresses caused by pressure from the above lying ice, on one hand, and by the ice sliding along a bed, sides and internal weakened surfaces, on the other hand. The character of the ice movement in a glacier is determined by the relationship between two main forces: the runoff force, stipulated by inclination and curvature of the bed, and that of spreading stipulated by the glacier surface slope with respect to its bed. The runoff force causes mainly the glacier laminar flow, while the spreading one primarily causes block sliding of ice. The temperature state of glaciers exerts great influence upon the velocity of the ice flow as the ice is more readily deformed under higher

temperatures. The ice flow velocities increase from several meters per year at small mountain glaciers to 50-200 $m{\cdot}year^{-1}$ at large mountain glaciers. The glacier flow velocities change from year to year and in different seasons of a year. At mountain glaciers, seasonal fluctuations of the velocities reach 50%. The velocities are the greatest at the glacier heads in late spring when a winter accumulation reaches its maximum. It results here in the formation of kinematic waves, which are transfers of portions of compression and extension forces down along the glacier that result in alternating mounds and depressions at the same places in the glacier surface topography. Stresses of extension and compression appear in glaciers during the ice flow. They are related to changes in the bed inclinations, narrowing or widening of the ice flow channel, changes in conditions on the bed and acceleration of the ice flow. If stretching stresses exceed the ice breaking strength they cause the formation of ice crevasses. Intensive glacier erosion, or exaration, takes place at the area of contact between the glacier and its bed. Climate changes bring permanent fluctuations in a glacier's mass and size. Fluctuations are recognized as forced oscillations caused by variations of external load, i.e. by the accumulation-ablation rate, and the relaxation auto-oscillations arising due to the non-stationary character of the kinematic links in the glacier; they are expressed in terms of glacier surges.

1. Introduction

Mountain glaciers slowly flow down along the slopes, glacier sheets and domes spread out from their centers to the peripheries. This motion of glaciers is governed by gravity, and it becomes possible owing to the property of ice that allows it to be deformed under pressure. Natural ice, being fragile in individual pieces, acquires a plastic property like solidified pine pitch, which cracks if one gives a stroke to it, but it slowly spread out on a surface if pressed at one place.

2. Mechanism of flow

2.1. Stresses and deformations in glaciers

The motion of glaciers is governed by two different mechanisms: the ice viscous-plastic deformations that depend on the ice temperature and a whole array of compressing and stretching stresses caused by pressure from the above-lying ice, on one hand, and sliding along the bed, sides and internal weakened surfaces, on the other one.

Stresses in glaciers are created throughout the entire mass of the ice. Gravity forces along with those of external and internal friction act on a glacier. Stresses of shear are equal to zero on the surface of a glacier and maximal on its bed or on surfaces of chipping. Inhomogeneous distribution of stresses in glaciers results in concentrations of them in separate parts. Where the stresses are greater than limits (about 10 000 $\boldsymbol{H\ m^{-2}}$for a break and 5 000 $\boldsymbol{H\ m^{-2}}$ for a shear), chipping takes place or crevasses are formed. Where the stresses are smaller, ice is deformed in accordance with the law of rheology.

Due to differences in ice velocities in glaciers, internal friction, i.e. friction between the ice particles and separate layers, arises. The value of this friction is proportional to a transversal gradient of motion in a plane-parallel flow and to a product of both stress and deformation tensors in a real glacier. Internal friction is especially great along the surfaces of breaks inside a glacier and in near-bottom layers where stresses and deformations are very strong.

Significant mechanical resistance arises at places where a glacier contacts its bed. This is caused by dry friction of the moraine fragments against the bottom and by the force of ice sticking to the bed. Liquid friction arises under the glacier as it slides along its bed (aprocess that causes the ice to melt). These frictional forces are created by the resistance of the ice to being deformed when flowing around the bottom roughness. These forces decrease when a water or water-clay film appears, which floods the bed bulges in some places, making the bed more slippery. The friction forces also decrease when parts of the ice break off from the bed (cavitation), which reduces the area of the glacier that contacts the bed. The forces increase when the sliding velocity increases (linearly at a small velocity and proportionally to its second degree at a greater velocity) and also when roughness grows (measured as the ratio between the bulge heights and distance between them).

