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Snow and Ice on the Earth's Surface

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EDITOR'S FOREWORD

"Cold Regions Science and Engineering" consists of a series of monographs summarizing existing knowledge and providing references for the use of professional engineers responsible for design and construction in Cold Regions, defined as those areas of the earth where frost is an essential consideration in engineering.

Sections of the work are being published as they become ready, not necessarily in numerical order, but fitting into this plan:

I. <u>Environment</u>

  A. General

    1. Geology and physiography
    2. Perennially frozen ground (permafrost)
    3. Climatology

  B. Regional

    1. The Antarctic ice sheet
    2. The Greenland ice sheet

II. <u>Physical Science</u>

  A. Geophysics

    1. Heat exchange at the earth's surface
    2. Exploratory geophysics

  B. The physics and mechanics of snow as a material

  C. The physics and mechanics of ice

    1. Snow and ice on the earth's surface
    2. Ice as a material

  D. The physics and mechanics of frozen ground

III. <u>Engineering</u>

  A. Snow engineering

    1. Engineering properties
    2. Construction
    3. Technology
    4. Oversnow transport

  B. Ice engineering

  C. Frozen ground engineering

  D. General

IV. Miscellaneous

F. J. SANGER

SNOW AND ICE ON THE EARTH'S SURFACE

by

Malcolm Mellor

CHAPTER I. NATURAL FORMS OF ICE

Water is an abundant and essential part of our environment, its distribution influencing every human activity. Perhaps 2% of the total amount of water on the earth is in the solid state as ice; this is approximately equal to the quantity of water in rivers and lakes and in the ground. Ice appears in a variety of natural forms, each with its own characteristic mode of formation and life cycle. In the present text, the aim is to describe snow and ice as they occur naturally, with particular emphasis on glaciers, since they represent the great bulk of the world's ice.

Snow

The atmospheric conditions necessary for snow formation exist over all parts of the world, but snow can only reach the ground without melting into rain if temperatures remain cold as it falls towards the surface. Snowfalls are therefore experienced only in regions where surface temperatures drop close to freezing at some time of the year; at high latitudes this includes all places, but at low latitudes only the high elevations receive snow. The amount and frequency of snow depend on the availability of moisture for precipitation and on the prevalence of low temperatures, factors which themselves depend upon continental and maritime situations, movement of weather systems, latitude and altitude, and local orographic effects.

The initial formation of snow usually occurs at high altitudes in the atmosphere, when supercooled water droplets change into tiny ice crystals by freezing, usually stimulated by seeding of nuclei, or when vapor condenses directly on solid nuclei. Terrestrial dust, particularly from clay minerals, appears to be the major source of freezing nuclei.[8]* The form of the crystal is governed by the temperature at the time of first freezing and by the temperature and supersaturation† conditions prevailing during its growth. The results of Nakaya's researches on the dependence of crystal type on air conditions are summarized in Figure I-2a. More recent work by Mason led to the modified scheme shown in Figure I-2b.

Nakaya drew up a general classification of snow crystals, which is reproduced here in Table I-1 and Figure I-3. Table I-2 gives probable dimensions for various types of crystals appearing in the classification. Nakaya's system is valuable scientifically, but it is too detailed for practical use in meteorology and related fields. A simpler international classification has therefore evolved so that snow can be tersely and accurately described in symbols independent of language barriers, and with codes suitable for telegraphic transmission (Fig. I-4). The notes of Table I-3 explain the relation of the international snow classification to Nakaya's general classification.

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*Numbers refer to references, listed after each chapter.

† Supersaturation s is given by $s = \frac{w + \rho'}{\rho_0}$, where w = liquid water content per unit volume, ρ' = saturation vapor density over supercooled water, ρ_0 = vapor density over ice.

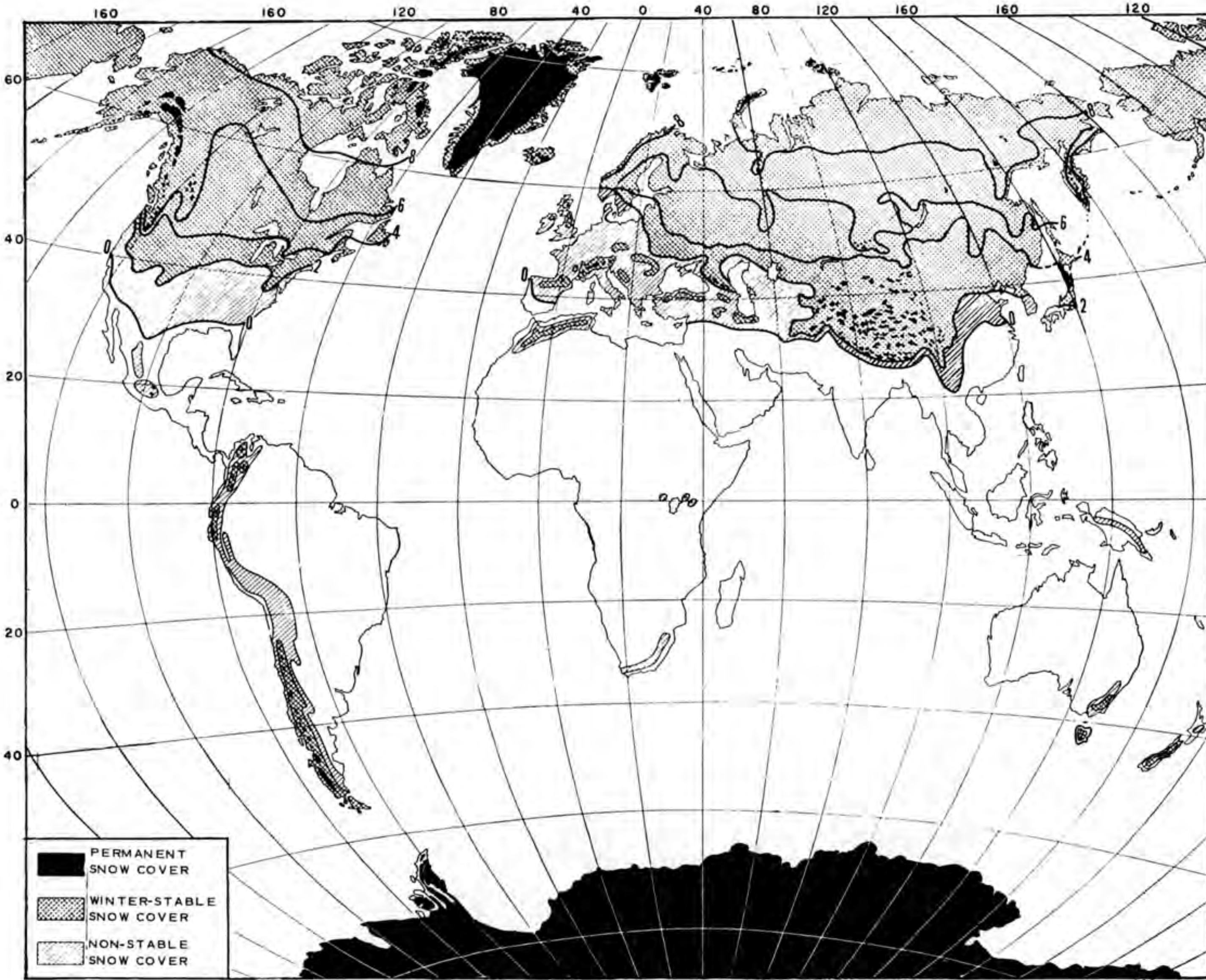
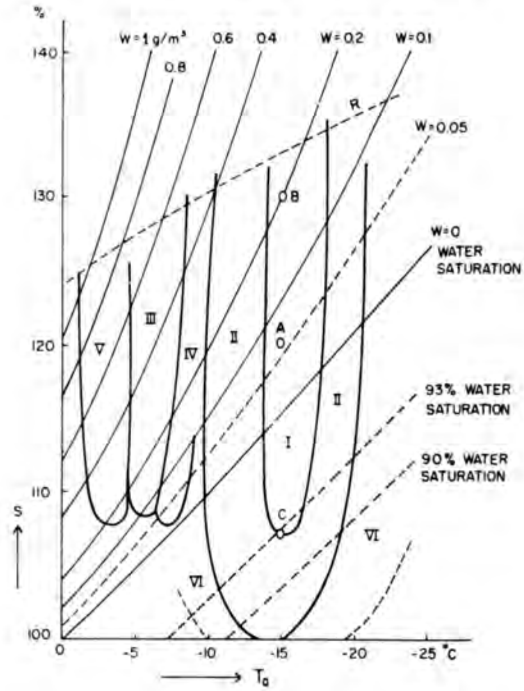
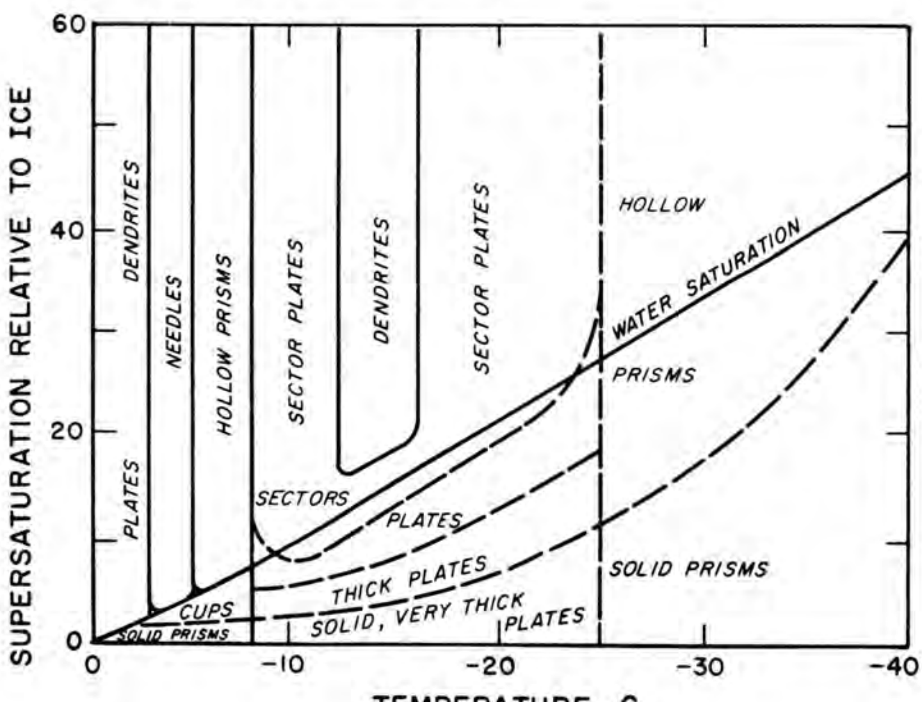


Figure I-1. World distribution of snow cover.
1. Permanent cover of snow and ice; 2. A stable snow cover of varying duration forms every year; 3. A snow cover forms almost every year, but is not stable; 4. No snow cover; 5. Duration of snow cover (months).
(Rikhter, ref. 17)



I-2a. (After Nakaya, ref. 12)



I-2b. (After Mason, ref. 8)

I-2. Conditions of temperature and supersaturation governing formation of various snow types.

Table I-I. General classification of snow crystals. (After Nakaya, ref. 10)

I	N	Needle crystal	1. Simple needle	a. Elementary needle b. Bundle of elementary needles
			2. Combination	
II	C	Columnar crystal	1. Simple column	a. Pyramid b. Bullet type c. Hexagonal column
			2. Combination	a. Combination of bullets b. Combination of columns
III	P	Plane crystal	1. Regular crystal developed in one plane	a. Simple plate b. Branches in sector form c. Plate with simple extensions d. Broad branches e. Simple stellar form f. Ordinary dendritic form g. Fernlike crystal h. Stellar crystal with plates at ends i. Plate with dendritic extensions
			2. Crystal with irregular number of branches	a. Three-branched crystal b. Four-branched crystal c. Others
			3. Crystal with twelve branches	a. Fernlike crystal b. Broad branches
			4. Malformed crystal	Many varieties
			5. Spatial assemblage of plane branches	a. Spatial hexagonal type b. Radiating type
IV	CP	Combination of column and plane crystals	1. Column with plane crystal at both ends	a. Column with plates b. Column with dendritic crystal c. Complicated capped column
			2. Bullets with plane crystals	a. Bullets with plates b. Bullets with dendritic crystals
			3. Irregular assemblage of columns and plates	
V	S	Columnar crystal with extended side planes		
VI	R	Rimed crystal (crystal with cloud particles attached)	1. Rimed crystal	
			2. Thick plate	
			3. Graupellike snow	a. Hexagonal type b. Lump type
			4. Graupel	a. Hexagonal graupel b. Lump graupel c. Conelike graupel
VII	I	Irregular snow particle	1. Ice particle	
			2. Rimed particle	
			3. Miscellaneous	

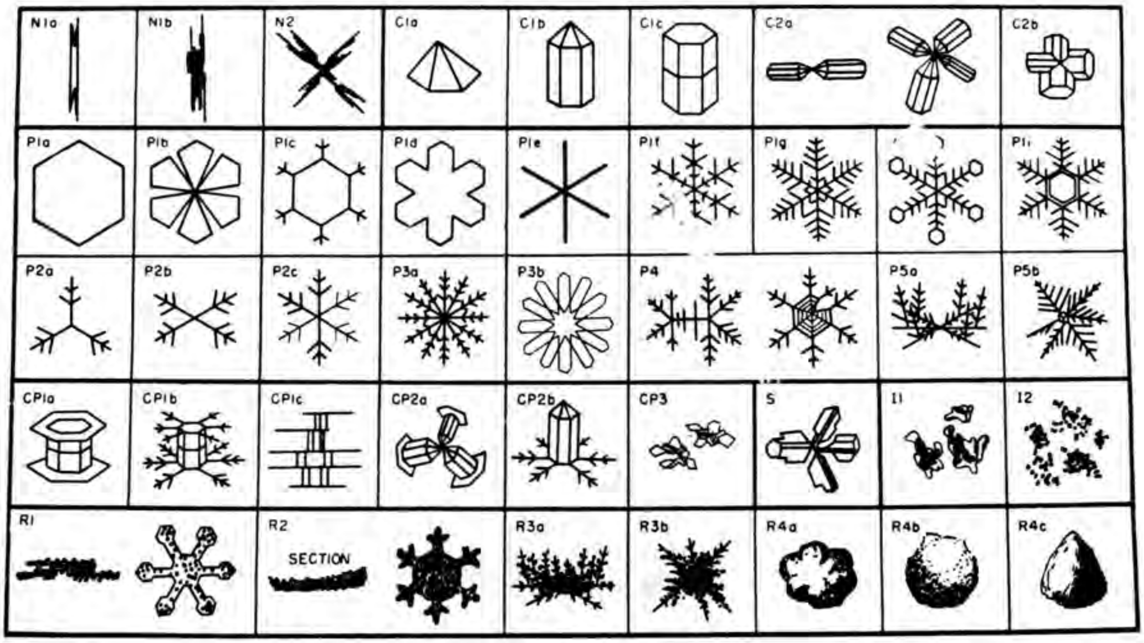
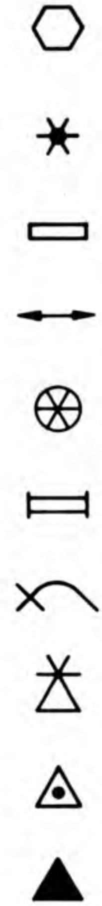


Figure I-3. General classification of snow crystals. (Nakaya, ref. 10)

Table I-II. Most probable dimensions of crystals. (After Nakaya, ref. 11)

No.		Most probable dimensions (mm)
1	Simple plate	0.25, 0.75
2	Branches in sector form	1.0
3	Simple stellar form	1.0, 2.5
4	Plate with simple extensions	1.5
5	Broad branches	2.0
6	Ordinary dendritic crystal	2.5
7	Plate with dendritic extensions	3.0
8	Fernlike crystal	4.0
9	Spatial dendritic hexagonal crystal	4.0
10	Spatial dendritic radiating crystal	2.5
11	Needle and combination of needles	1.75
12	Columnar crystal	0.5
13	Combination of bullets	0.75
14	Column with dendritic planes at both ends	1.0, 2.5
15	Column with plates at both ends	0.75
16	Spatial assemblage of plates	0.75
17	Combination of bullets with planes	1.0
18	Combination of columns with or without planes	1.0

1 SOLID PRECIPITATION



TYPE OF PARTICLE	SYMBOL	GRAPHIC SYMBOL
PLATE	F 1	
STELLAR CRYSTAL	F 2	
COLUMN	F 3	
NEEDLE	F 4	
SPATIAL DENDRITE	F 5	
CAPPED COLUMN	F 6	
IRREGULAR CRYSTAL	F 7	
GRAUPEL	F 8	
ICE PELLET	F 9	
HAIL	F 0	

MODIFYING FEATURE	BROKEN CRYSTALS	RIME COATED CRYSTALS	CLUSTERS	WET
SYMBOL SUBSCRIPT	p	r	f	w

SIZE OF PARTICLE D MEASURED IN MILLIMETERS.

Figure I-4. Abstract of the international snow classification. (From ref. 29)

Table I-III. Relation of international snow classification and Nakaya's general classification.

International classification code	
F1	Includes P1a, P1b, P1c, P4 of general classification
F2	P1d, P1e, P1f, P1g, P1h, P1i, P2a, P2b, P2c, P3a, P3b, P4
F3	C1a, C1b, C1c, C2a, C2b
F4	N1a, N1b, N2
F5	P5a, P5b
F6	CP1a, CP1b, CP1c, CP2a, CP2b
F7	CP3, S, I1, I2, I3
F8	R4a, R4b, R4c
F9	Frozen raindrops, called sleet in the U. S. (in England a mixture of snow and rain is called sleet)
F0	Hail is solid precipitation with concentric laminar structure, formed by successive freezing of water layers, usually in thunder clouds
p	Broken crystals of types F1, F2, F3, etc.
r	R1, R2, R3 of Nakaya's classification (not sufficiently coated to be classed as graupel). Clusters of crystals of types F1, F2, F3, etc. (snowflakes)
w	Wet or partly melted particles
D	The greatest dimension of the crystal or particle in millimeters. In the case of a snowflake, the average size of the individual particles.

