

PHYSICAL PROPERTIES OF GLACIAL AND GROUND ICE

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Summary

Glacier ice covers about ten per cent of the Earth's land surface at the present time and covered about three times as much during the ice ages. The great ice sheets are of Greenland and Antarctica. Classification of glaciers is according to ice temperature and amount of surface melting. Very few glaciers show the entire sequence: dry snow zone, percolation zone, wet-snow zone, superimposed-ice zone. The transformation of snow to ice in the percolation and wet snow zones differs from that in the dry snow zone. A superimposed-ice zone represents the extreme case. Density increase in ice results from compression of air bubbles by the overburden pressure. Crystal size increases at a constant rate if the temperature remains constant. A single ice crystal deforms under an applied stress. A single ice crystal normally deforms by gliding on its basal planes. In glaciers and ice sheets, the ice is deformed for hundreds or thousand of years, to total strains of perhaps 10 or more. Ablation includes all processes by which snow and ice are lost from the glacier. Ice in glaciers is re-crystallizing continuously. Crystal size of the ice depends on salinity. The petrofabric analysis of ground ice may help to infer growth processes and conditions. The ice grain diameter, density and shear modulus increase with the increase of the depth. Porosity of solid transparent ice is as high as 1 kg (cm³/kg) and of solid sea ice is 1–50 cm³/kg. Effective thaw heat of sea ice decreases abruptly with increase of salinity. Regelation is movement of heavy objects through the ice without any disturbances to the ice. Due to molecular thermal oscillations and defects of ice crystals the water molecules can diffuse. The electrical properties of ice are distinguished by the fact that the charge carriers in ice are protons.

1. History of Glacier Study

Glacier ice covers some ten per cent of the Earth's land surface at the present time and covered about three times as much during the ice ages. However at present all but about 4 per cent of this ice is in areas remote from the normal activities humankind in the great ice sheets of Greenland and Antarctica. Thus it is not surprising that relatively small glaciers were the first to attract attention. Description of glaciers can be found in the 11th century Icelandic literature, but the fact that they move, does not appear to have been noticed, or at any rate recorded until some 500 years later. Altmann in 1751 correctly recognized that gravity was the cause of glacier motion. But he thought that movement consisted entirely of the ice sliding over its bed. Many glaciers do slide in this way, but the ice itself can flow, somewhat a very viscous fluid, as Bordier suggested in late 18th century. A proper understanding of the mechanism of glacier flow has been reached by application of the ideas of solid state physics and metallurgy. This followed the realization that as ice is a crystalline solid; it should deform like other crystalline solids such as metals near their melting points.

Agassiz began systematic measurements on glaciers about 1830 in the Alps. He found that the velocity is greatest in the central part and decreases progressively towards each side. He also found that a glacier moves more slowly near its head and terminus than elsewhere.

Noteworthy early work on polar ice sheets is the study by Koch and Wegener of snow stratigraphy during their crossing of Greenland in 1913. Koch and Wegener also measured temperatures in the ice in one instance down to a depth of 24 m. Wegener's Greenland Expedition of 1930–1931, which wintered in the central part of the ice sheet, studied the way in which snow is transformed to ice. They also made seismic measurements of ice thickness— a method first tried earlier in the Alps.

That glaciers advance and retreat is response to change in climate is common knowledge; but the relationship is much more complex than is usually assumed. Ahlmann between 1920 and 1940 carried out classic investigations on this subject on glaciers in Scandinavia, Spitsbergen, Iceland, and Greenland (Paterson 1981).

The dynamics of glaciers and other large ice masses, temperature regime of glaciers, glaciers' balance and detail study of glaciers' physical properties have been considered in detail by Budd (1969) and Paterson (1981)

Modern study of glaciers is based on detail investigations of their microphysical properties (Hondoh, Fukuda, Mae, Watanabe and others), deformation, texture, and grain growth (Alley, Budd, Gow, Woods, Duval, Dahl-Jensen, Langway, Goodman, Thorsteinsson, and others) isotope (Dansgaard, Grootes, Johnsen, Jouzel, Lorius, Kotlyakov, Merlivat, Petit, Stuiver, Thompson, Vasil'chuk) and physical properties, electrical properties and chemistry, palaeoclimate and palaeoatmosphere (Dansgaard, Grootes, Hammer, Jouzel, Thompson, Vasil'chuk), firn densification, close-off and air bubbles and clathrate hydrate (Hondoh, 2000).