Due to these stresses, ice deformations, both viscous-plastic and discontinuous, occur in glaciers. A simple shear, which is the simplest type of viscous-plastic deformation, occurs at the point of maximum velocity on a vertical near the point of intersection between the glacier axis and the equilibrium line. A complex shear deformation with parallel current lines without longitudinal expansion or compression, occurs when there is a velocity change along not only a vertical axis, but also across the glacier. This may be observed in a glacier's cross-section near the equilibrium line. Changes in ice velocity and discharges along the glacier due to accumulation and ablation results in deformations of the expansion along the current lines as well as compression across them in the accumulation area. These changes also cause compression deformations along the current lines and expansion across them in the ablation area. On the longitudinal section along the glacier axis, where no transversal deformation takes place, a plane tense-deformation state is observed.

Longitudinal expansion and compression change along a glacier following changes in the bed topography. A glacier's surface and floating ice shelves are in a state of one-axis or two-axis horizontal expansion or compression. The rest of the mass of ice, lying in valley glaciers with variable cross-sections or spreading to the edges of the ice sheets, is in a state of complicated three-dimensional deformation that also includes components of rotation (angle turn).

Typical rates of deformations in glaciers are 0.01 $year^{-1}$, while at cliffs they reach 0.1 $year^{-1}$, and at a moment of glacier surge it increases up to several units per year, though such states last only 1-2 months.

2.2. Thermal state of ice and glacier flows

The thermal state of glaciers exerts a significant influence upon the velocity of the ice flow because the higher a temperature, the more easily ice deforms. The thermal regime of the glacier is determined by the ice transport with its own temperature taken in other place, then by the temperature transfer from the air, bed and adjacent parts of the glacier by the thermal diffusivity, radiative heating through the transparent ice, heat released in the process of freezing water infiltrating into the glacier thickness, and heat released by internal friction.

A profile of the temperature distribution along a depth inside a glacier is formed by the joint effect of all the above factors. The temperature of the active layer undergoes seasonal variations owing to thermal diffusion and the heat released during the process of freezing water that has infiltrated the glacier. At the low boundary of the active layer, the ice temperature is determined by the mean annual air temperature and infiltration freezing. The volume of the latter increases with growth in the firn thickness. At deeper levels, the temperature profile is determined by the relationship between the heat flux from geothermal heat below, the heat caused by friction against the glacier's bed and inside near-bottom layers, and the heat conduction and advection of cold ice from above and along a horizontal. The greater the role of advection, the more pronounced is the temperature vertical profile, and, sometimes, even an inverse gradient may appear.

Additional heat is released as a result of the ice's friction against the bed, and if a glacier is cold, this heat is spent on warming the ice. In a glacier where the temperature is close to the melting point, this heat is spent on ice melting or increasing its moisture. Temperate glaciers flow faster than cold ones. Heat released from the glacier's motion also accelerates the flow.

2.3. Viscous-plastic flow and block sliding

A field of the ice velocities, caused by the deformations, can be calculated with some assumptions for isotherm glaciers. Velocities of sliding cannot be theoretically calculated, and they are empirically determined by the subtraction of calculated velocities of the viscous-plastic flow from measured total velocities. The viscous-plastic flow arises in a glacier due to viscous-plastic deformations, so that velocities of the movement are close to integrals over the deformation velocities along a streamline from its start to a measurement site. Glaciers with viscous-plastic movement are characterized by the ice freezing to the beds and sides, by a smooth increase in the velocities from the bed and sides to the axis of flow, and by a parabolic profile of the transversal diagram of the velocities. Sliding is especially typical for a glacier whose thickness is at the temperature of melting, while it is less inherent to cold glaciers. On mountain glaciers these two mechanisms of motion often play approximately equal roles (Figure 1).