Freshly deposited snow is initially a loose mass of snow particles, its density depending on the type of snow falling and on the prevailing weather conditions. Dendritic crystals settling gently to earth make a soft fluffy deposit with very low density, whereas small crystals fragmented and driven by strong winds pack into firm, dense layers. After deposition the snow particles change shape and begin to bond together at sub-freezing temperatures as the processes of sublimation and molecular diffusion re-distribute ice and progressively reduce the specific surface of the grains. High air temperatures or strong radiation lead to melting, with consequent transformation of snow to ice when temperatures again fall below freezing. Settlement of the snow under gravity gradually compacts it to higher densities; on polar glaciers snow is transformed to ice in this way without a need for melting and re-freezing.

Frost, rime and glaze

Frost, rime and glaze are all forms of ice deposited directly from the atmosphere onto cold solid objects. Frost (hoar) is formed by direct condensation (sublimation) of water vapor on a cold surface; rime results from rapid freezing of small supercooled water droplets; and glaze is the hard clear ice produced by slow freezing of supercooled droplets.

Frost deposits usually have well-defined crystalline forms and do not adhere strongly to the parent surface. Nakaya[10] differentiates five crystalline forms: (1) needle, (2) feather-like, (3) plate, (4) cup, and (5) dendritic. Crystalline frosts often form from cold still air which is not highly supersaturated; common types can be seen on cold window panes, inside refrigerators, and on the walls and ceilings of unheated undersnow, under-ice, and permafrost cavities.

Rime deposits have a rough, white appearance resulting from entrapped air and they are generally quite firmly bonded to the receiving surface. As small supercooled droplets, such as supercooled fog particles, strike a solid object they freeze instantaneously. Successive collisions build up a layer of rime ice composed of particles which have largely retained their rounded shapes during freezing, so that air is included between them. The rate of deposition is controlled by the liquid water concentration of the air, the temperature, the droplet size, and the rate of collection. The rate at which droplets strike a surface increases as wind speed increases, and so rime builds up faster if there is some wind. Deposition in a wind produces the characteristic feathery growths on the windward side of tree branches, posts, power lines, etc. Rime does not form in high winds when temperatures are relatively high, as the wet-air kinetic temperature rise brings the droplets to a temperature too high for instantaneous freezing.

Glaze is a deposit of hard clear ice, which adheres strongly to the surface where it forms. It is produced by relatively slow freezing of supercooled droplets, and it generally forms at high temperatures and with larger droplets than rime does. As the droplets strike the surface they spread and flow, freezing as they do so. The result is that air is excluded from the deposit, which becomes smooth and rounded from the flowing water. Glaze deposited during windy conditions builds out on the windward side of solid objects, but develops a smooth aerodynamic profile. It is more common than rime at high wind speeds. "Freezing rain" is another form of glaze.

Both rime and glaze can cause serious trouble when they form on aircraft, structures, wires, etc., and so the problems of forecasting and preventing icing of this kind continue to receive attention.

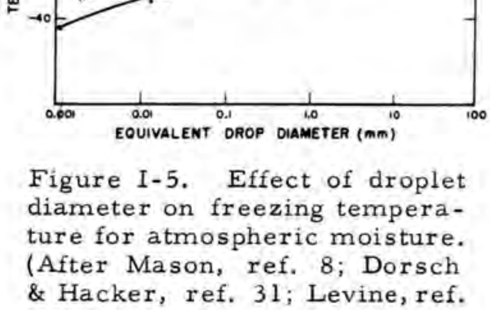


Figure I-5. Effect of droplet diameter on freezing temperature for atmospheric moisture. (After Mason, ref. 8; Dorsch & Hacker, ref. 31; Levine, ref. 32)

Although supercooled water droplets can exist in the atmosphere at temperatures as low as -35C (see Fig. I-5), the probability of icing occurring is greatest when temperatures are around -5C. Pilots use dewpoint spread and air temperature as indicators of possible icing conditions. If air temperature is below freezing and the difference between air temperature and the dewpoint is 2C or less, icing is highly probable. Appleman[1] found empirically that icing in stratiform clouds is probable when air temperature T_a is between the dewpoint T_d and a value equal to $0.8T_d$. There is high probability for temperatures between 0 and -12C, moderate probability between -12 and -22C, and low probability for temperatures lower than -22C.[24]

Glacier ice

Glacier ice is massive perennial ice formed and sustained by accumulation and metamorphism of snow. It is found where the environmental conditions are such that more snow is deposited than can be dissipated by local wastage processes, such as melting and evaporation.

Snow accumulating on the surface, by direct precipitation and by drifting, is gradually transformed into hard impermeable ice by processes of thermal metamorphism and pressure metamorphism.* The snow grains of the initial deposit change their size and shape and develop intergranular bonds as a result of sublimation, which can occur even at very low temperatures. The snow can also be modified by melting and melt-water infiltration; this gives rise to ice layers and ice lenses in the snow pack. The buried snow layers are under pressure from the snow above, and they compress, slowly but continuously, in a viscous manner. Eventually a given snow layer reaches a state of compression where the air spaces between the grains seal off and become bubbles, thus making the material impermeable. This transition from a permeable to an impermeable material is taken as marking the change from snow to ice; it occurs when the density has reached about 0.8 g/cm^3, which corresponds to depths of 150-300 ft in the cold snow areas of ice caps. The rate of compression, or densification, is governed by the initial density of the surface snow, the temperature, and the rate of snow accumulation at the surface; these variables determine the depth at which the change from snow to ice occurs.

A mass of glacier ice cannot build up indefinitely, since the stresses developed by its weight cause it to flow. This flow, which may be termed visco-plastic flow, permits a balance between accumulation and wastage to be reached by transporting ice either directly to the coast, where it is discharged in icebergs, or into warmer zones, where it is dissipated by melting.

Lake ice

The surface of a lake freezes when heat losses from the surface (radiation, convection, evaporation) exceed the heat gains (air convection, radiation, condensation, geothermal flux). As a fresh water surface cools, a stable thermal stratification is set up, since the density of fresh water decreases as temperature falls below +4C. Therefore, in the absence of mixing currents, the surface can freeze over while the main body of water is at temperatures above freezing.

When the surface water cools to 0C, or a fraction of a degree lower, tiny spicular and platelike crystals are formed, and aggregations of these initial crystals eventually produce a continuous ice cover. Weather conditions prevailing at the time of formation affect the crystalline structure of the ice: calm weather favors orderly crystal growth, with optic axes of the crystals predominantly parallel, while wind and waves (and snowfall) tend to produce a random pattern of primary crystals.[27]

Soluble impurities in the water are rejected during the freezing of primary crystals, and they eventually become concentrated along the boundaries of adjacent crystals. This leads to preferential melting along crystal boundaries when the ice is exposed to strong solar radiation during the following spring, and the ice "candles".

Once the lake has frozen over, the growth rate of the ice is controlled by the rate of heat conduction through the sheet. Rate of conduction depends on the ice thickness, its thermal conductivity, and the temperature difference between the upper and lower faces. Unless the lake is entirely frozen the lower face is always at the freezing point, but the temperature of the upper surface depends on air temperature, net radiation, wind velocity, and snow cover (which acts as an insulator). Of these, air temperature has the greatest influence, and it is possible to relate ice thickness with the amount of "cold" the surface has been subjected to since freeze-over. The "cold" is expressed in "degree-days of frost", obtained by taking a summation of mean daily negative air temperatures over the period since freeze-over. The relationship is of the form

$$h = K\sqrt{S}$$

where $\underline{h}$ is ice thickness, $\underline{S}$ is the number of degree-days of frost since freeze-over, and $\underline{K}$ is a constant which takes into account snow cover, wind conditions, and size and situation of the lake (see Assur, ref. 2, for typical values of $\underline{K}$).

*Lying snow which is undergoing metamorphism is often called "firn".

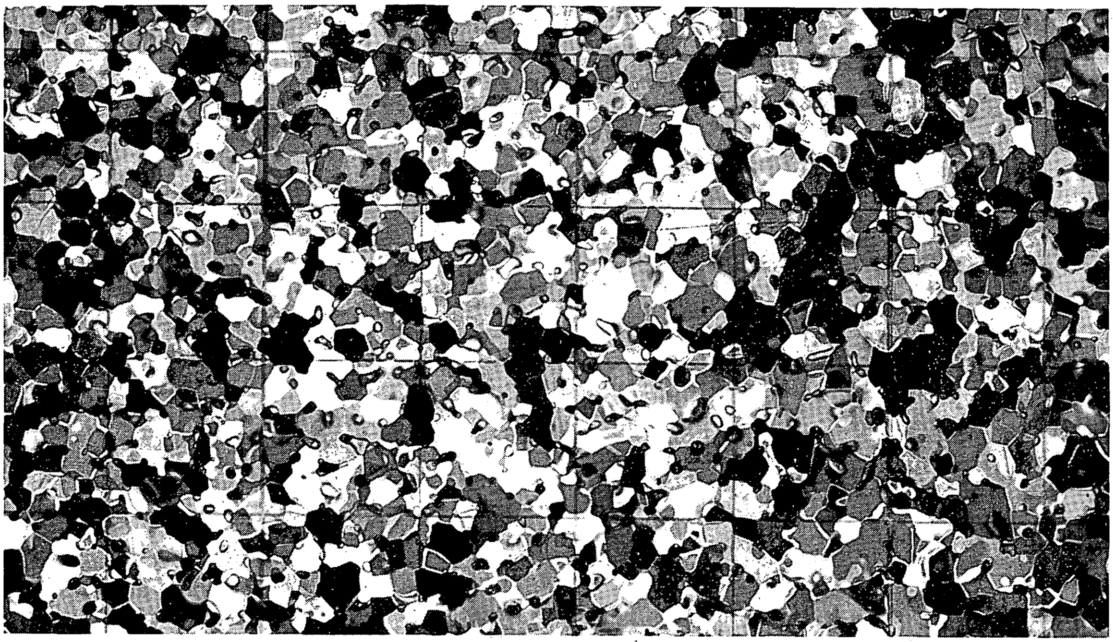


Figure I-6. Thin section of glacier ice formed in a cold, non-melting environment (Byrd Station, Antarctica). Crystal boundaries and air bubbles can be seen in the section, which was taken from a depth of 66 m (216 ft). Scale is provided by the 1-cm grid overlay. (Photo by A. J. Gow)

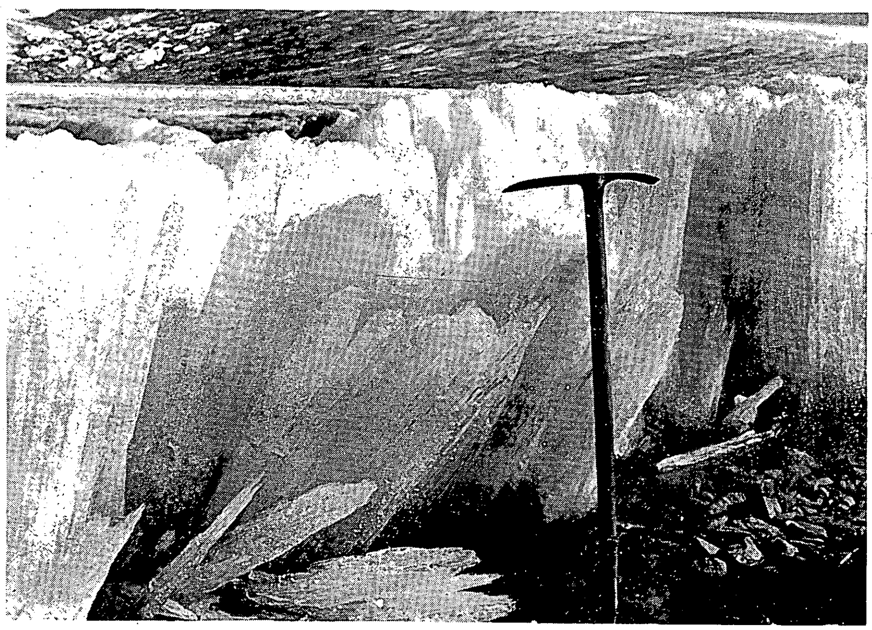


Figure I-7. Candled lake ice. The columnar structure results from radiation melting at the grain boundaries. Each "candle" is a single crystal. (CRREL photo).

River ice

In the more placid reaches of a river, an ice cover forms in much the same way as it does on a lake, although there is some vertical mixing of the water and therefore freeze-over is somewhat later than on a lake in the same area. In turbulent water, however, a stable ice cover is unable to form and supercooling occurs when air temperatures fall rapidly. This leads to the formation of tiny ice crystals which mix intimately with the turbulent water, but mat together on encountering a solid object or after being carried into calm water. The spongy matted deposits of this kind of ice bear some resemblance to forge cinders, and this led to the French-Canadian name of frazil.

Frazil forms, usually at night, in rapids, sluice tail-races, etc. when the water is supercooled by cold, rapidly falling air temperatures. Very little supercooling is required for its formation — only about 0.01C below zero — but unless the water is exceedingly turbulent air temperatures have to fall rapidly to achieve this. Cooling rates higher than 0.01C/hr are normally a necessary condition, and frazil usually forms only in the absence of sun.[18] Frazil can also form on water bodies which are normally placid if the surface becomes agitated by wind.

Initial formation occurs at the surface of the water, where thin plates crystallize. These primary platelike crystals have a ratio of diameter to thickness of about 50:1, and the initial diameter is less than 1 mm. After they have grown to diameters in excess of 1 mm, secondary dendritic growth may occur around the edges.[18]

The crystals have little buoyancy, and are easily carried below the surface by turbulent eddies. The vertical distribution of frazil crystals is governed by the turbulent exchange, which varies with stream velocity. With surface velocities greater than 10 ft/sec there is a fairly even distribution of frazil through the whole stream depth; with velocities of about 4-10 ft/sec crystals are carried mainly in the upper layers; and with velocities less than 4 ft/sec most of the frazil concentrates near the surface and tends to mat into a continuous sheet of ice.[21, 25]

Frazil can cause serious problems on rivers. At hydroelectric plants it clogs the intakes, and elaborate precautions have to be taken to prevent and combat its formation on the screens. Frazil runs also collect under sheets of surface river ice, restricting flow and causing ice jams and floods. Serious flood damage has been caused by frazil forming a frozen foam below an overfall spillway which had supercooled water on the very still upstream side. (See III B, "Ice Engineering".)

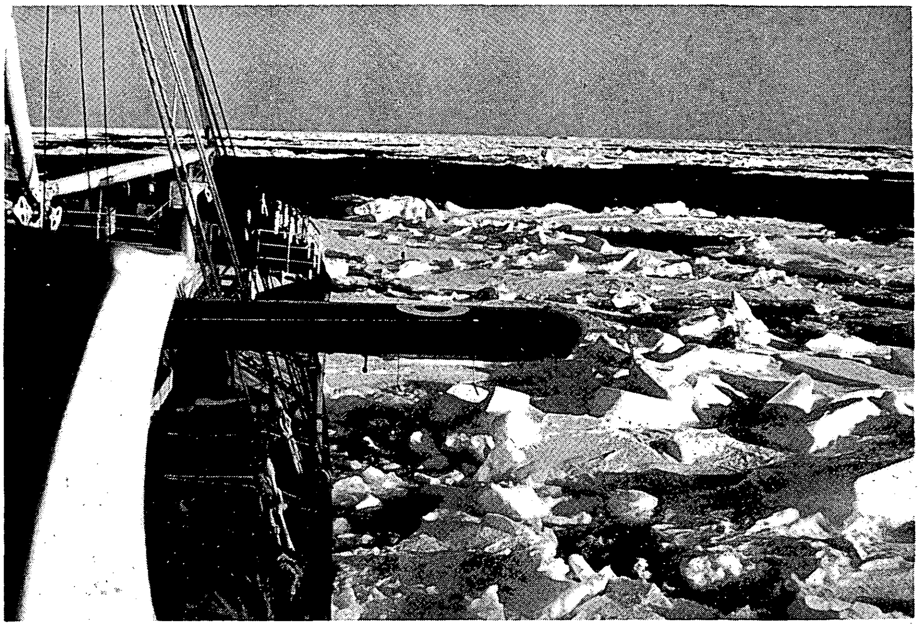
Sea ice

Ice encountered on the sea is of two main types: true sea ice, formed by freezing of the sea's surface water, and glacier ice, formed by snow metamorphism on land or on an ice shelf. Icebergs are of the second type; they should not be included in the term 'sea ice', and they are discussed separately.

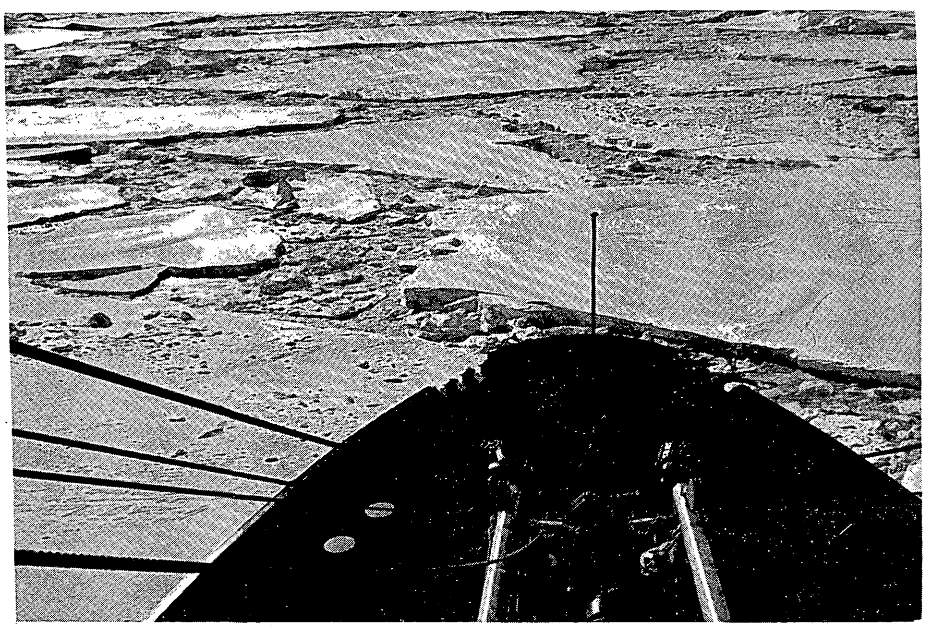
Sea ice forms in continuous sheets when the sea freezes in the autumn and early winter. Along coastlines in the Arctic and Antarctic it usually remains firmly anchored to the land during the winter and early spring, and it is broken only by narrow tide cracks formed as the ice rises and falls with the tides. In this state it is called 'fast ice', and after one winter's growth it reaches a thickness of 3 to 8 feet in Arctic and Antarctic regions.