2. Structure of Ice Crystal

The three nuclei of ice molecule can be pictured as forming an isosceles triangle with the oxygen nucleus at the apex and hydrogen nuclei (protons) at the other two corners. The oxygen atom has eight electrons, two of which circle close to the nucleus. Another two rotate in eccentric orbits, each of which also contains the electron from one of the hydrogen atoms. Each of these orbits thus encloses the oxygen nucleus and one proton. The remaining four electrons rotate in two other eccentric orbits. The four eccentric orbits radiate tetrahedrally from oxygen nucleus. The electron orbits completely screen the oxygen nucleus. They do not screen all positive charge of the protons, however, and they also provide an excess negative charge in the direction of two orbits without protons. Thus the ice molecule is regular tetrahedron with positive charges in two corners and negative charges in the other two. Each negative corner attracts a positive corner in a neighboring molecule, joining the molecules together by hydrogen bonds. Thus four other molecules surround each molecule in a regular tetrahedral arrangement. The spacing between molecules in ice is 0.276 nm.

A substance in which every atom has four neighbors in regular tetrahedral arrangement can crystallize hexagonally or cubically. Molecules or rather, the oxygen atoms are arranged in layers of hexagonal rings. The atoms in rings are not in one plane but in two: alternate atoms are in the upper and lower planes. The spacing between these two planes is much less than the spacing between layers. Adjacent layers are mirror images of each other. The structure of ice crystal resembles that of a hexagonal metal such as magnesium

or cadmium. The plane of a layer of hexagonal rings is called the basal plane of the crystal. The direction at right angles to the basal plane is the optic axis or c-axis.

3. The Transformation of Snow to Ice

The term snow is usually restricted to material that has not changed since it falls. The strict meaning of “firn” is wetted snow that has survived one summer without being transformed to ice. In this sense the term should not be applied to transformed snow in regions where is no melting. The broad definition suffers from the drawback that there is no clear division between snow and firn and we may sometimes use snow when firn would be more appropriate. The difference between firn and snow is clear; firn becomes glacier ice when the interconnecting air passages between the grains and is sealed off. A grain may be a single crystal or an aggregate of several. This occurs at a density of about 830 kg/m^3 . In glacier ice air is present only as bubbles and any increase in density results from compression of them (Table 1).

New snow (immediately after falling in calm)	50–70
Damp new snow	100–200
Settled snow	200–300
Depth hoar	100–300
Wind packed snow	350–400
Firn	400–830
Very wet snow and firn	700–800
Glacier ice	830–910

Table 1: Typical densities (kg/m^3) (after Paterson 1981)

Model of physical densification of snow, firn and ice (Arnaud et al., 2000) is constructed on the basis on the grain sliding in the first stage and than by plastic deformation and structural evolution of a porous material in later stages. Differences in initial firn structure are included through the density of the snow–firn transition. Empirically this density depends on temperature. Three stages of densification are generally distinguished. During the first stage, the densification of snow is mainly a structural rearrangement of grains by grain boundary sliding. This particle rearrangement is complete at critical density of about 0.55 g/cm^3 , corresponding to a relative density 0.6, which can be regarded as the maximum packing density. This critical density increases with temperature and the depth at which this density increases as the temperature decreases. At -57°C it is approximately 30 m and at -23°C it is approximately 12 m. In addition to the dependence on temperature, the accumulation rate for a given temperature can also affect to the depth of the critical density. The second stage is firn (consolidated snow). Because seasonal variations of the temperature disappear, the transformation of the firn into ice can be considered as isothermal. The end of this stage is characterized by growth of the impermeable pore space fraction under the increasing overburden pressure. The average number of bonds per grain (called coordination number) increases with density from approximately 6 at the start of this stage to 16 for the ice (Arnaud et al., 2000). Plastic

deformation increases this coordination number and causes neck growth; particle motion is negligible. During the final stage, atmospheric air is trapped in cylindrical or spherical pores; further densification of this bubbly ice is driven by pressure lag between the ice matrix and the air in bubbles.