When a glacier slides along a bed at the temperature of melting at its points of contact with the bed, the tension at the bed is determined by the glacier's mass, which is parallel to its movement, and by the friction force counteracting it. The friction force depends on the sliding velocity and the sizes of transversal cross-sections of the bed ledges, i.e. its sides and bottom. According to J. Weertman, the sliding mechanism is formed by the combination of ice melting before the ledges under pressure followed by overflow and re-freezing on the opposite side and of accelerated ice flow under the influence of the tension concentrations that occur near ledges. Large ledges retard the first mechanism, while small ledges interfere with the second one. A glacier slides with a velocity corresponding to the condition of velocity equality of both mechanisms, i.e. obstacles of average sizes. In this case, the velocity of the sliding is proportional to a square of the shear tension near the bed.

Fluctuations of the sliding velocity are connected with the presence of a water film near the bed. Under normal conditions, such a film appears due to the pressure and heat of friction on the contact surface, the ice becomes slippery, and the friction coefficient decreases down to 0.02-0.03. When ledges on the contact surface are prominent, sliding against the rough bottom is caused by deformation and phase transitions at a flow over the surface irregularities. When the film thickness exceeds the height of controlling obstacles, then conditions are formed that favor ice surges. Sliding increases under cavitation when there is not enough time for ice to fill depressions between obstacles.

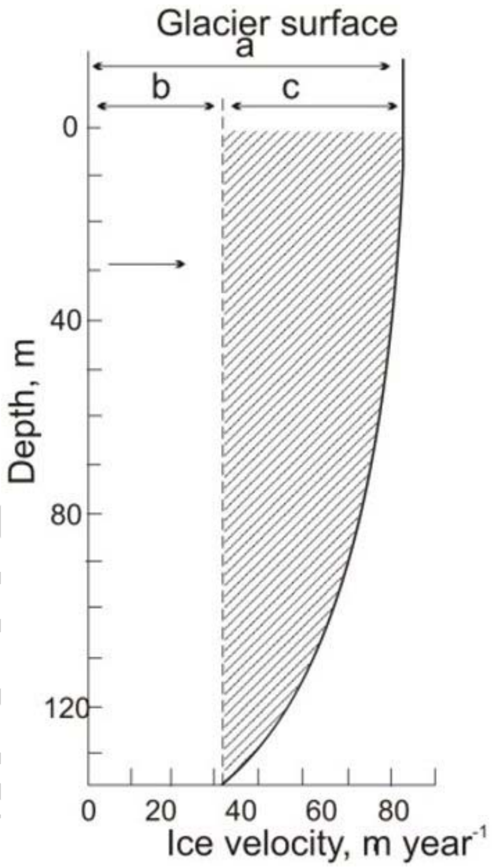

Figure 1. Total ice flow in the accumulation area of the Grossaletschgletscher (Swiss Alps) from August 1948 till September 1950:
a is total movement; *b* is movement due to sliding along the bed; *c* is movement due to ice deformation

Often sliding proceeds along breaks in a glacier, i.e. the ice block displacements relative one to another along chipping or weakened layers. Such sliding runs irregularly as jerks in movement alternate with rests during periods of tension accumulation. Sometimes, the displacement proceeds in a thin (several centimeters thick) weakened layer where ice deform exceeds by 2-3 orders the deformation inside the blocks. This layer is permeable to water and becomes moist ice when volume moisture reaches 15% owing to the effect of the friction heat.

When a real glacier moves, the relation between the viscous-plastic flow and the block sliding can be very different. In cold glaciers, ice frozen to the bed can move only by viscous-plastic deformations, while glaciers with a water film on their beds can move only by block sliding.

The smoothness of the ice velocity variations in glaciers indicates that glacier flow has features like a liquid. The velocity of a glacier that moves by ice viscous-plastic flow is determined mainly by the thickness of the ice, its temperature and the inclination of the glacier surface. Ice also flows down the surface inclination in cases where it meets ledges with inclinations inverse to the glacier bed. There is a regular relationship between the ice thickness, the surface inclination and the ice velocity in a glacier: if the surface is steeply inclined, ice is thin and moves faster, while ice is thick and slow where the inclination is insignificant. This may be seen in different parts of the same glacier and in different glaciers. Small irregularities on a glacier's surface do not influence its velocity if they are smaller than its thickness.