In spring or summer sea ice breaks up under the action of wind and waves (after being weakened by high temperatures and melting) and large flat segments, termed 'floes', drift with wind and current. This is 'pack ice'; it may take the form of discrete floes drifting freely with open water between them, or it may be a mass of floes packed tightly together or even thrust over each other (rafted) by the pressure of wind or current. Terminology for various forms of sea ice is given in Chapter 10.

Pack ice may disperse and melt in the open ocean, or it may remain and refreeze in a matrix of new ice the following winter. Much of the ice of the Arctic Ocean survives for many years, reaching an equilibrium thickness when summer melting equals winter accretion.



(a) Open pack ice. The floes are well separated by open water, and ice-strengthened ships can move freely. Some evidence of old rafting can be seen in the foreground.



(b) Large ice floes cemented together by lightly frozen brash ice.

Figure I-8. Pack ice.

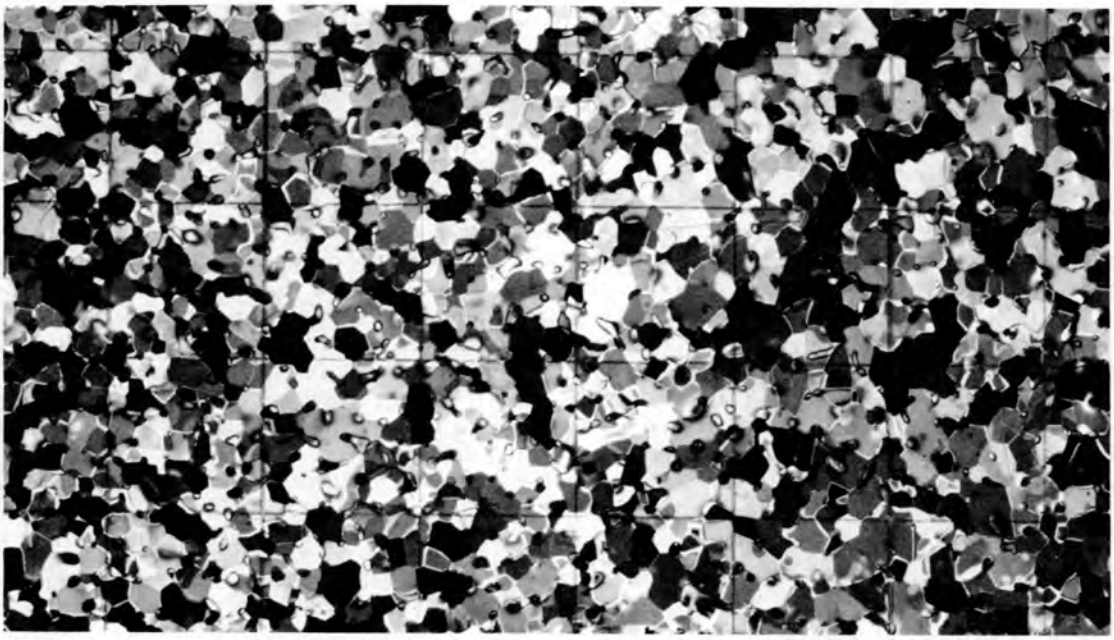


Figure I-6. Thin section of glacier ice formed in a cold, non-melting environment (Byrd Station, Antarctica). Crystal boundaries and air bubbles can be seen in the section, which was taken from a depth of 66 m (216 ft). Scale is provided by the 1-cm grid overlay. (Photo by A. J. Gow)

Figure I-7. Candled lake ice. The columnar structure results from radiation melting at the grain boundaries. Each "candle" is a single crystal. (CRREL photo).

(c) A ship trapped in heavy, rafted pack ice.

Figure I-8. (Cont'd) Pack ice.

Land-fast sea ice which remains in place for more than one year can also grow to great thickness. This type is termed bay ice.

The freezing of salt water is different from the freezing of fresh water. Sea water of 35‰* salinity begins to freeze at about -1.9C, but water of this salinity does not reach maximum density until its temperature has been lowered to -3.6C. This produces density currents in the water as the surface cools, but since surface heat loss is often more rapid than convective heat transfer from the depths, ice may form before the deep layers have been cooled to the freezing point.[15,22]

Initially, free-floating crystals form at the surface. These are usually platelike, with widths very much greater than their thicknesses. In calm conditions the initial crystals soon mat together to form a continuous sheet, the c-axes tending to have initially a preferred vertical orientation.[26] In rougher water the crystals are agitated and freeze-over is delayed until a heavy, slushy aggregation of crystals has been formed (the distinctive "pancake ice" is formed under these conditions).

In the initial crystal formation, salt is rejected during freezing, leaving the crystals "fresh". When the ice mats into a sheet, however, brine is included between and within the crystals, the amount trapped increasing with rate of freezing. During subsequent ice growth, brine concentrates at crystal boundaries and along selected planes with the crystals. Over a period of time, concentrated brine trapped in pockets within the ice migrates to the base of the sheet, primarily by gravity drainage in the summer, but also as a result of diffusion along the temperature gradient during winter. This leads to gradual freshening of the ice, to the point where it may become potable.

Icebergs and ice islands

Icebergs are large pieces of glacier ice which have broken off from land-based ice or from an ice shelf. They range in size from pieces as big as a house to enormous flat-topped slabs tens of miles long and hundreds of feet thick (occasionally even bigger). Their buoyancy in sea water depends on the average density of the ice; dense

* ‰ = parts per thousand, by total weight

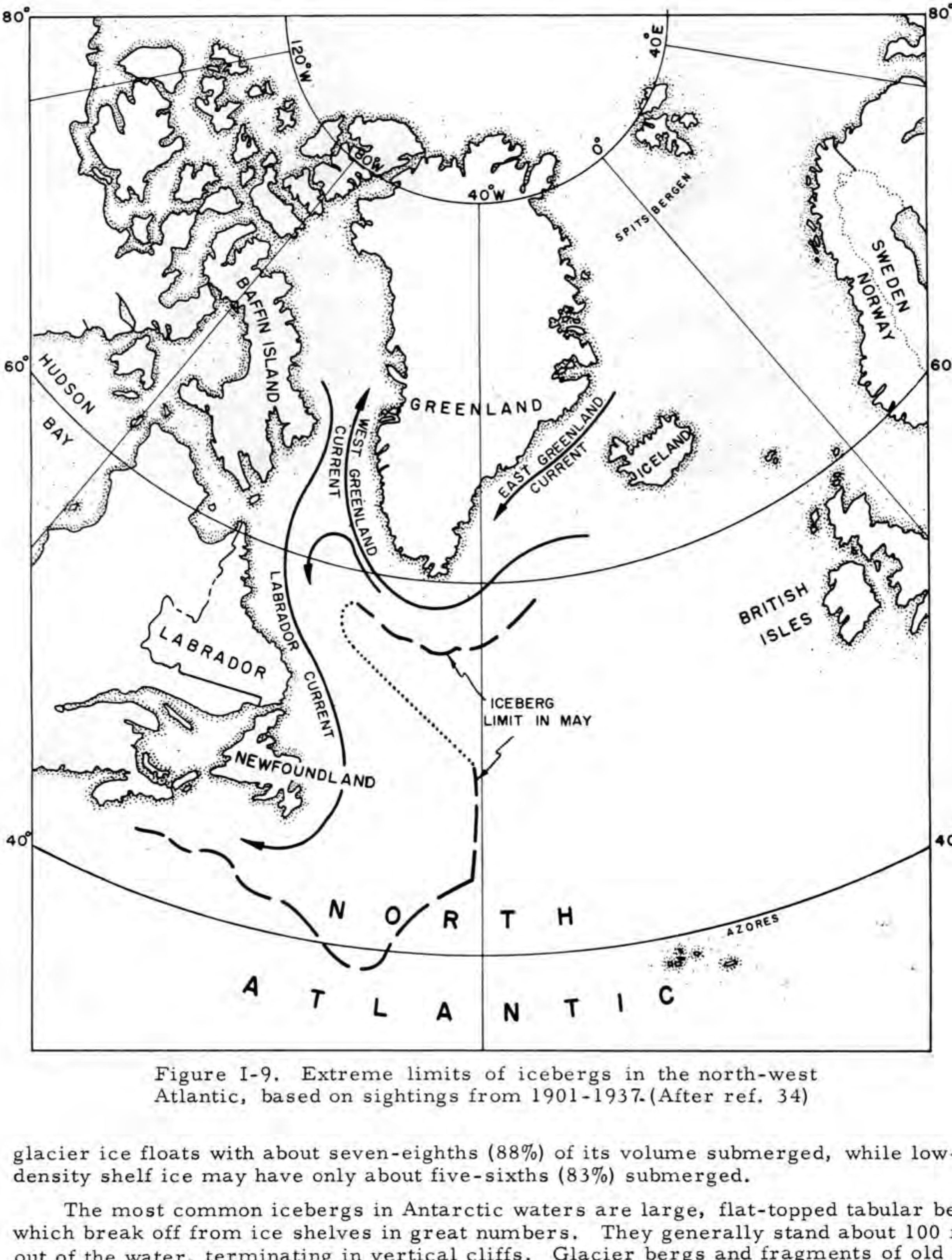


Figure I-9. Extreme limits of icebergs in the north-west Atlantic, based on sightings from 1901-1937. (After ref. 34)

glacier ice floats with about seven-eighths (88%) of its volume submerged, while low-density shelf ice may have only about five-sixths (83%) submerged.

The most common icebergs in Antarctic waters are large, flat-topped tabular bergs, which break off from ice shelves in great numbers. They generally stand about 100 ft out of the water, terminating in vertical cliffs. Glacier bergs and fragments of old tabular bergs are also encountered in large numbers. Arctic icebergs come mainly

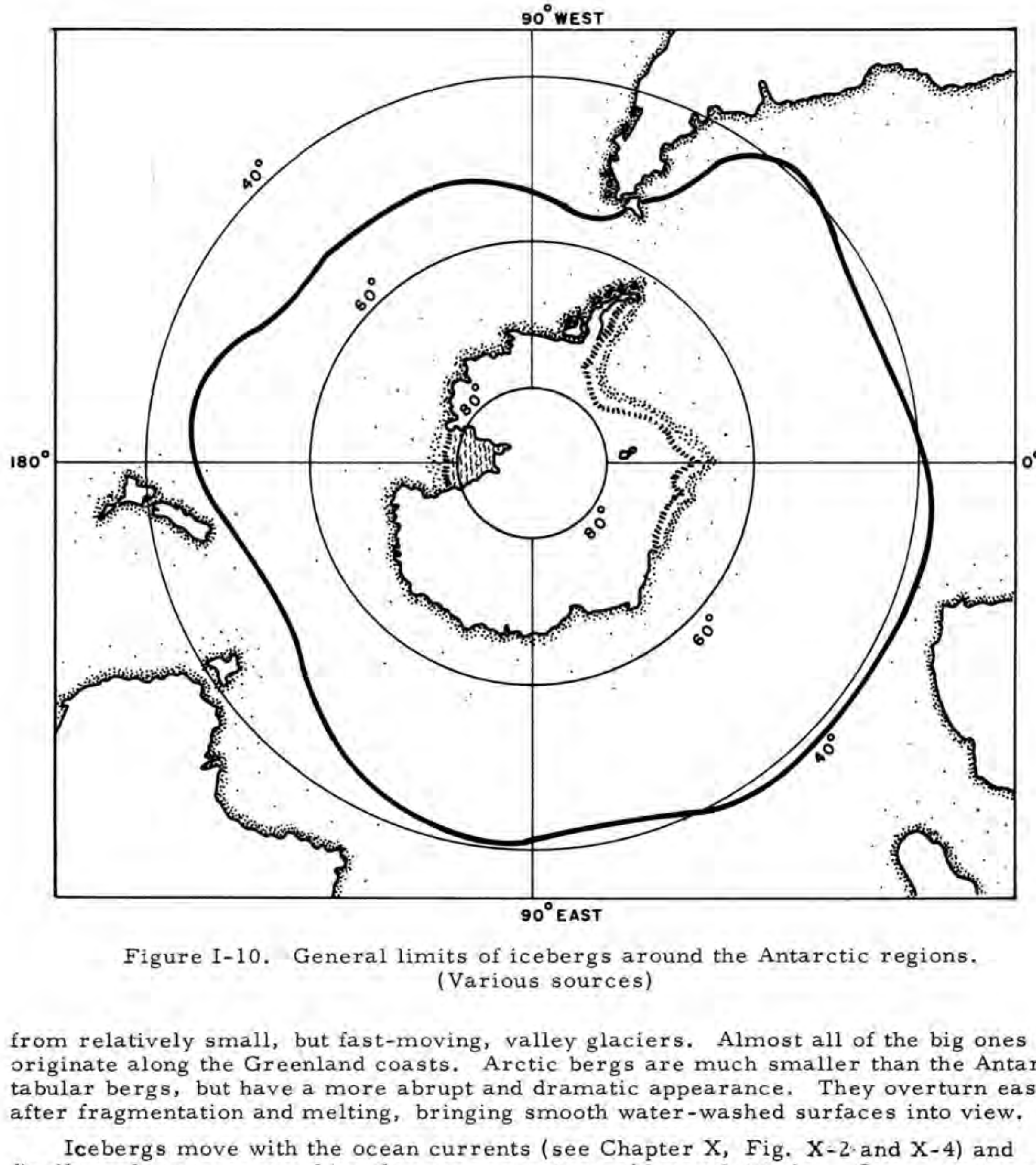


Figure I-10. General limits of icebergs around the Antarctic regions. (Various sources)

from relatively small, but fast-moving, valley glaciers. Almost all of the big ones originate along the Greenland coasts. Arctic bergs are much smaller than the Antarctic tabular bergs, but have a more abrupt and dramatic appearance. They overturn easily after fragmentation and melting, bringing smooth water-washed surfaces into view.

Icebergs move with the ocean currents (see Chapter X, Fig. X-2 and X-4) and finally melt away on reaching the warmer waters of lower latitudes. Some run aground on submarine banks close to their place of origin and remain there for many years, sometimes receiving nourishment from snowfall and deforming into domelike shapes. Since they are deeply immersed, icebergs can move in a different direction than surrounding pack ice, which is more sensitive to surface winds. In the Southern Ocean icebergs present few problems, since shipping in high latitudes is confined to whalers and expedition ships, and the bergs never reach the well-traveled shipping lanes further north. In the North Atlantic the situation is quite different and fog-shrouded bergs were enough of a hazard to shipping for the International Ice Patrol to be brought into existence (as a result of the 'Titanic' disaster in 1912). Radar alleviates the situation but does not eliminate the danger.