It is suggested a link between density and reconstructed past close-off depth and ages for Vostok core. At Vostok, the transition from firn to ice, corresponding to the end of the pore close process was observed for a density of about 0.84 g/cm^3 . The age of the ice at this transition is around 2800 yr. At the Byrd, the firn to ice transition is observed for a density of 0.825 g/cm^3 and the age of ice is only 250 years (Barnola et al., 1991).

4. Glacier Classification

Geophysical classification of glaciers bases on ice temperature and mount of surface melting. This categories are temperate, sub-polar, and high polar. A temperate glacier is at a pressure melting point throughout. There is no surface melting in a high polar glacier. However conditions vary from one point of glaciers to another; very few glaciers can be fitted into a single category. Thus to speak about different zones in glacier is better than trying to classify entire glaciers. The characteristics of the zones starting from the head of the glacier are described. Very few glaciers show the entire sequence:

1. *Dry snow zone*: No melting occurs here, even in summer. The boundary between this zone and the next one is called dry snow line.

2. *Percolation zone*. Some surface melting occurs in this zone. Water can percolate a certain distance into snow at temperatures below 0°C before it re-freezes. If water encounters a relatively impermeable layer it may spread out laterally for the some distance. When it re-freezes an ice layer or ice lens is formed. The vertical water channels also refreeze, when their water supply is cut off to form pipe-like structures called ice glands. As the freezing of 1 g of water releases enough heat to raise the temperature of 160 g of snow by 1 K, re-freezing of melt water is the most important factor in warming the snow. As summer advances successively deeper layers of snow are raised to the melting point. The amount of melt water produced during a summer generally increases with decrease of elevation. Thus as we are go down glacier, we eventually reach a point where by end of the summer, all the snow deposited since the end of previous summer has been raised to the melting temperature. This point the wet snow line is the boundary of the next zone.

3. *Wet-snow zone*: In this zone, by the end of the summer all the snow deposited since the end of the previous summer has been raised to 0°C . Some melt water also percolation into the deeper layers, which were deposited in previous years, though not necessarily in sufficient quantity to raise their temperature to 0°C . Percolation into these layers may also occur in the lower part of the percolation zone. It is important to find out where this happens because, when it does mass balance measurements cannot be restricted to the current year's layer.

4. *Superimposed-ice zone*. In the percolation and wet-snow zones the material consists of ice layer, lenses, and glands, separated by layers patches of snow and firn. At lower

elevation, however, so much melt water is produced that ice layers merge to continuous mass of ice. This is superimposed ice. We restrict the term superimposed ice zone to the regions where there is an annual increment of superimposed ice exposed at the surface. Superimposed ice is also formed in the lower part of the wetted zone, but there it is buried beneath the firn. The boundary between wet-snow and superimposed ice zones is called a snow line. It has also called a firn line, firn edge, and annual snow line. Its location is easily determined. It is a boundary between firn and ice on the glacier surface at the end of the melt season. The lower boundary of the superimposed ice zone is taken at the equilibrium line. This line is important in mass balance studies. Above it the glacier has a net gain of mass over the year. Below it there is a net loss. Some superimposed ice is formed below the equilibrium line, but it is lost by the end of the summer.

5. *Ablation area*: This is area below the equilibrium line.