If the velocity of ice is small, the glacier moves in accordance with the laws of a viscous liquid, but when the velocity increases the glacier's movement becomes plastic and transforms into block sliding. As a whole, the ice deformation mechanism can be presented as follows:

- A slow shift in parallel to the basic plane of the ice crystal causes viscous flow without changing the ice structure.
- A small (less than 1 kg cm^{-2}) shear stress (tangential stress) activates a slow flow of polycrystal ice which is followed by migration re-crystallization, an increase in the crystal sizes along with ordering of orientation of their main axes. This is the basic mechanism of deformations in glaciers.
- If the stresses increase, ice starts to flow faster. It results in the breaking of the crystal spatial lattice, break-down of stress grains and primary re-crystallization.
- Under still greater loads and rates of deformation, links between crystals are broken, and the mechanism of deformation becomes similar to what is typical for snow and partly for firn: crystals slide one relative to another, their sizes decrease, and a chaotic structure is formed in the shear zone. The last two mechanisms are typical for near-bottom layers of quickly moving glaciers.
- Under tangential stresses exceeding 10 kg cm^{-2}, and small normal (perpendicular) loads, breaks and chipping appear in the ice.The ice is destroyed and slides along the planes of the fracture.
- If similar high tangential stresses are combined with large normal loads and sufficiently high temperature, it causes partial melting inside the ice along the planes of chipping. This mechanism is operating when a glacier is acting under block movement.

Progressive increases in the shear stresses as well as the prolonged action of these stresses cause the ice to go through stages of viscous, plastic and fragile body. They also introduce into action different mechanisms of plasticity, i.e. from sliding along the basic planes to displacement of crystals relative to each other.

On the whole, the dependence between irreversible deformations of the ice and stresses caused them determines the rheologic properties of the ice. Ice's ability to be irreversibly deformed characterizes its plasticity, while its viscosity is defined as the ice's resistance to the flow that is directly proportional to its velocity. Experiments have made it possible to find a great variety of ice viscosity coefficients, ranging from $10^3$to 10^8 *MPa s*. This variability of the viscosity coefficient is connected with the stress increases over time as well as by changes in the ice properties from those of a viscous (Newton) liquid. That is why ice can be imagined as an ideal plastic body with its limit of fluidity close to 0.1 *MPa*. Such a body is not deformed under stresses smaller than the ice fluidity, but it is deformed under larger stresses with any rate depending on external friction. Actually, a creeping takes place in the ice, i.e. the ice flow under stresses is lower than the fluidity limit.

2.4. Equations of the ice flow. Law of Glen

When considering glacial flows, the main problem is how to calculate distribution of the ice thickness and its velocities, which hold up the glacier in a stationary state under the prescribed rates of accumulation and ablation along with surface temperatures, as well as how to calculate the glacier system's reaction to variations of these characteristics. Such a calculation is made based on an analysis of the equations of discontinuity and motion (balance) of forces with due regard to the rheologic law for ice.

Equation of discontinuity expresses the law of conservation of mass for a continuously moving medium. It means that a sum of masses, entering a volume under consideration and leaving it, is equal to the change of the mass inherent in the volume for the same time $\boldsymbol{\tau}$ due to change of density $\boldsymbol{\rho}$:

$$\frac{\partial \rho}{\partial \tau} + \frac{\partial (\rho \mathrm{u_i})}{\partial \mathrm{x_i}} = \mathrm{v}.$$

Here, $\boldsymbol{U_i}$ are vector components of the movement velocity along directions of orthogonal axes $\boldsymbol{x_i}$ ***while v*** are internal sources of mass (mainly, the water freezing in pores of the firn).

In a monolith and practically impenetrable ice, its density does not change, and internal sources of mass are absent. Then, the discontinuity equation reduces to $\partial(\rho \mathrm{U})/\partial \mathrm{x} = 0$. In snow and firn, density changes are possible due to compression as a result of subsidence, horizontal extension and infiltration freezing. The discontinuity equation is broken if crevasses and cavities are formed in the ice, and, in such cases, these cavities should be taken into account as the density changes.