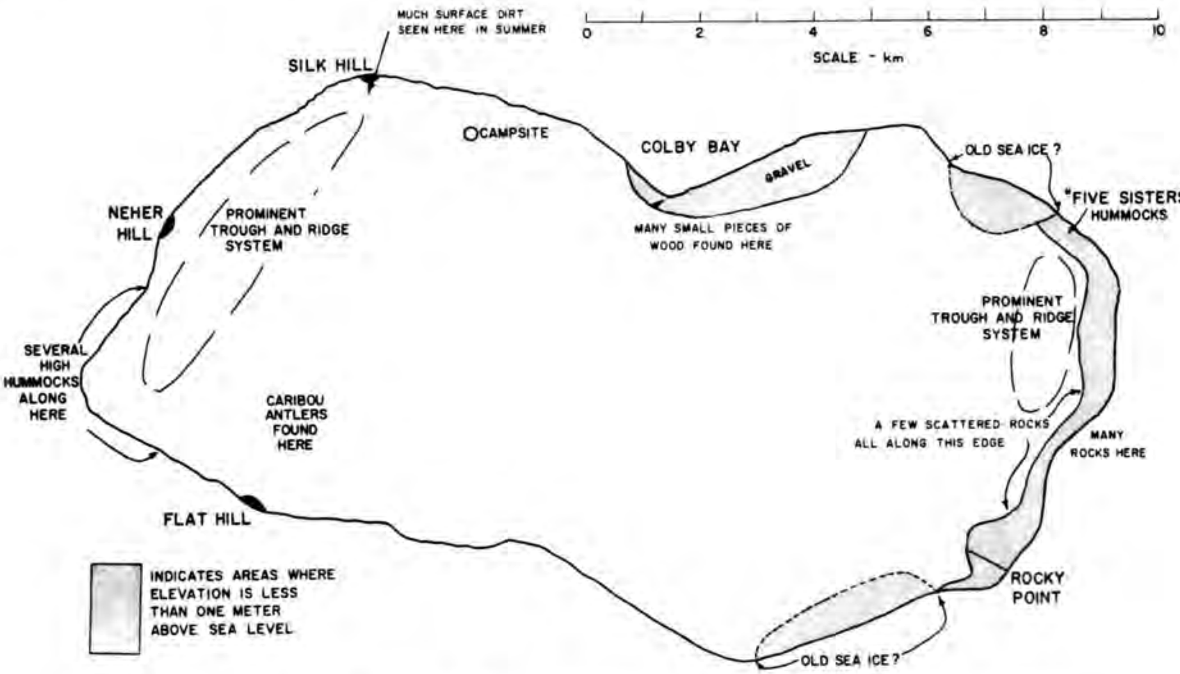


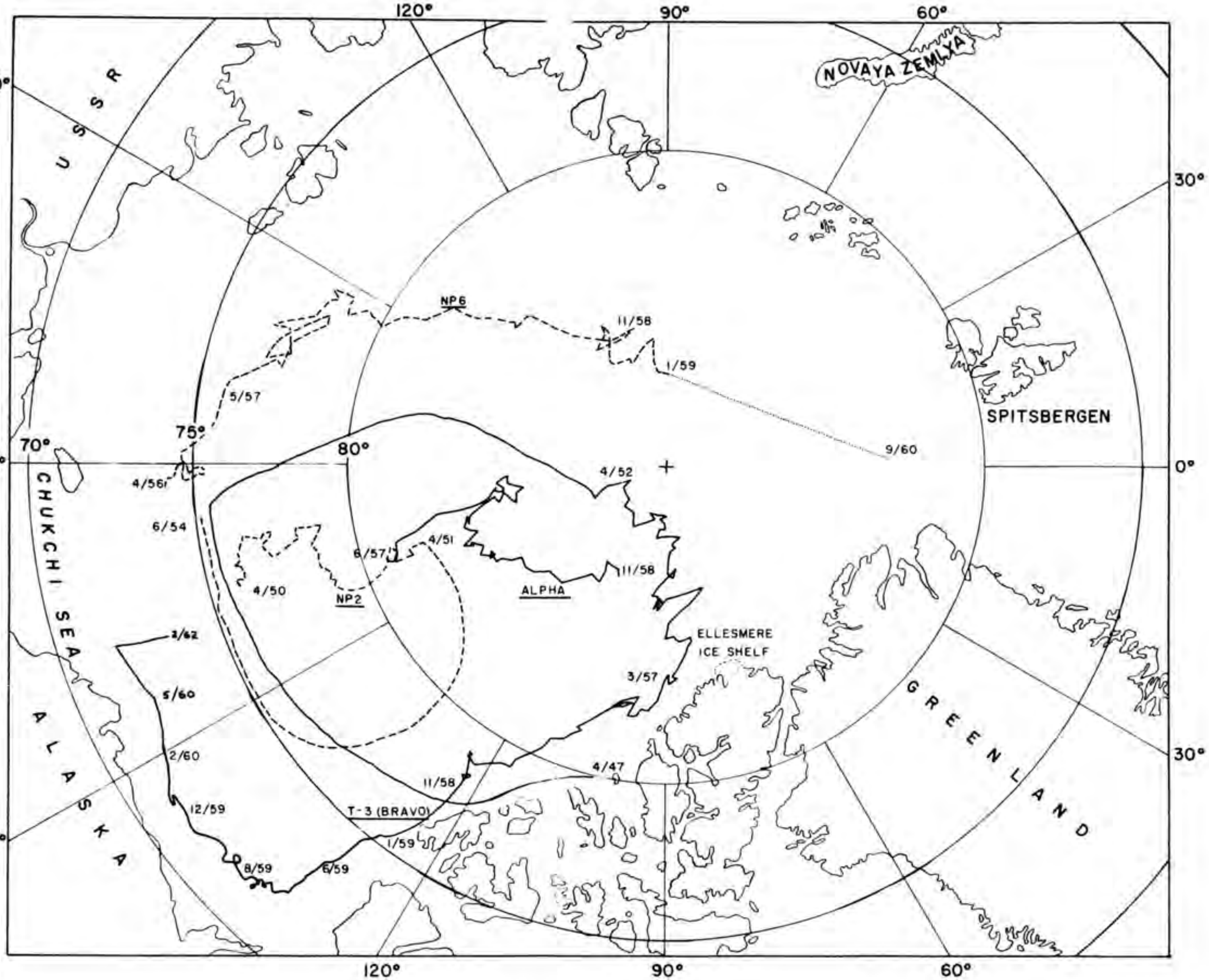
Figure I-11. Map of the floating ice island T-3 prior to the break-up which reduced its size about 1961. (After Crary, ref. 5)

The floating ice islands of the Arctic Ocean are large tabular icebergs, up to 200 sq miles in area. Several ice islands in the North American sector of the Arctic Ocean are kept under observation, and some of them have been occupied by scientific parties from time to time. These big slabs have broken away from the ice shelves on the north coast of Ellesmere Island and they drift in clockwise paths on the Alaskan side of the Lomonosov Ridge. The roughly elliptical paths run westerly though the northern parts of the Beaufort and Chukchi Seas, swing north to pass over the Pole, and return to the north of Ellesmere Island. The islands drift in a direction about 30°-40° to the right of the wind direction, and have speeds from 1% to 2% of the surface wind speed. Before it went aground near Barrow, Alaska, the island designated T-3 had an average drift rate of about 1 nautical mile per day.

T-3 was found to be about 160 ft thick, and to consist of a basement of glacier ice and old sea ice covered by 80-90 ft of granular iced firn, with interspersed lenses of meltwater ice. The ice masses on Ellesmere Island, where it originated, are estimated to be 3000-5000 years old. The surface of T-3 is characterized by alternating parallel troughs and ridges, the crests being spaced at intervals of about 300 ft, with the troughs up to 20 ft deep. Meltwater lies in most of the troughs. Surface ablation varied with the latitude of the island; net loss was small in the region of the Pole (where there may even be net accumulation), but farther south annual net ablation losses of more than 1 ft were measured.[5]

Ice islands are big and strong enough to survive for many years while drifting in the Arctic Ocean, and quite elaborate research stations have been built upon them. These are usually established and maintained by large ski-aircraft. Station Bravo (on T-3), Arlis II, and the Russian Stations SP-6 and SP-8 are examples.* Other drifting stations in the Arctic have been established on thick floes of true sea ice, which have a much shorter life. The average operational life of a drifting station on a floe is about

* SP — Severnyy Polius = North Pole



——— U. S. DRIFTING STATIONS - - - - - - RUSSIAN DRIFTING STATIONS 6/57 - MONTH & YEAR

Figure I-12. Tracks of some U. S. and Russian drifting stations (compiled from various sources).

$1\frac{1}{2}$ years, and the longest period of occupancy is 3 years. Examples of drifting stations on floes are the U. S. stations Alpha, Arlis I, and Charlie, and the Russian stations SP-1 through SP-5.

The tracks of some drifting stations are shown in Figure I-12.

Ground ice

In the permafrost regions, which cover great areas of the earth's surface, ice is found beneath the ground surface, and sub-freezing temperatures persist to very great depth in some places. Where mean annual temperatures are well below freezing, at high latitudes and on high mountains, ice can persist indefinitely in the ground since summer warmth only causes melting to shallow depths.

Ground ice can be divided into two broad categories: ice which has formed within the soil or rock, and ice which has been formed on the surface and subsequently buried ('relict' or 'fossil' ice).

For ice formed within the soil, four major sub-types can be distinguished:

(a) Ice lenses. These are produced by migration of water to a freezing front, very noticeable in silty soils.

(b) Ice in pores. This is produced by the freezing *in situ* of natural pore water in the soil or rock. In unsaturated granular soils it may appear as ice crystals on the mineral particles.

Types (a) and (b) occur also with seasonal freezing in temperate regions.

(c) Ice in pingos and other ice mounds (Fig. I-14a). This results from the freezing of water forced in under hydrostatic pressure.

(d) Wedge ice (Fig. I-14b). This is produced by condensation of moisture and percolation of running water into thermal contraction cracks.[33] The cracks filled by ice wedges can be single ones or can form a polygonal network (Fig. I-14c).

Buried ice consists of massive bodies, which may be the remains of snow beds, icings, glacier ice, sea, lake or river ice. These ice masses may have been buried by water-borne deposits, earth flows and slides, wind-blown material, or rock which has fallen from scree slopes.

Corte[4] has shown that a relationship exists between some ground ice types and patterned ground (morphological features of the surface) in the Thule area of Greenland; besides ice wedge polygons there are features associated with other kinds of ground ice. Ice masses or lenses are generally associated with areas of sorted soils (sorted nets, sorted circles). Elevations and depressions, or furrows, are related to a particular kind of buried ice he calls relict ice. Corte also demonstrates that the physical properties of the ice crystals (size, shape, c-axis orientation, bubble and particle inclusions) are important elements for the determination of ground ice type and origin.

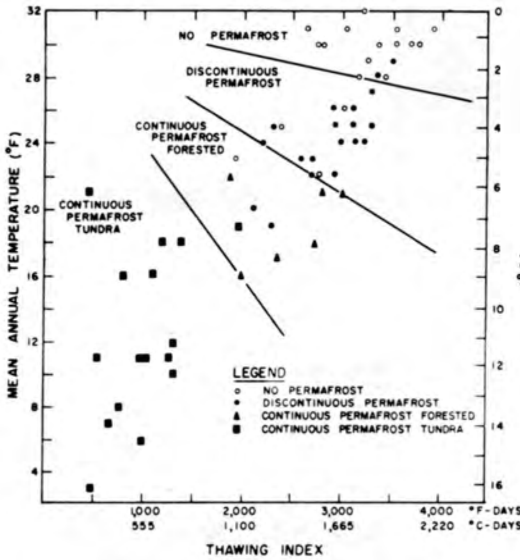


Figure I-13. An approximation of probable permafrost occurrence, from mean annual temperatures and thawing indexes for 61 localities in northern Canada. (After Pihlainen, ref. 14)

(a) A pingo on the coastal plain east of the Mackenzie River delta. (ACFEL ref. 35)

Figure I-14. Ground ice.

(b) An ice mass in glacial till revealed by excavation. The building above has been erected on a fill of free-draining stony material, which prevents the ground from thawing.

(c) Soil polygons on the Arctic coastal plain (Frost, ref. 36)

Figure I-14. (Cont'd.) Ground ice.

BIBLIOGRAPHY

1. Appleman, H. S. (1954) Design of a cloud-phase chart, Bulletin of the American Meteorological Society, vol. 35, no. 5.

2. Assur, A. (1956) Airfields on floating ice sheets, U. S. Army Snow, Ice and Permafrost Research Establishment, Corps of Engineers, Technical Report 36.

3. Corte, A. E. (1961) The frost behavior of soils, Part I: Vertical sorting, U. S. Army Cold Regions Research and Engineering Laboratory, Research Report 85.

4. ________ (1962) Relationship between four ground patterns, structure of the active layer, and type and distribution of ice in the permafrost, USA CRREL Research Report 88.

5. Crary, A. P. (1958; 1960) Arctic ice island and ice shelf studies, Parts I and II, Arctic (Arctic Institute of North America).

6. Kuroiwa, D. (1958) "Icing and snow accretion," in Microwave propagation in snowy districts. Research Institute of Applied Electricity, Hokkaido University, Sapporo, Japan.

7. Lewis, W. (1951) "Meteorological aspects of aircraft icing," in Compendium of meteorology. Boston: American Meteorological Society.

8. Mason, B. J. (1958) The growth of ice crystals from the vapor and the melt, Advances in Physics, vol. 7, no. 26.

9. ________ (1961) The growth of snow crystals, Scientific American, vol. 204, no. 1.

10. Nakaya, U. (1951) "The formation of ice crystals," in Compendium of meteorology. Boston: American Meteorological Society.

11. ________ (1954) Snow crystals, natural and artificial. Cambridge: Harvard University Press.

12. ________ (1954) Formation of snow crystals, USA SIPRE Research Report 3.

13. ________ (1959) Six lectures on snow and ice, presented on radio from the Sorbonne, Paris.

14. Pihlainen, J. A. (1962) An approximation of probable permafrost occurrence, Arctic, vol. 15, no. 2.

15. Pounder, E. R. (1961) "The physics of sea ice," in An analysis of environmental factors affecting army operations in the polar basin. Arctic Institute of North America (for CRREL).

16. Rikhter, G. D. (1945) Snezhnyi pokrov, ego formirovanie i svoistva (Snow cover, its formation and properties). Moscow-Leningrad: Izdatel'stvo Akademii Nauk SSSR, SIPRE Translation 6, 1954.

17. ________ (1960) Geografiia snezhnogo pokrova (Geography of the snow cover). Moscow: Izd-vo Akad. Nauk, SSSR.

18. Schaefer, V. J. (1950) The formation of frazil and anchor ice in cold water, Transactions of the American Geophysical Union, vol. 31, p. 885-893.

19. ________ (1951) "Snow and its relationship to experimental meteorology," in Compendium of meteorology. Boston: American Meteorological Society.

20. Shumskii, P. A. (1959) "Ground ice," in Osnovy geokriologii. Moscow: Izd-vo Akad. Nauk SSSR.

BIBLIOGRAPHY (Cont'd)

21. Strel'tsov, A. M. (1951) Shuga i protivoshugovye meropriiatiia na gidroelektrostantsiiakh (Frazil ice and countermeasures employed on hydroelectric stations), Gidrotekhnicheskoe Stroitel'stvo, vol. 20, no. 4.

22. Sverdrup, H. U. (1956) "Arctic sea ice," in The dynamic north. U. S. Navy.

23. _______________, Johnson, M. W. and Fleming, R. H. (1942) The oceans, their physics, chemistry and general biology. New York: Prentice-Hall.

24. U. S. Air Force (1960) Forecasters' guide on aircraft icing, Air Weather Service Manual.

25. Vartazarov, S. IA. (1939) Metody kolichestvennogo ucheta shugi (Quantitative methods of calculating the amount of frazil ice), Meteorologiia i Gidrologiia, vol. 5, no. 9. p. 102-109.

26. Weeks, W. F. (1958) "The structure of sea ice: a progress report," in Arctic sea ice. U. S. National Academy of Sciences, National Research Council, Pub. 598.

27. Wilson, J. T., Zumberge, J. H. and Marshall, E. W. (1954) A study of ice on an inland lake, USA SIPRE Technical Report 5.

28. U. S. Navy (1957-58) Oceanographic atlas of the polar seas, Part I: Antarctic, Part II: Arctic. H.O. publication no. 705.

29. International Association of Hydrology (1954) The international classification for snow, Commission on Snow and Ice.

30. Ragle, R. H. (1962) An investigation of the formation of lake ice in a temperate climate, USA CRREL Research Report 107.

31. Dorsch, R. G. and Hacker, P. T. (1950) Photomicrographic investigation of spontaneous freezing temperatures of supercooled water droplets, National Advisory Committee for Aeronautics, Technical Note 2142.

32. Levine, J. (1950) Statistical explanation of spontaneous freezing of water droplets, National Advisory Committee for Aeronautics, Technical Note 2234.

33. Lachenbruch, A. H. (1962) Mechanics of thermal contraction cracks and ice-wedge polygons in permafrost, Geological Society of America, Special Paper 70.

34. U. S. Navy Hydrographic Office (1946) Ice atlas of the Northern Hemisphere, Washington: H. O. publication no. 550

35. Arctic Construction and Frost Effects Laboratory (1962) Airphoto pattern reconnaissance of Northwestern Canada, prepared for Office of the Chief of Engineers, Civil Engineering Branch, Engineering Division, Military Construction.

36. Frost, R. E. (1950) Evaluation of soils and permafrost conditions in the territory of Alaska, vol. I, prepared by Engineering Experiment Station, Purdue University, for St. Paul District, U. S. Army Corps of Engineers

CHAPTER II. GLACIERS

Terminology

The general term "glacier" covers the complete range of snow-fed bodies of land ice and floating ice shelves, from small permanent snow beds to ice sheets of continental proportions. It is also frequently used, without qualification, to describe small dependent components of a major glacier system. In the present monograph the term "glacier," or "glacier system" will be used for a complete and independent glacier, and dependent components will be described by a qualifying adjective, e.g., piedmont glacier. Some terminology is given below.

Ice sheet. A continuous sheet of ice of immense size. It submerges the land mass almost completely, and flows outwards in several directions. An ice sheet may have almost every other type of glacier among its components. The only present-day glaciers to which the term is commonly applied are the ice sheets of Greenland and Antarctica.

Ice cap. A broad mass of ice, considerably smaller than an ice sheet, which flows outwards in several directions and submerges all but the major topographic features of the underlying land. Examples of ice caps can be found on Baffin Island (e.g., Penny Ice Cap), in Iceland (e.g., Vatnajökull), in Alaska (e.g., Juneau Ice Field), and in Norway (e.g., Jostedalsbrae).

Valley glacier. A glacier flowing down a valley bounded by exposed rock. It may be an independent glacier of the alpine type, or an outflow glacier from an ice sheet or ice cap.

Cirque glacier. A localized glacier occupying a bowl-shaped niche (cirque, cwm, corrie) on a mountainside or at the head of a valley. Common in Norway and the Alps.

Piedmont glacier. An apron of ice formed when valley glaciers issue onto low-lying areas, spread, and unite. A piedmont glacier is not an independent glacier. The Malaspina Glacier, Alaska, is an example. (When a single glacier debouches from a valley and spreads on flat land, the spreading portion is called a foot-glacier, or expanded-foot glacier).

Ice shelf. A thick and extensive slab of ice floating on the sea, but attached to the coastline of a glacier-covered land mass. It is fed by inflow of adjacent land glaciers and by snow accumulation on its surface. It is not an independent glacier. Examples are the Ross and Filchner Ice Shelves in Antarctica, and the Ward Hunt Ice Shelf of Ellesmere Island.

Ice stream. A course along which flow in an ice sheet is locally rapid. The ice stream flows much faster than the surrounding ice mass, but there are often no visible rock boundaries. Ice streams are common in Antarctica. The term is also used by some to denote the distinct longitudinal streams, separated by medial moraines, in a compound valley glacier. Each of these streams is donated by a separate tributary glacier. These features can be seen on most major valley glaciers.