The only dry-snow zones are in the interiors of Greenland and Antarctica and near the summits of the highest mountains in Alaska and Yukon. It was found that the dry-snow zone in Greenland roughly coincides with the region where the mean annual temperature is -25°C or less. The mean annual temperature is in a dry-snow zone also is approximately equal to the temperature at the depth of 10 m in the firn (Paterson, 1981). However in some summers a little melting occurs even at the highest elevation in Greenland. Thus the whole sequence of zones may be found in parts of Greenland and Antarctica. On the other hand, the dry snow line on the Filchner Ice Shelf in Antarctica appears to lie about 5 km from the outer edge on the shelf. Thus the ice shelf and the part of the ice cap that drains into it is entirely a dry snow zone except a small percolation zone near the sea. The whole mass loss results from the calving of icebergs. The whole sequence, except for dry-snow zones, occurs in some large glaciers in Northern Ellesmere Island and Axel Heiberg Island. In cold summers there may be dry snow zones on the highest icefields in these areas. The Barnes Ice Cap in Baffin Island on the other hand appears to consist only of superimposed ice and ablation zones in most years. All these are 'cold' glaciers, that is, the temperature are below the pressure melting point.

In a 'temperate' glacier the ice is at the pressure melting point throughout, except for a surface layer, some 10-m thick, in which the temperature is below 0°C for the part of the year. Temperate glaciers can not have percolation zones because in that zone, by definition, the temperature of part of the current year snow pack, and thus the temperature of deep layers, never reaches 0°C . Again, superimposed ice only forms if the firn the firn temperature is below 0°C . On temperate glaciers the extent of any superimposed ice zone is insignificant and, for practical purposes, the equilibrium and snow line coincides. A temperate glacier thus has only wet snow and ablation zones. The reverse statement is not necessarily true. Each winter's cold wave will normally penetrate through several annual layers in the firn. For a glacier to be temperate the cold wave must be completely eliminated by the end of the summer. For a region to be classified as a wet-snow zone, it is only necessary that the cold wave be eliminated from the current year's snow.

The gradual transformation of snow to glacier ice results from changes in the constituent crystals and the intervening pore space, such as: mutual displacement of crystals changes in size and shape, internal deformations.

The relative importance of these mechanisms changes as the density of the material increases. The net result, however, is that the crystals grow and join together while the air spaces between them are eliminated.

Changes in crystal size and shape occur readily because, unlike other solids, ice is usually near its melting point. Molecules are thus relatively free to move both within the ice lattice (volume diffusion) and over the crystal surface (surface diffusion). In addition sublimation occurs readily. At least in the early stages of transformation, sublimation is more important than either volume or surface diffusion.

The net directions of movement of molecules are governed by the thermodynamics principal that the free energy of the system tends to a minimum. A reduction in surface area reduces a free energy. Thus the molecules tend to be redistributed. In a way that reduce the total surface area of the crystals. Thus fresh snowflakes, with their complex shapes, are gradually transformed to spherical particles. The larger crystals tend to grow at the expense of the smaller ones at this further reduce the free energy.

However the most important factor in the initial stages of transformation is settling, that is, the displacement of individual particles relative to their neighbors. The rounding of the particles makes this easier. We can estimate the increase in density that settling can bring about, by considering the group of spheres, all of the same size. Packing experiment spheres show that, in practice, one can never reduce the porosity below about 40 %, for spheres of ice of density 910 kg/m^3 and porosity 40% corresponds to a density of 550 kg/m^3 . Other mechanism must be responsible for any further increase in density. A packed arrangement of spherical particles is not the end result. Transfer of material can further reduce the total surface area to the points of contact between particles to the point s of contact between particles, to form bonds. This process is called sintering. As the density increases and the firm becomes less porous, sublimation is greatly reduced. At the same time the load and the area contact between grains are increasing. Re-crystallization becomes the dominant process: the size and shape of grain change in such a way as to reduce the stress of them. When density reaches about 830 kg/m^3 , the air spaces between grains close off, much of the air has escaped to the surface: the remainder is now present as a bubbles. The firm has become glacier ice. Further increase in density results from bubbles. Most of these processes are sensitive to the temperature; thus the rate of transformation varies from place to place.