Another basic equation expresses the law of conservation of momentum. A sum of projections onto coordinate axes of all forces exerted to the volume is equal to acceleration given to a body:

$$\frac{\partial s_{ij}}{\partial x_i} + \rho g_i = \rho \frac{du}{d\tau}.$$

Here, $\boldsymbol{x_i}$ are coordinates, ***U*** is velocity, τ is a time, ρ is the density, $\boldsymbol{g_i}$ is a component of the free fall acceleration in direction $\boldsymbol{x_i}$. Under small ice velocities in glaciers, the acceleration (inertial force) is very small (8-10 orders smaller than the other members). Consequently, the equation of motion reduces to the equation of equilibrium $\partial s_{ij}/\partial x_i + \rho g_i = 0$. Stresses on the ice facets and the gravitational force act upon the ice volume.

At the very beginning of the 20th century, it was supposed that ice is deformed as a Newtonian viscous body.

But, by the end of the 1940s, it was understood that ice, being a polycrystal solid body, should be deformed like other polycrystal solid bodies, i.e. like metals and stone at temperatures close to their melting temperatures. Analysis, performed by E. Orovan and J.

Nye and based on a hypothesis that ice behavior is similar to that of an ideal plastic material, clarified somewhat the properties of ice flow. Later on, J. Glen determined on the basis of laboratory experiments that the relationship between the rate of ice deformation and stress is non-linear, at least within the limits of stresses significant for glacier movement.

Thus, it was clarified that viscosity is not a constant, but depends on stress. The law of Glen was taken as the basis for a theory of glacier flow.

Ice velocity can change with time. The deformation of stochastically oriented polycrystal ice first slows down, then it does not change for a long time (secondary flow), and after the deformation reaches a value of about 10%, it transfers to a tertiary, still faster flow.

The initial slowing down is connected with an interaction between differently oriented crystals. Transfer to the tertiary flow is reached by the ordering of the ice crystal orientation in the course of re-crystallization.

In the case of secondary flow within a range of stresses typical for glaciers (0.05-0.2 MPa), the relationship between the velocity of shear ε and the shear stress σ is expressed by the law of Glen:

$$\varepsilon_{xy} = k\sigma_{xy}^{n},$$

where ***k*** and ***n*** are empirical parameters. Values of these parameters are determined in a laboratory or in the field from deformation of glacier boreholes under the assumption of the shear; ***n*** changes from 1.5 to 4.5 with the transfer from the secondary flow to the tertiary one. An average value of ***n*** is about 3. Different types of rheology laws are presented in Figure 2.

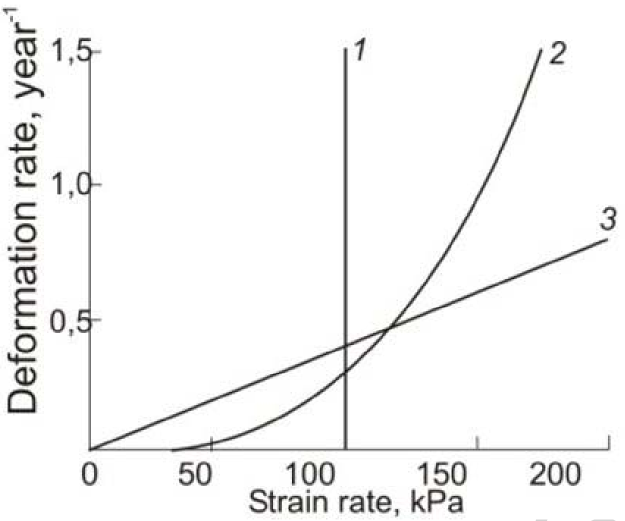

Figure 2. Different types of ice flow laws:
1 is ideal plasticity with limit of fluidity 100 *kPa;* 2 is the Glen law with n = 3 and k = 0.28 10^{-15} $year^{-1}$ Pa^{-3}; 3 is the Newtonian viscous flow with a viscosity of 8 TPa s

The curve of dependence of the ice flow velocity on the stress of the shear τ has a point of inflection corresponding to some critical stress τ_0. Until this stress is reached, the flow runs at a very small velocity according to a law similar to the law of flow of viscous Newtonian liquid (***n*** is equal to approximately 1.5). When the stress exceeds a critical point, the exponent ***n*** increases 2-3 times, and the flow velocity sharply increases taking on the character of a plastic flow. Thus, ice flow starts under even the smallest stresses, and the ice has no limit of fluidity. Changes in the velocity of cold glaciers are determined by the acceleration of the deformations with a respective rise in the temperature and increase of the glacier thickness.