Rock glaciers. Streams or lobes of rocky material, with interstitial ice, which bear a resemblance to common glaciers. They exist in some cirques and valleys of mountains which are free of permanent snow cover, and are apparently formed when intimate mixtures of scree, boulders, and ice are preserved as a type of permafrost on steep slopes. The active ones move sluggishly with speeds of the order of a few feet per year.

Description and classification

In the past a number of schemes for classifying glaciers have been put forward.[25] None of them is entirely satisfactory when used alone, but collectively they offer a means of describing individual glaciers. Clear description is probably preferable to rigid classification, however, since a comprehensive classification system might turn out to have as many subdivisions as there are glaciers.

Figure II-1a. Valley glacier in a region of coastal mountains, East Greenland. The picture shows many glacial features of the ice and the topography.

Figure II-1b. Valley glaciers (outflow glaciers) flowing down to the sea from the Greenland Ice Cap.

Figure II-1c. A lobe of ice spreading on flat ground where a valley glacier debouches from the hills (see "Piedmont glacier").

Morphological description. For description of form and size a simple scheme might be adapted from Ahlmann's classification[1].

A. Extensive continuous ice sheets moving outwards in several directions

(1) Major ice sheet of very large area (10^6-10^7 km^2), e.g., Antarctica, Greenland.

(2) Minor ice sheet, or ice cap, very much smaller than A(1) (say 10^3-10^4 km^2) e.g., North-East Land, Ellesmere Island, Vatnajökull (Iceland).

B. Glaciers with defined courses and unidirectional flow, both independent glaciers and outflow glaciers from group A types.

(1) Valley glacier, e.g., Alpine glacier, Greenland outflow glacier.

(2) Ice stream, along which flow in an ice sheet is locally rapid, often with no visible rock boundaries (common in Antarctica and Greenland).

C. Glacier ice spreading out over flat land, or on water, at the edge of a glacierized region (dependent components of a glacier system).

(1) Piedmont glacier, formed when valley glaciers issue onto low-lying areas, spread, and unite, e.g., Malaspina Glacier, Alaska.

(2) Ice shelf, floating on the sea and fed by land-formed glaciers and snowfall, e.g., Ross and Filchner Ice Shelves.

The distribution of area with altitude can be described by means of hypsometric curves or area distribution curves[1,4,5] On a hypsometric curve, surface altitude is plotted against cumulative percentage of the total surface area which lies below that level. Examples of hypsometric curves are given in Figure II-2. To obtain area distribution curves, Ahlmann[1] divided the total altitude range of the glacier into ten parts and plotted the percentage of the surface area which lay in each tenth of the altitude range. Examples are shown in Figure II-3. Ahlmann and Bauer have used area distribution curves to classify glaciers[1,4,5] but the usefulness of this procedure seems questionable (widely different glacier types may yield similar curves).

Description of physical characteristics. Further description of a glacier can be made on the basis of detectable characteristics of the upper layers, such as temperature regime, surface snow or ice types, and rate of movement.

There is a significant difference between snow and ice at subfreezing temperatures and snow and ice at the melting point, since heat transfer and metamorphic processes change when temperature reaches the melting point. "Warm" (0 C) ice is therefore differentiated from "cold" (below 0 C) ice. It is, however, unwise to generalize and say that an entire glacier is "warm" or "cold", since temperatures vary with surface altitude, with depth below the surface, and with the time of year (see Chapter VIII). The old classifications of glaciers by temperature are perhaps best avoided.

The condition of surface snow or ice offers another means of describing glacier zones. Most glaciers suffer a net loss of ice by melting and evaporation in their lower parts, which are known as ablation zones. Higher up there may be an annual surplus of snow, but this will be soaked with summer melt water which later refreezes. Further up still there is a definite accumulation of snow, but small quantities of melt water may still percolate into it. At the highest levels, if the climate is cold enough, there is no melting at all and the snow accumulates in dry-cold conditions. Benson[7] designates these characteristic areas as the ablation zone, the soaked zone, the percolation zone, and the dry snow zone; his scheme is illustrated in Figure II-4a. The transition between the ablation zone and the higher accumulation zones is termed the firn limit.

The rapidity or sluggishness of a glacier's movement is a useful characteristic to describe. Measurements of surface flow rate are needed for quantitative description, but some qualitative estimation of movement can be made by observing crevasses, "flow lines", slopes, moraines, etc.

Some indication of a glacier's mass economy is given by describing the intensity of nourishment and wastage (amount of snow accumulation, melting, iceberg calving), and also the wastage process which operates (whether it terminates on land by melting, or terminates in sea cliffs).

Some surface features

Many of the features observed on glacier surfaces have been studied and their modes of formation explained, so that appearance of these features gives useful information on glacier behavior or on local meteorological conditions.

Snowline and firn limit. The lowest level at which recent snow persists at any given time of year is called the snowline, or temporary snowline. The lowest level at which recent snow persists by the end of the melt season in any given year is the annual snowline for that year. The lowest level to which there is perennial snow, or firn, is called the firn limit, or firn edge. The firn limit separates the accumulation zone from the ablation zone.

Melt streams. When melting occurs on an impermeable ice surface, the water liberated trickles into streams which cut deep, smooth-walled channels. Flow velocities are generally high, (e.g., 10 ft/sec), and the discharge from moderate-sized streams may be of the order of 100 ft^3/sec. The surface drainage over a crevasse-free area develops a river system, with small tributaries feeding the main streams. Large melt streams constitute major obstacles to surface travel. Streams may flow directly off the

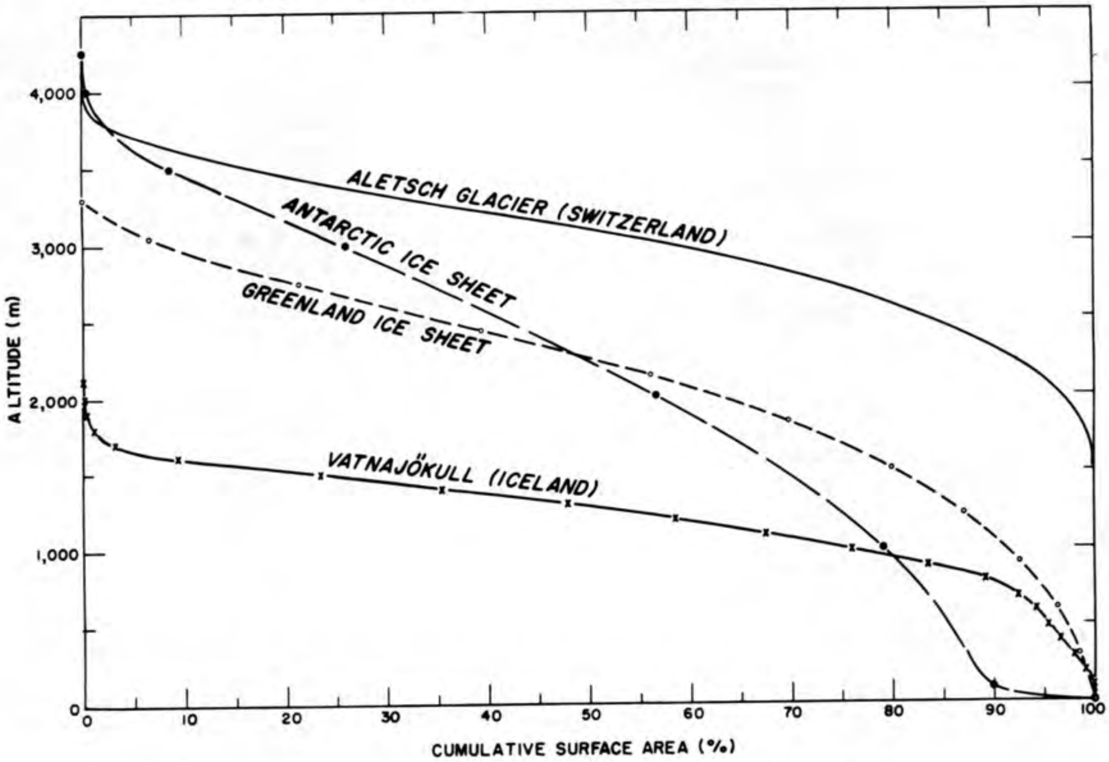


Figure II-2. Hypsographic curves for various glaciers.
(After Bauer, refs. 3, 4, 5)

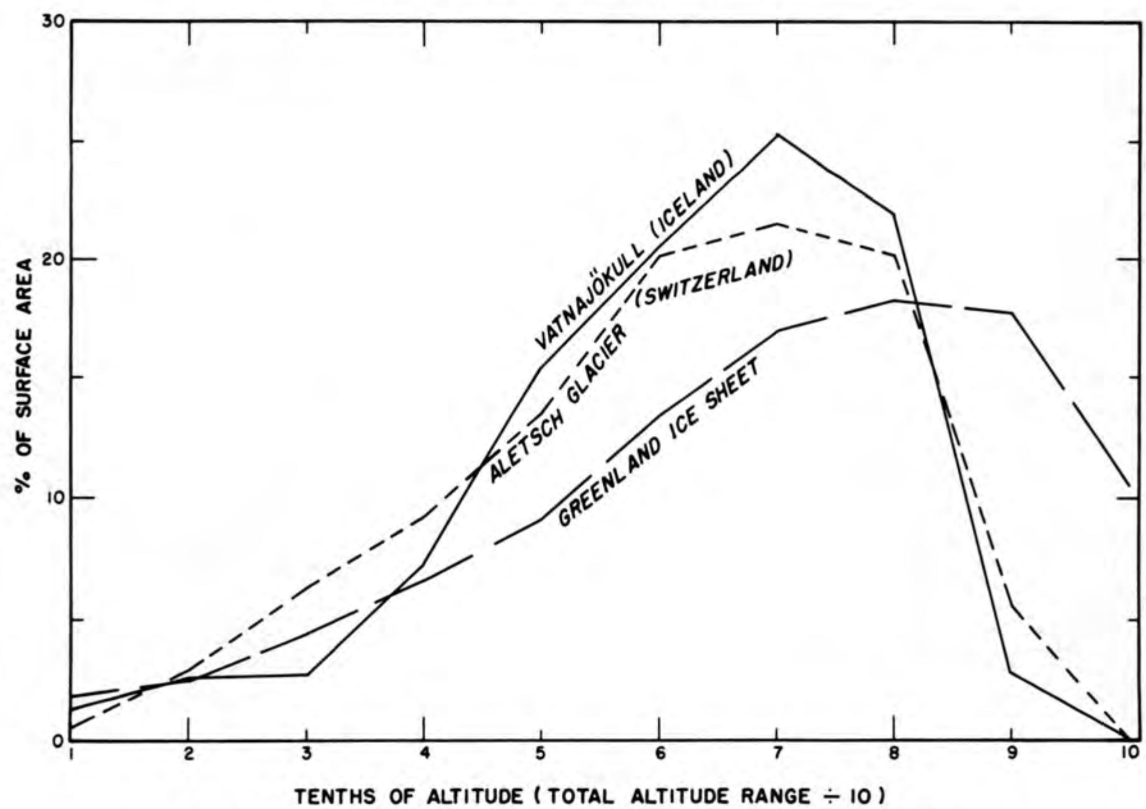
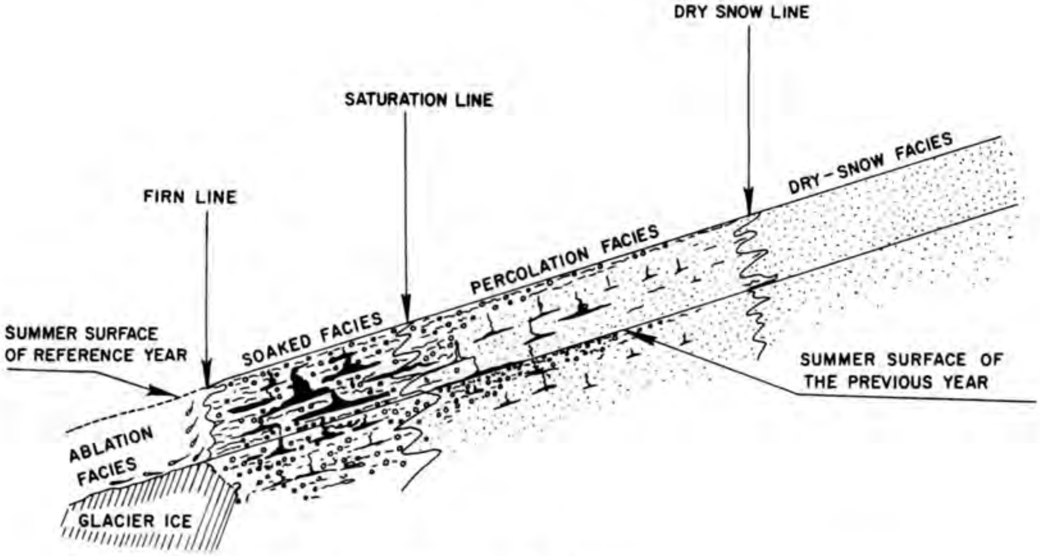
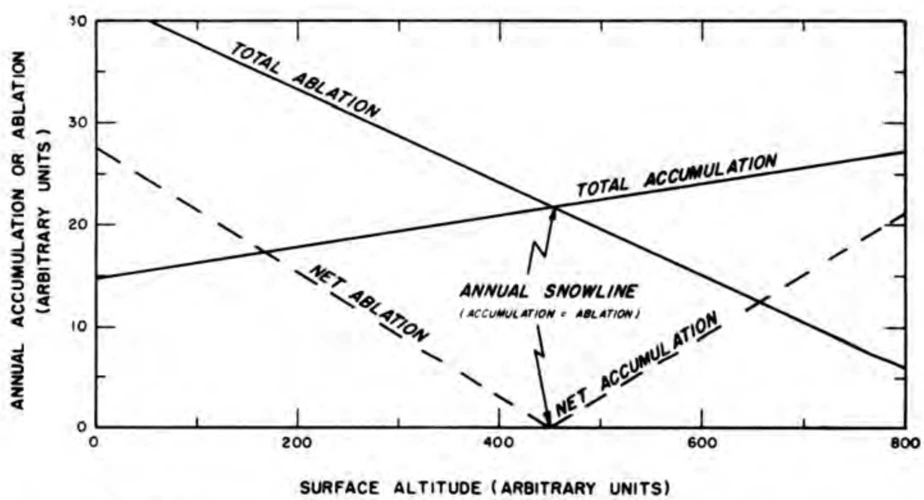


Figure II-3. Area distribution curves for various glaciers.
(After Bauer, ref. 4)



(a) Schematic illustration of the characteristic zones of accumulation and ablation. (After Benson, ref. 7)



(b) Schematic diagram illustrating what is meant by total and net accumulation and ablation. The net accumulation or ablation at any point is the difference between total accumulation and total ablation.

Figure II-4. Schematic illustration of accumulation and ablation.

glacier surface at the margins, or may disappear into sink-holes known as "moulins." These sink holes are usually carved out where a stream dives into a crevasse, and they allow surface water to find its way to the bed of the glacier, where subglacial streams are formed. Subglacial or englacial streams can often be seen issuing from the snout of a land-terminating glacier, and discolored currents near the terminal cliffs of a sea-terminating glacier indicate their presence.

Melt lakes. Melt water sometimes ponds in depressions on the ice surface to form sizable lakes (they can be big enough for landing a float plane). A melt lake may persist over several years, its surface freezing in the winter and being thawed or flooded in the summer. It can also be drained suddenly by opening of a crevasse in its bed. If the water of a deep lake drains out while the surface is frozen, an ice-roofed chamber is left; when such a roof collapses a subsidence feature (an "ice doline") is left.

Snow swamps. When melt water ponds in depressions where permeable snow overlies the impermeable ice layer, a water-logged mass of slush results. Snow swamps are commonly several feet deep. A snow swamp is an obstacle to surface travel, and in certain lighting conditions it may be undetectable from the air (or even from the ground).

Glacier tables. A glacier table is a column of ice topped by a large boulder. It forms because the boulder protects the ice beneath it from ablation, while the surrounding ice is melted down. The same effect is obtained when a tent or hut is set up on a rapidly ablating surface.

Dirt cones (sand cones, gravel cones). Dirt cones appear to be steep conical piles of dirt standing on the glacier surface; they are actually cones of ice with a covering of sandy or gravelly material. They form by the same process as glacier tables, i.e., the debris (which must be a fairly thick pocket) protects the ice beneath it from ablation while the surrounding ice is lowered by melting. The conical shape results from sliding of the dirt as the center becomes elevated. They are commonly up to 10 ft high, but giant cones 100 ft high occur on some glaciers.

Cryoconite holes (dust wells). While a thick cover of soil or rock protects the ice beneath from melting, a light scattering of dust or gravel accelerates melting by its capacity to absorb solar radiation. A pocket of dust, pebbles, or organic matter absorbs radiation and melts down a water-filled hole in the ice. The dirt stops sinking when it reaches a depth where it is effectively shielded from radiation. Hole diameters are commonly from a few inches to a foot, and maximum depths are around 2 ft. When the melt water in a cryoconite hole refreezes it forms quite clear ice (in contrast to bubbly glacier ice), with a central vertical column of bubbles and radiating bubble chains.

Penitents (Nieves penitentes). Penitents are ablation features found on cold dry snowfields subject to intense sunshine. They consist of slender conical pillars of snow in close array, and their heights range from a few inches (micro-penitents) to 10 ft or more. They incline so that they point in the direction of maximum radiation, and develop from differential ablation. A scattering of dust helps to initiate penitent formation. The name stems from a fancied resemblance to hooded penitent monks.

Hummocks. Hummocks are ablation features which sometimes occur on glacier ice. They can be steep and closely spaced, constituting an obstacle to vehicle travel. Though formed chiefly by differential absorption of radiation, melt streams may aid the surface dissection.

Crumbly ice. The cohesion between crystals of surface glacier ice may be lost (usually in cold sunny weather) as a result of preferential melting at the crystal boundaries, where impurities are concentrated. The loose crystals are frequently columnar. indicating recrystallization before melting.