In certain circumstance the transformation of recently fallen snow proceeds in a way quite different from the previously described. Instead of small round grains large crystals in the shape of prisms pyramids or hollow hexagonal cups are produced. These are called depth hoar. This is the most coarse-grained type of firm that can be formed in the absence of melt water. The average grain size is in the range of 2 to 5 mm, but some crystals can be much larger than this. A layer of depth hoar is highly porous and its density is low (100 to 300 kg/m^3). As name implies depth hoar is produced within snow pack in unconsolidated snow, but not on the surface. Conditions seem to be especially favorable when the snow is lying on the top of the much denser material such as ice. A strong vertical temperature gradient, to produce a strong gradient of vapor pressure, is essential. Such conditions usually occur in autumn when the surface is cooling rapidly and the underlying layers are still relatively warm. Evaporation takes place in the lower layers. The vapor rises and

condenses to form depth hoar crystals in the cold upper layers, especially at the places where they are relatively impermeable. Some vapor may also escape to the snow surface. Growth of depth hoar crystals represents an increase in the internal energy of the system; this is supplied by the temperature gradient.

Depth hoar occurs widely. A layer usually only a few millimeters thick develops in polar glaciers each autumn. It is easily recognized in the walls of a pit and provides a marker for measuring the annual snow accumulation. Depth hoar forms at ground snow interface in winter and spring in snowfields. In Central Alaska the entire snow pack (0.5 to 0.7 m) develops into depth hoar. Temperatures and temperature gradient in this area are higher than in Polar Regions and conditions suitable for depth hoar formation persists throughout the winter. Depth hoar layers are of particular concern in mountain areas because they act as fracture and sliding planes for avalanches.

The transformation of snow to ice in the percolation and wet snow zones differs from that in the dry snow zone. There is of course a difference only during summer and down to the maximum depth to which melt water penetrates. Packing of grains is still the important factor in the initial stages. Melting increases the rate at which grains become rounded, because the grains melt first at their extremities. The average grain size increases because the smaller grains tend to melt before the larger ones. In addition grains may join together in clusters by regelation (re-freezing after pressure melting). Joining the grains is particularly rapid in the surface layers because they undergo a daily cycle of freezing and thawing. Melt water accelerates packing by lubricating the grains and permits very close packing because the surface tension of a water film tends to pull the grains together. Thus the maximum density which can be attained by packing is higher in a melt area than in dry-snow zone.

Re-freezing of melt water also speeds up the later stages of transformation. Air spaces are filled in this way. Re-freezing of large quantities of melt water to form ice layers and lenses represents a rapid transition from snow to ice. The time needed to complete the transformation will vary widely between different areas, according to the amount of melt water. A superimposed-ice zone represents the extreme case. Elsewhere, and at other times of year, the processes are the same as in dry-snow zone.

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geochronology, Quaternary Geology, stratigraphy, geocryology, glaciology and geomorphology. He undertook field investigations in nearly all permafrost regions of Eurasia, such as Gydan and Yamal Peninsulas in the North of Western Siberia, Central and Northern Yakutia, Chukotka, Magadan region, Trans-Baikal region and Arctic Islands. Yu.K.Vasil'chuk is the author of over 200 publications, from them there are 7 monographs, such as: "Oxygen-Isotope Composition of Ground Ice" (Application to paleogeocryological reconstructions) 2-volum issued in 1992 and the textbook "Principles of Isotope Geocryology and Glaciology" (coauthored with Academician RAS V.M.Kotlyakov) issued in 2000 et al., about 20 papers he has published in "Transactions of Russian Academy of Sciences" and more than 25 ones in the International Journals, such as *Radiocarbon*, *Permafrost and Periglacial Processes*, *Nuclear Instruments and Methods in Physics Research B*, *Earth and Planetary Science Letters* etc. His recent textbook, "Soil Engineering» (2005, Lomonosov' Moscow University Press), was co-authored with V.T.Trofimov et al. This textbook characterized the ground ice as a base for constructions. Currently he prepared the new book "Ice wedge: Heterocyclity, Heterogeneity, Heterochroneity" issued in 2006.