-
-
-

Bibliography

Gudmundsson G. H. (1994).*Converging glacier flow – a case study: the Unteraarglacier*, 120 pp. Zürich: ETH. [Uses conceptual models of general flow characteristics of covering flow to draw inferences on stress and strain regime of glacial confluence of Unteraargletscher, Switzerland.

Gudmundsson G. (1994). *Glacier sliding over sinusoidal bed and characteristics of creeping flow over bedrock undulations* (Versuchsanstalt für Wasserbau, Hydrologie. Mitteilungen 130), 102 pp. Zürich. [Describes ice flow close to bedrock undulations in detail.]

Hooke R. L. (1998). *Principles of glacier mechanics*, 248 pp. Upper Saddle River, NJ: Prentice-Hall. [Detailed mathematical textbook on flow of glaciers and origins of landforms produced by glaciers.]

Hutter K. (1983). Theoretical glaciology, material science of ice and the mechanics of glaciers and ice sheets. Tokyo: Reidel.

McIntyre N. F. (1985). The dynamics of ice sheet outlets. *Journal of Glaciology*, **31**, 99–107.

Soushez R. A. and Lorrain R. D. (1991). *Ice composition and glacier dynamics*, 207 pp. Berlin: Springer–Verlag. [General view of relationships between ice composition and ice flow is considered, followed by more detailed study of key basal zone of ice caps, ice sheets and Alpine glaciers.

Van der Veen C. J. (1999). *Fundamentals of glacier dynamics*, 462 pp. Rotterdam: A.A. Balkema. [Introduces modeling flow and dynamics of glaciers, including role of ice in climate system, mechanics of glacier flow, basal sliding, large-scale dynamics of ice sheets and Alpine glaciers.]

Willis I. C. (1995). Intra-annual variations in glacier motion: a review. *Progress in Physical Geography*, 19, 61–106.

Biographical Sketch

Vladimir Mikhailovich KOTLYAKOV (born in 1931) is a member of the Russian Academy of Sciences (elected in 1991). He is Director of the Institute of Geography, Russian Academy of Sciences. With particular interest in glaciology and physical geography in polar and mountain regions, he directed the twenty-year project resulted in the World Atlas of Snow and Ice Resources (published in 1997).

V.M. Kotlyakov participated in many expeditions. He worked and wintered in the Arctic, the Antarctica, at the slope of the highest summit of Europe, the Elbrus, headed the high mountain glaciological expeditions to the Pamirs.

The main theoretical results of V.M. Kotlyakov`s works consist in elucidation of laws of snow and ice accumulation of the Antarctic ice sheet as well as ice sheets in general (1961), the snowiness of the Earth and its fluctuations within time and space (1968), the tasks and abilities of the space glaciology (1973), the application of isotope and geochemical methods to the study of the environment and its evolution (1982), the study of the past for four glacial-interglacial cycles (1985 and further on). During the last years, V.M. Kotlyakov dealt with the global changes of the environment, geographical aspects of global and regional ecological problems, the problems of interaction between the Nature and society.

V.M. Kotlyakov is the vice-president of the Russian Geographical Society and the President of the Glaciological Association. In 1983–87, V.M. Kotlyakov was elected the President of the International Commission of Snow and Ice, in 1987–93, he was the member of the Special, and later Scientific, ICSU Committee of the International Geosphere-Biosphere Programme, in 1988–96, the vice-president of the International Geographical Union. Now he is a member of the Earth Council.

V.M. Kotlyakov is elected a member of the Academia Europaea and the Academy of Sciences of Georgia, a honorary member of the American, Mexican, Italian, Georgian, and Estonian Geographical Societies.