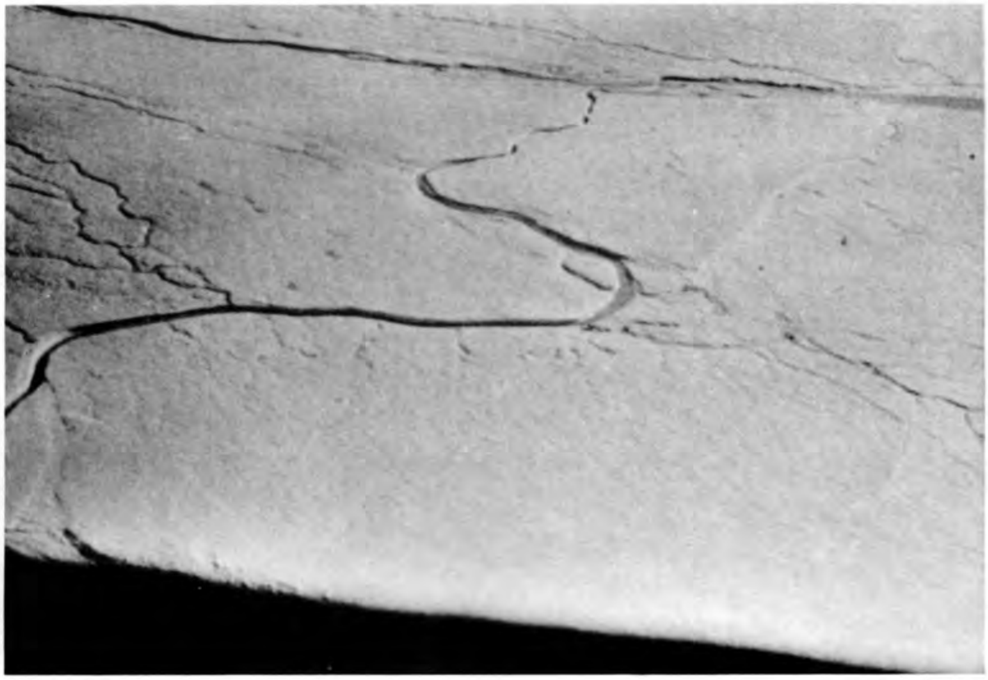


Figure II-5a. Melt streams on the ice cap surface in south-west Greenland. (CRREL photo)

Figure II-5b. Hummocks on the surface of a Spitsbergen glacier. Lines of hummocks follow foliation planes, and some dissection is apparently due to melt water flow.

Figure II-6a. Snow swamp in south-west Greenland. (CRREL photo)

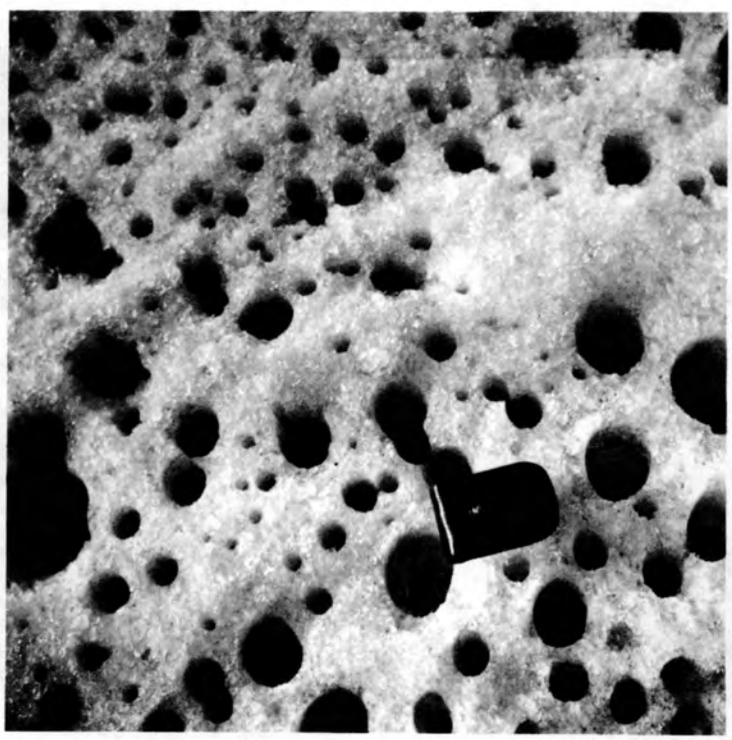


Figure II-6b. Cryoconite holes. (Photo by V. Schytt)

Figure II-7. Basal moraine in Antarctic glacier ice.

Moraines.* Moraine is glacier-transported debris. Surface moraine usually forms in distinct longitudinal strips along the glacier edges, where loose rock falls and collects (lateral moraines). When two separate valley glaciers unite, their adjacent lateral moraines become a medial moraine. Moraine is also collected beneath the surface wherever ice shears over the rock (walls or bed), and when separate glaciers unite in a compound system this material can become lodged inside the ice mass of the main stream. Moraine forms characteristic deposits of unsorted angular material after it is dumped by the ice.

"Flow lines" (longitudinal septa, longitudinal foliation). The surface of ice which is flowing under lateral confinement often displays longitudinal stripes which follow the flow direction. On investigation the ice of these stripes proves to be structurally different from the ice of the main mass, apparently due to shear deformation on more or less vertical planes.

Crevasses. Crevasses are surface cracks (commonly as much as 100 ft deep) which open when the ice is subjected to tensile or shear strains. They are indicative of a flow condition in which the ice is tending to stretch, and they form at right angles to the directions of tensile strains (see Chapter V).

Arc-shaped banding (arcuate foliation). In the ablation area of a valley glacier, transverse, arc-shaped bands are often seen on the ice surface. These are flow structures made visible by differences of bubble content and crystal orientation. They give a visual impression of transverse distribution of velocity, i.e., flow in the center is faster than at the edges.

*Moraine studies have yielded a wealth of information on glacial geology. For a detailed discussion of moraines, see Flint, ref. 13.

Ogives. Ogives are a special kind of transverse arc-shaped band, concave on the upstream side, appearing as alternating light and dark bands, with or without associated surface undulations. They form at the bases of certain ice falls by some periodic process (several mechanisms have been suggested[21]).

Sastrugi. Sastrugi are sharp-edged, wind-sculptured surface irregularities which form by erosion and re-deposition where the snow is cold and dry. They are aligned with their long axes in the direction of the wind which formed them; on ice caps this is usually the prevailing katabatic wind, which tends to flow down the line of steepest surface slope.

Snow dunes. Snow dunes form when snow falls in strong wind conditions. They have a rounded whaleback shape and are much larger than sastrugi. Their long axes are aligned in the direction of the snow-bearing wind which laid them down; this is generally a cyclonic wind.

Ripples, barchans, terracing, pocking. A variety of surface irregularities may be formed by wind on a dry snow surface. Ripples, which are transverse to the wind direction, are laid down in gentle winds. This is true to a lesser extent of the crescentic snow barchans. The rugosities described by such terms as "terracing" and "pocking" usually result from deflation of a crusted snow surface.

Area and thickness of existing glaciers

Area. The areas of present-day individual glaciers range from less than 1 km^2 for small snow beds to almost 13,000,000 km^2 for the Antarctic Ice Sheet. The range therefore covers about seven orders of magnitude.

Snow beds, cirque glaciers, and similar small ice bodies generally have areas between 1 and 10 km^2. Minor valley glaciers, such as those of the Alps, are approximately 10 to 100 km^2, while major valley glaciers and most ice caps have areas from about 100 km^2 to more than 1,000 km^2. The biggest Arctic ice caps, excluding Greenland, are of the order of 10,000 km^2 in area. From snow beds to big ice caps there is a continuous grading of sizes, but the major ice sheets are in a class apart, with the Greenland one (1,700,000 km^2) roughly two orders of magnitude larger than its nearest rival. At the top of the range the Antarctic Ice Sheet is nearly an order of magnitude bigger than the Greenland Ice Sheet.

Thickness. The thickness of glacier ice runs from almost zero, for marginal ice, up to about 4300 m for the deepest parts of the Antarctic Ice Sheet. If the smallest glaciers are excluded, average thicknesses for complete glacier systems range through about one order of magnitude, from 200 m or so for minor valley glaciers to 2300 m for the Antarctic Ice Sheet. Typical thicknesses for valley glaciers are from 200 to 600 m, and sizable ice caps might perhaps be 700 or 800 m deep. The Greenland Ice Sheet has an average thickness of about 1500 m and the Antarctic Ice Sheet about 2300 m.

Figures 8-13 give a number of ice depth profiles for various kinds of glaciers.

World distribution of glaciers

Table II-1 gives some statistics on the geographical distribution of glaciers. It can be seen from this table that most of the world's glacier ice is in the Antarctic, the ice sheet and its surrounding islands accounting for 85% of the world total glacier area and 91% of the world total glacier volume.

Most of the remaining ice is in Greenland, which has 12% of the total glacier area and 8% of the total glacier volume. Other glacier-covered parts of the earth combine to provide only 3% of the total area, and 1% of the total volume, of glacier ice in the world.

Outside of Greenland and Antarctica, the Arctic islands and Iceland carry more than half of the remaining glacier ice. Continental Asia has extensive glaciers, and continental North America has quite an abundance. The glaciers of continental Europe, which are such famous attractions, make an almost insignificant contribution.

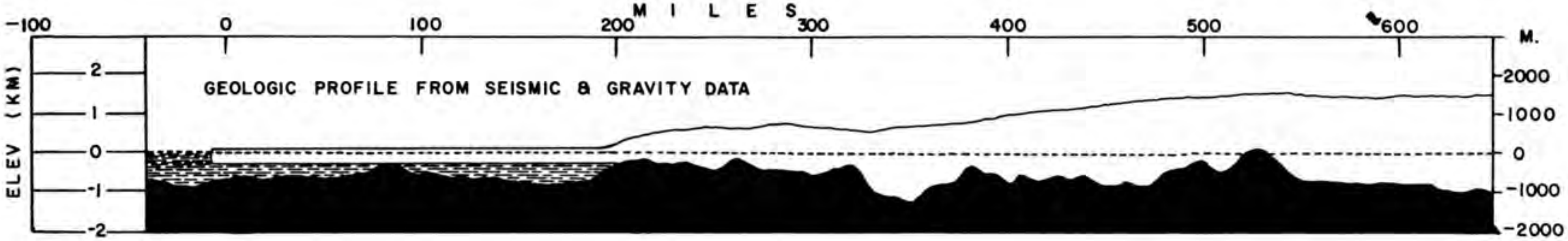


Figure II-8. Ice depth profile in Antarctica — Little America to Byrd Station. (After Woollard, ref. 24)

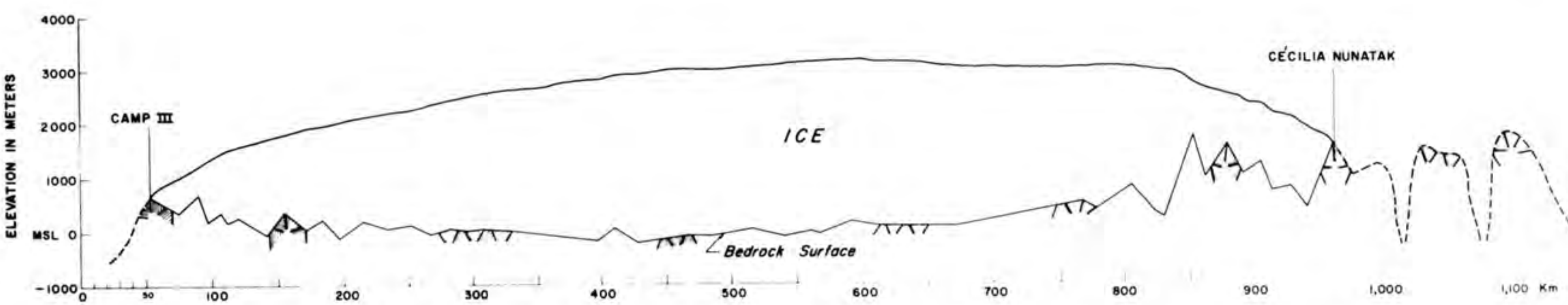


Figure II-9. Ice depth profile across Greenland. (After Holtzscherer, ref. 14)

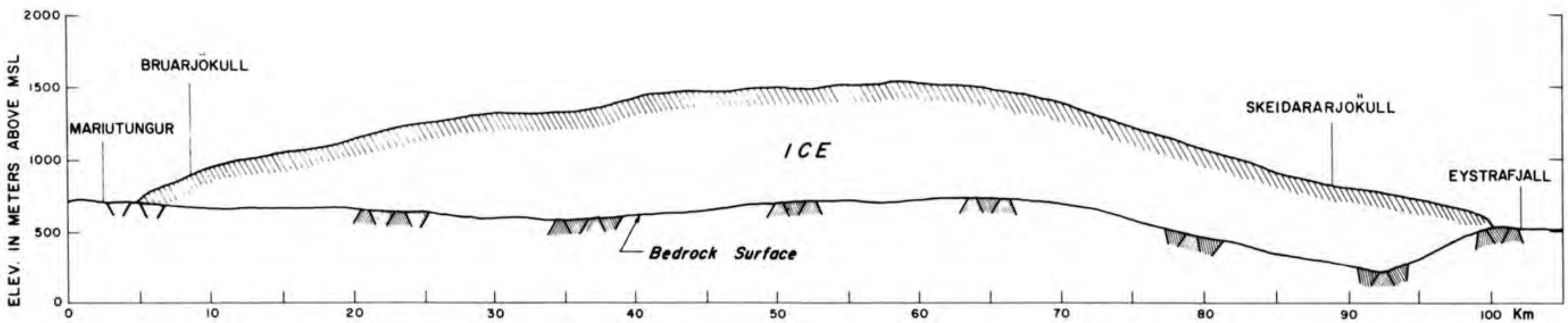


Figure II-10. Ice depth profile across Vatnajökull, Iceland. Vertical exaggeration X10. (After Joset and Holtzscherer, ref. 15)

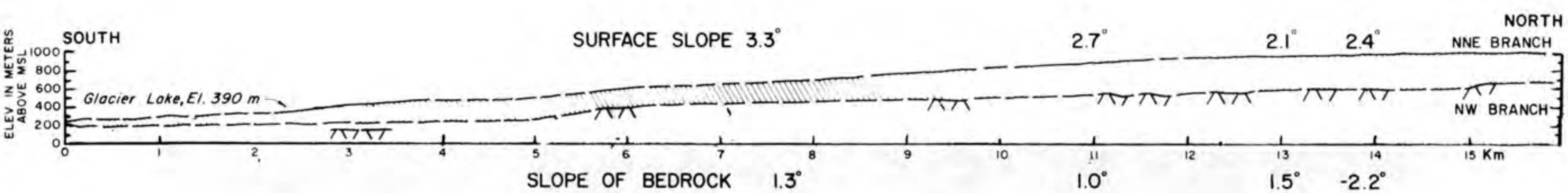


Figure II-11. Ice depth profile along the Highway Glacier, Baffin Island. (After Roethlisberger, ref. 19)

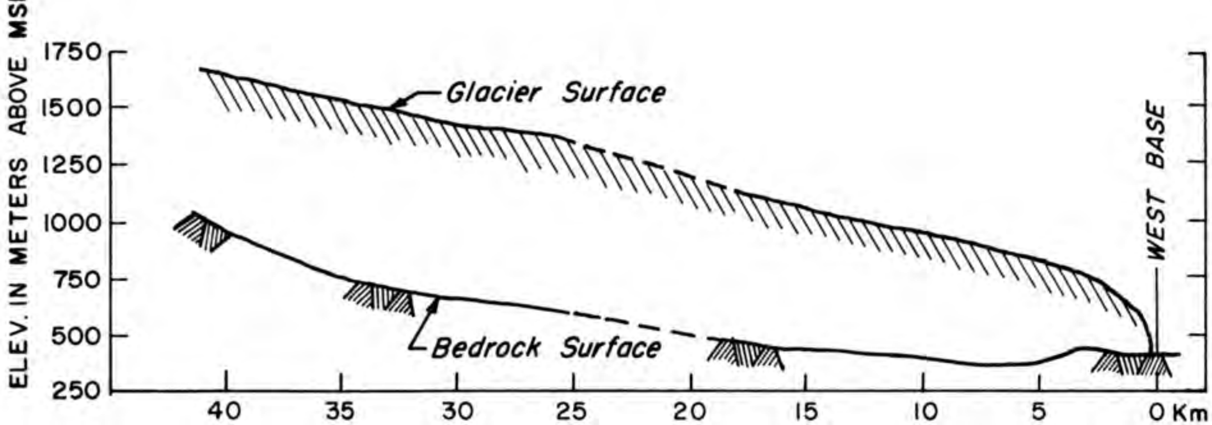


Figure II-12. Ice depth profile along Gilman Glacier, Ellesmere Island. (After Weber <u>et al.</u>, ref. 26)

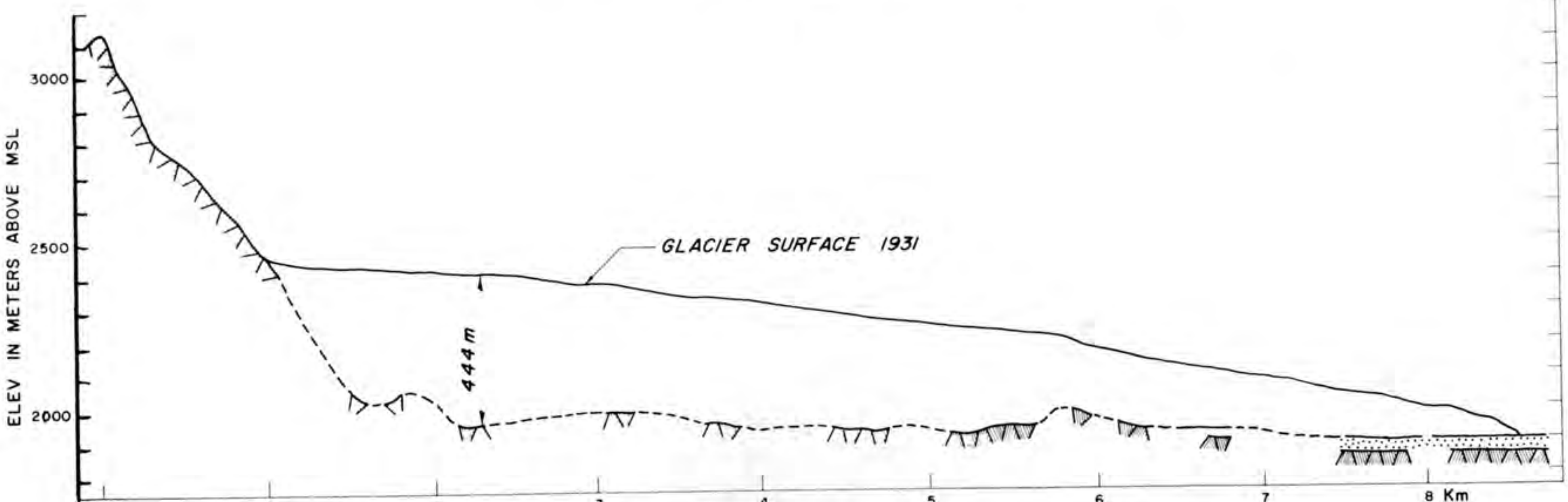


Figure II-13. Ice depth profile along the Unteraar Glacier, Switzerland. Vertical exaggeration X2. (After Mercanton and Renaud, ref. 17)

Table II-I. World distribution of glaciers.

(Based on estimates and assessments by Bauer[3,6] Crary[9] Donn *et al.*[10] Flint[13] Novikov[18] Thorarinsson[22] and others)

Region	Glacier area (Millions of sq km)	Percentage of world total glacier area	Estimated ice volume (Millions of cu km)	Percentage of world total ice volume
Antarctica & Sub-Antarctic Islands	13	85%	29	91%
Greenland	1.8	12%	2.6	8%
Arctic Islands & Iceland	0.3			
Continental Asia	0.1			
Continental North America	0.08	3%	0.2	1%
South America	0.03			
Continental Europe	0.01			
New Zealand, Africa, New Guinea	0.001			

The area of the earth's surface covered by present-day glaciers represents about 10% of the total land surface.

Glacier ice is a significant item in the worlds's water budget, representing an amount of water roughly equal to that contained in all the lakes, rivers and aquifers (Table II-II). Glacier ice and fresh waters each contribute about 2% of the world total water volume, the remaining 96% being sea water (atmospheric water is negligible).

Table II-II. Comparison of water volumes in oceans, glaciers, rivers and lakes.

Item	Percentage of world's total water volume
Oceans and seas[16]	96%
Glaciers (equivalent water volume)	2%
Rivers, lakes and ground water[16]	2%
Atmosphere	negligible

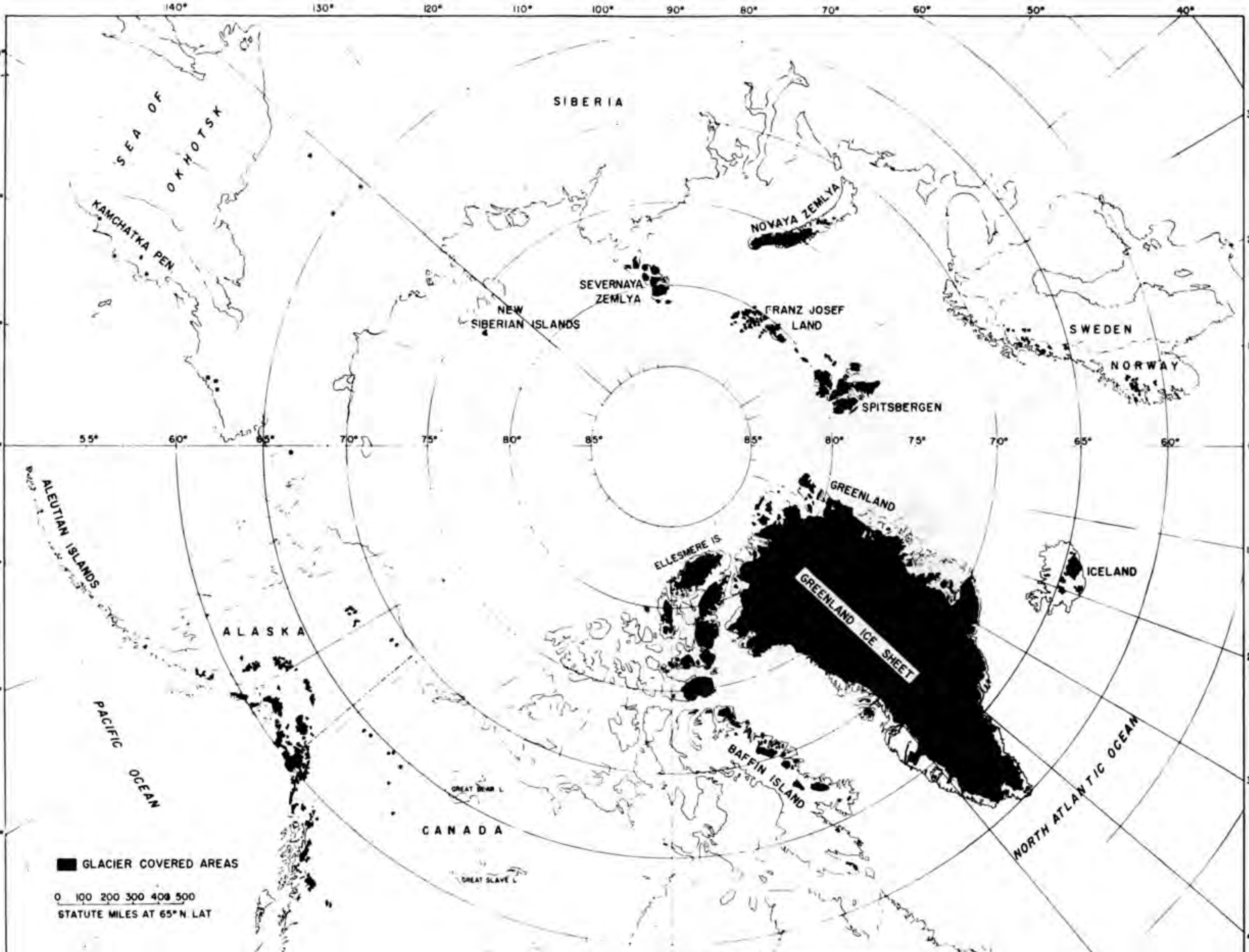


Figure II-14. Distribution of glaciers in the Arctic regions.

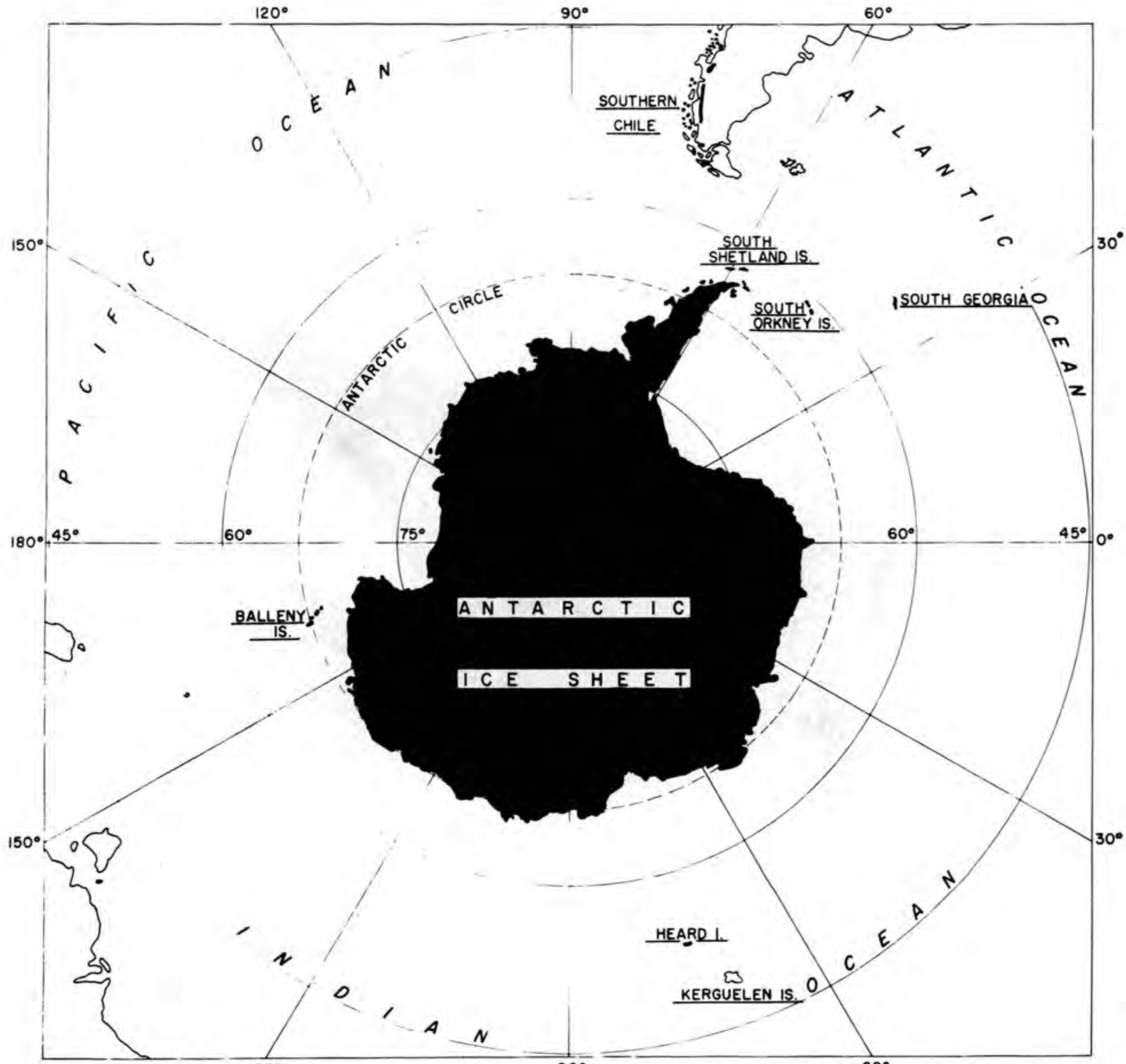


Figure II-15. Glaciers in the Antarctic regions.

BIBLIOGRAPHY

1. Ahlmann, H. W. (1948) Glaciological research on the North Atlantic coasts, Royal Geographical Society (London), Research Series, no. 1.

2. Armstrong, T. and Roberts, B. (1956; 1958) Illustrated ice glossary, Polar Record, vol. 8, no. 52; vol. 9, no. 59.

3. Bauer, A. (1961) "Nouvelle estimation du volume de la glace de l'inlandsis Antarctica, (New estimate of the ice volume of the Antarctic ice sheet)" in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55.

4. ________ (1955) Contribution à la connaissance du Vatnajökull, Islande (Contribution to knowledge of the Vatnajökull, Iceland, Jökull, no. 5 (Iceland Glaciological Society).

5. ________ (1955) The balance of the Greenland Ice Sheet, Journal of Glaciology, vol. 2, no. 17.

6. ________ (1955) Über die in der heutigen Vergletscherung der Erde als Eis gebundene Wassermasse (The present glacier ice cover of the earth), Eiszeitalter u. Gegenwart, vol. 6, p. 60-70.

7. Benson, C. S. (1960) Stratigraphic studies of the snow and firn of the Greenland Ice Sheet, USA CRREL Research Report 70.

8. Charlesworth, J. K. (1957) The Quaternary Era. London: Arnold, vol. 1.

9. Crary, A. P. (1960) Status of U. S. scientific programs in the Antarctic, IGY Bulletin no. 39; Transactions of the American Geophysical Union, vol. 41.

10. Donn, W. L.; Farrand, W. R. and Ewing, M. (1962) Pleistocene ice volumes and sea level lowering, Journal of Geology, vol. 70, no. 2.

11. Field, W. O. (1956) "Some aspects of glaciers and glaciology," in The dynamic north, U. S. Navy.

12. ________, et al. (1958) Geographic study of mountain glaciation of the northern hemisphere. New York: American Geographical Society.

13. Flint, R. F. (1957) Glacial and Pleistocene geology. New York: Wiley.

14. Holtzscherer, J. J. and Bauer, A. (1954) Contribution à la connaissance de l'inlandsis du Groenland, (Contribution to the knowledge of the Greenland ice sheet) Expéditions Polaires Francaises, Paris.

15. Joset, A. and Holtzscherer, J. J. (1954) Expédition Franco – Islandaise au Vatnajökull, Résultats des sondages seismiques (Franco – Icelandic expedition to the Vatnajökull, results of seismic soundings), Jökull, no. 4.

16. Kuenen, P. H. (1955) Realms of water. New York: Wiley.

17. Mercanton, P. L. and Renaud, A (1948) Les sondages seismiques de la Commission Helvétique des glaciers, (Seismic soundings by the Swiss Committee on Glaciers), Extrait des publications du Bureau Central Seismologique International, Série A., Travaux Scientifiques, Fasc. 17.

18. Novikov, V. (1960) The study of the Antarctic is continuing, Priroda, vol. 8, p. 43-52.

19. Roethlisberger, H. (1955) Studies in glacier physics on the Penny Ice Cap, Baffin Island, Part III, Journal of Glaciology, vol. 2, no. 18.

20. Sharp, R. P. (1956) "Glaciers in the Arctic," in The dynamic north. U. S. Navy.

21. ________ (1960) Glaciers. University of Oregon Press.

BIBLIOGRAPHY (Cont'd)

22. Thorarinsson, S. (1940) Present glacier shrinkage and eustatic changes of sea level, Geografiska Annaler, vol. 22, p. 131-159.

23. Wahrhaftig, C. and Cox, A. (1959) Rock glaciers in the Alaska Range, Bulletin, Geological Society of America, vol. 70, no. 4.

24. Wright, C. S. and Priestley, R. E. (1922) Glaciology, British (Terra Nova) Antarctic Expedition 1910-13, London.

25. Woollard, G. P. (1958) Preliminary report on the thickness of ice in Antarctica, IGY Bulletin no. 13, National Academy of Sciences.

26. ______________ (1957) Discussion on the classification of glaciers, Journal of Glaciology, vol. 3, no. 21.

27. Weber, J. R., Sandstrom, N. and Arnold, K. C. (1961) "Geophysical surveys on Gilman Glacier, northern Ellesmere Island," in Snow and Ice Commission, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 54.

CHAPTER III. DEPOSITION AND ACCUMULATION OF SNOW

General

When precipitation falls in the form of rain, the rain falls directly onto the surface below it, but this is not necessarily true of falling snow, which can be carried horizontally for long distances by wind, both during and after the original snowfall.

If snow falls when the weather is calm, flakes settle gently to the ground and collect in a loose fluffy mass of low-density snow. Later, when the wind begins to blow, loose snow is picked up and carried along as a stream of particles concentrated near the ground, and as a higher "blizzard cloud" in which particles are borne aloft by turbulent eddies of the wind. The blown snow is subsequently re-deposited, where and when the wind loses its carrying capacity. The re-deposited snow is firm and dense compared to snow laid down in wind-free conditions, and its surface is characterized by drifts, dunes, and sastrugi resulting from turbulent deposition.

Snow which is actually precipitated from clouds during windy conditions cannot fall directly to the ground; it is carried along by the wind. When the concentration of snow suspended in the air becomes greater than the carrying capacity of the wind, some deposition occurs, although snow particles may still migrate along the surface.

After a windy period the surface snow stabilizes by "age-hardening", a sublimation process which develops bonds between adjacent snow grains. An age-hardened surface is not easily disturbed by subsequent winds blowing over it, although it is still subject to evaporation.

The initial properties of surface snow are determined by the weather conditions prevailing during its deposition. The initial density depends on the size of the snow crystals, and on the amount of wind-working the snow particles and the surface have received. The density of snow which has been lying for several months can also be correlated with mean air temperature during the period, since the rate at which the snow settles, or densifies under its own weight, is influenced by temperature.

On most glaciers snowfall is usually accompanied by wind, and conventional precipitation gages are quite useless for measuring local precipitation. The alternative is to measure the net accumulation* of snow on the surface, by observations on stakes or by inspection of the snow strata revealed in pit walls or drill cores. On flat open snowfields, the snow blown away from one locality may be replaced by an equal amount from upwind; in this case the net accumulation gives a good approximation to the precipitation. In general, though, snow is swept away from the area in which it fell and is replaced unequally with snow from upwind areas. Accumulation represents the following balance:

Net accumulation = precipitation + snow blown in + condensation (frost)

- snow blown away - evaporation.

For most glaciological studies, net accumulation has greater immediate significance than precipitation, but precipitation can be estimated from the net accumulation if measurements of the transport of blowing snow are made. The condensation and evaporation are often small balancing quantities, so that these terms can usually be neglected.

Mechanics of drifting snow

There are two principal snow transport mechanisms: Turbulent suspension (when particle settlement under gravity is balanced by upward eddies), and saltation (when particles bound along the surface impelled by the wind shear there)[9, 15] Turbulent suspension accounts for high blizzard clouds, which move large quantities of snow and

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* Net accumulation on a glacier is the annual surplus of snow in the accumulation area; total accumulation is the total quantity of snow falling onto the entire glacier in one year.

Figure III-1a. Smooth, soft snow laid down in wind-free conditions. (Photo by V. Schytt)

Figure III-1b. Patches of sharp-edged sastrugi and discontinuous "pavements" of pocked hard snow.

Figure III-1c. Oblique aerial view of longitudinal snow dunes on the Greenland Ice Cap. The dune orientations indicate at least two major wind directions.

Figure III-1d. Barchan formed on the snow surface in northern Greenland. (Photo by C. Benson) form big snowdrifts behind surface obstacles. Saltation creates sheets and streamers of densely concentrated particles which blow along within a few inches of the ground. When strong winds blow over loose snow, turbulent suspension and saltation are both active, but in light winds saltation can take place alone.

The rate of snow transport (mass flow per unit time) at any given height above the surface is doubly dependent upon wind velocity: the drift concentration (concentration of snow in air) is determined by wind speed, and the velocity of flow is directly related to wind speed. Snow transport depends not only on wind speed, however; it depends also on the available supply of loose snow. A moderate wind blowing over a dense, age-hardened snow surface will not generate much drifting, but if it blows over soft fluffy snow it will raise a high drift cloud. If new precipitation is falling there may be too much snow for the wind to carry, and deposition will occur. In this connection it is convenient to introduce a concept of "saturation": If a snow-bearing wind is under-saturated it is capable of picking up loose snow (eroding), but if it is over-saturated it will deposit snow.

The theory of steady turbulent suspension is based on a vertical equilibrium condition for a snow particle subject to upward turbulence and downward gravity force:[7,10]

$$w\,n + \left(\frac{A}{\rho}\right)\frac{\partial n}{\partial z} = 0 \tag{1}$$

where $\underline{w}$ = particle fall velocity, $\underline{n}$ = drift concentration (mass of snow per unit mass of air) at height $\underline{z}$, ρ = air density (assumed constant), $\underline{A}$ = coefficient of exchange at height $\underline{z}$.

Moderate and strong winds over snowfields appear to have a vertical distribution of velocity confirming to a logarithmic profile with small roughness parameters, so that

$$v = \frac{u_*}{k}\log_e\left(\frac{z}{z_0}\right) \tag{2}$$

and

$$A = k\rho u_*\, z \tag{3}$$

where $\underline{v}$ is velocity at height $\underline{z}$, u_* is the shear velocity (ρu_*^2 = the boundary layer shear stress), $\underline{k}$ is von Kármán's constant (0.4), and z_0 is the roughness length (in general, $z_0 << z$).

Thus eq 1 has the solution

$$\frac{n}{n_h} = \left(\frac{z}{h}\right)^{-\frac{w}{ku_*}} \tag{4}$$

where $\underline{h}$ is a fixed reference height.

Substituting for u_* (from eq 2) in the logarithmic form of eq 4 and using the roughness height z_0 (where the wind velocity is zero, according to eq 2) as reference level gives

$$\log\left(\frac{n}{n_{z_0}}\right) = -2.303\,\frac{w}{k^2}\left[\log\left(\frac{z}{z_0}\right)\right]^2 v^{-1} \tag{5}$$

showing that, at any given height, a straight line relationship between log n and $\frac{1}{v}$ is to be expected provided n_{z_0} does not change appreciably with wind conditions (which is an observed fact).

Field measurements of snow transport and wind profiles give results which appear to be in agreement with the theory.[7]

The rate of snow transport at a given level, $\underline{q}$, is rather a complicated function of wind speed, since

$$q = v\,n$$

Integration over a layer leads to an analytical expression for the snow transport which is a combination of power and exponential functions, and analysis of field observations over a limited range of wind velocities supports this result. It is interesting to note (W. Budd, personal communication) that as the wind becomes very strong the snow transport ends up being a linear function of wind velocity, since turbulence then has produced an almost even distribution of snow throughout the drift layer.

The foregoing* can be summarized as follows:

The amount of snow carried by a wind is determined by the abundance of available loose snow, and by the wind characteristics. Theory and experiment lead to the following conclusions on the effect of wind speed:[7]

(a) Drift density, $n\rho$ (g/m^3) varies with height above the surface according to a power law for a given wind strength.

(b) Drift density at a given height is an exponential function of the reciprocal of wind velocity at that height.

(c) The rate of snow transport in a given horizontal layer is an exponential function of the wind velocity at a fixed reference level (say 10 m, the standard anemometer height) for moderate and strong winds; when the wind becomes very strong (the standard of reference is provided by the particle fall velocity), the dependence of drift transport on wind velocity asymptotically approaches a linear relation.

Size of drift particles

The size of drift snow particles has received relatively little attention, although a number of writers mention "average" particle sizes around 0.1 mm.[4, 8, 10, 14] When a predominant particle size is given, it is not always stated whether that size refers to the most numerous particles, or to the particles which make the greatest contribution to the total mass. Figure III-2, which is adapted from a study of 4000 drift particles by Lister,[9] shows that particles in the size range 0.06-0.1 mm occur most frequently, but particles in the size range 0.1-0.2 mm make the biggest contribution to the total mass.

Lister's study was made at Southice, Antarctica, but the size distributions are probably typical for most of Antarctica and Greenland. When fresh precipitation is contributing to the blowing snow, some larger unfragmented snow crystals may well be present; typical sizes for various kinds of snow crystals are given in Chapter I.

Blowing snow and surface relief

A snow surface which has been modified by blowing snow exhibits a characteristic micro-relief. During heavy blizzards caused by the precipitation and wind of a cyclonic storm, snow is usually deposited in the form of elongated dunes, which have a rounded "whaleback" profile and their long axes parallel to the wind direction. The formation of discrete longitudinal dunes instead of a continuous deposit seems to indicate a transverse instability in deposits laid down by strong winds (cf. Bagnold's study of sand dunes[1]). Katabatic winds, which re-distribute surface snow laid down earlier, tend to form a more sharply sculptured surface. The sharp-edged sastrugi formed in katabatic winds are much smaller than whaleback dunes, but their long axes are also aligned along the wind direction. Light winds will produce transverse ripples, which usually are only stable at low wind speeds, changing to longitudinal features if wind velocity increases. Crescent-shaped barchans also form on the snow.

When flying over the snowfields of a windy region, one may notice two or more well-defined wind directions indicated by the surface micro-relief. Longitudinal whaleback dunes, commonly about 100 ft long and 10-30 ft wide, point the direction of the

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*** These findings may be modified in the light of new data.**

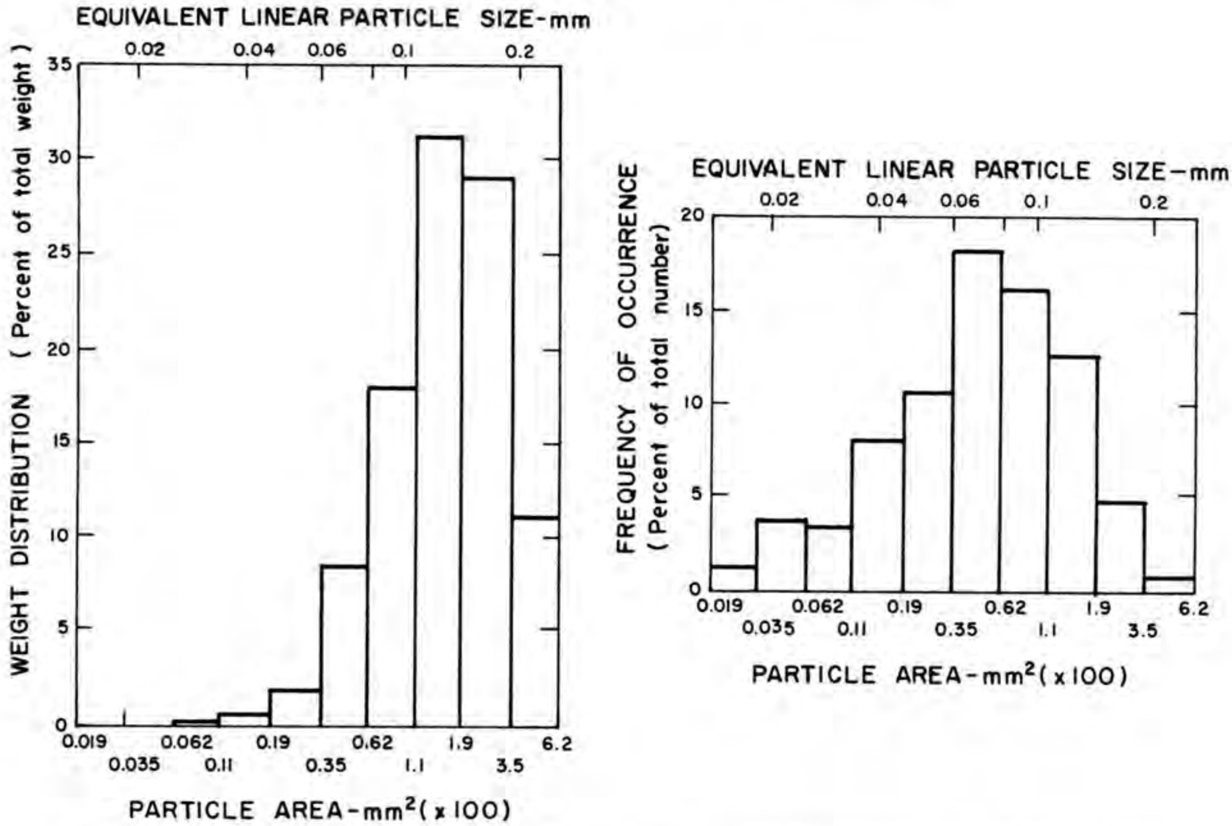


Figure III 2. Grading diagrams showing frequency of occurrence of the various particle sizes in Antarctic drift snow, by number and weight fractions. (Adapted from Lister, ref. 9)

main snow-bearing (cyclonic) winds, while the smaller sastrugi (a few feet long and a foot or so wide) give local directions of katabatic winds. Katabatic winds are gravity winds resulting from temperature inversions over the snow surface, and they follow the steepest surface slopes (with some deviation due to the Coriolis force). The sastrugi they produce therefore indicate general direction of surface slope.

The heights of sastrugi can be correlated with the speeds of the winds which formed them, but the additional influence of initial surface roughness confuses the problem. In moderately windy areas sastrugi are commonly 6 in. to 1 ft high (trough to crest); in very windy places heights up to 2 ft are common. The giant sastrugi (up to 6 ft high) found in parts of Antarctica are usually due to erosion of, or deposition behind, whaleback dunes.

During the formation of sastrugi, while the snow particles are still unbonded, they migrate slowly along in the wind direction. After a period of age-hardening, however, they become resistant to disturbance by later winds (Fig. III-3).

Initial properties of surface snow

Most mechanical, thermal and electrical properties of dry snow can be related to density, temperature and grain structure of the snow. Since these three parameters are easy to measure, they provide a convenient means of describing snow and offer the possibility of predicting a range of physical properties. Physical properties are discussed in detail elsewhere,[2, 13] but the influence of meteorological factors on the character of surface snow is outlined here.

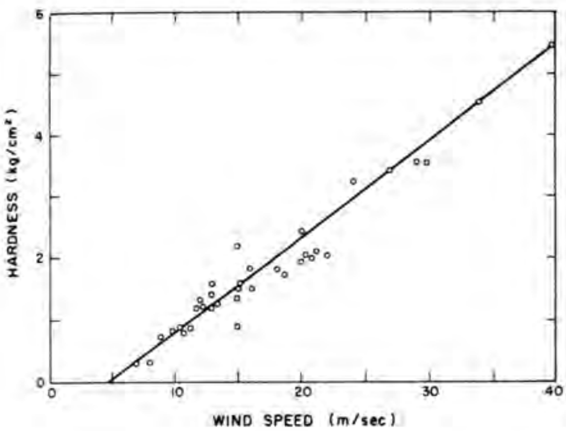


Figure III-3. Relationship between snow hardness and the wind speed necessary for erosion of that snow. (After Kotlyakov, ref. 8)

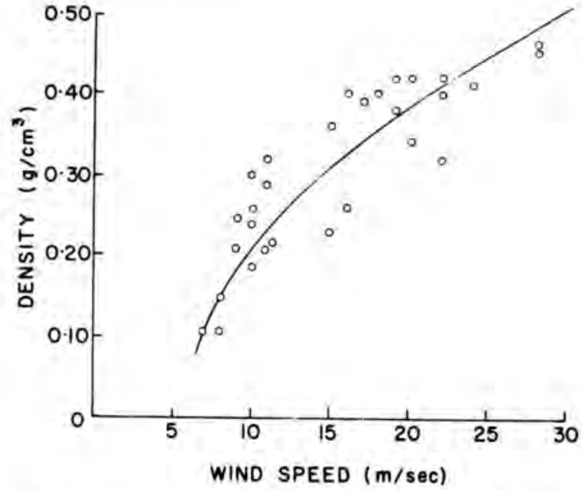


Figure III-4. Effect of wind speed on the density of newly deposited snow. (After Kotlyakov, ref 8)

If snow falls in cold, wind-free conditions, the initial density of snow on the surface is small — around 0.1 g/cm³. As long as there is no wind, density variations arise chiefly from differences in the size and shape of the snow crystals, small simple crystals (formed at low temperatures) packing more closely than larger stellar crystals (more common at higher temperatures).

Wind-blown snow, however, packs down to higher densities as a result of crystal fragmentation and agitation of the surface, and initial densities in excess of 0.4 g/cm³ can be realized. The effect of wind speed on initial density has not received much study, but Figure III-4 shows the general increase of density with wind speed found by Kotlyakov[8] in Antarctica. Some of the scatter in this plot can probably be attributed to temperature effect, which does not appear to have been isolated from the wind effect.

Surface snow gradually settles under its own weight, the rate of settlement being dependent on temperature. Since the temperature of snow near the surface corresponds closely to air temperature over long periods, a correlation between mean annual air temperature and average density of a one-year surface accumulation might be expected. Figure III-5 shows such a plot from data gathered in Antarctica; wind effects have not been removed.[14,16]

The bearing strength, or "hardness", of a snow surface depends upon grain structure as well as upon density and temperature. When snow is first laid down it is a cohesionless mass of grains, but ice bonds soon form between adjacent grains and "age-harden" the surface. The rate of sublimation (the process which produces bonds) is controlled largely by temperature and by the abundance of very fine particles in the snow, but substantial hardening usually occurs in the first two weeks after deposition, even at quite low temperatures.

Snow accumulation

Snow accumulation shows broad regional variations which are dependent upon general locations relative to tracks of low pressure systems, distance from the sea, surface elevations, and prevailing surface winds. It also varies on a local scale where abrupt changes of elevation occur, or where topographic irregularities influence drifting and deposition. On wind-swept snowfields there is variation on an even smaller scale due to the irregularity and discontinuity of deposition and erosion within small areas; this effect is shown up well by old vehicle tracks, which disappear intermittently beneath new dunes and sastrugi.

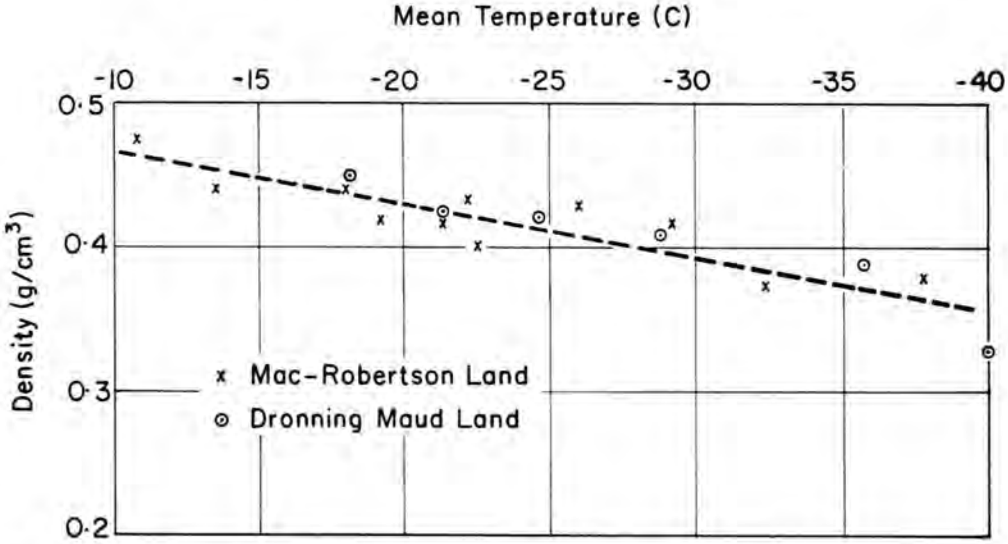


Figure III-5. Relation between mean annual temperature and average density of the surface snow layers. (After Schytt, ref. 16, and Mellor, ref. 14)

Local climatological conditions govern the type of snow accumulating in any given area. On high-altitude polar snowfields melting never occurs and the deposited snow stays dry; on slightly warmer snowfields, where some melting takes place in summer, water percolates from the surface into the lying snow to form dispersed pipes, layers and lenses of ice; at lower elevations, where appreciable melting occurs, water soaks the entire accumulation layer for the year to produce iced firn.[4] To allow comparison of accumulation quantities in areas with differing types of snow, accumulation is expressed in terms of its equivalent amount of water, i.e., the snow quantity (snow depth) is multiplied by the average specific gravity for the layer to give the water equivalent.

Seasonal weather changes alter the condition of falling and lying snow through the year, producing stratification in the snow mass and permitting the net accumulation of past years to be measured by examination of vertical sections revealed in pit walls or in drill cores. On some snowfields the annual snow layers are quite obvious; on mountain glaciers, for example, there may be distinct dust layers marking each summer surface. On the other hand, the cold dry snowfields of Greenland and Antarctica do not display a simple stratification, particularly in windy areas, and careful studies of grain structures and measurable physical properties are needed to interpret the profiles.

BIBLIOGRAPHY

1. Ahlmann, H. W. (1948) Glaciological research on the North Atlantic coasts, Royal Geographical Society, London. Research Series, no. 1.

2. Bader, H. (1962) "Snow as a material" in Cold Regions Science and Engineering (F. J. Sanger, Editor) U. S. Army Cold Regions Research and Engineering Laboratory monographs, Part II, Sect. B.

3. Bagnold, R. A. (1941) The physics of blown sand and desert dunes, London: Methuen and Co. Ltd.

4. Benson, C. S. (1962) Stratigraphic studies of the snow and firn of the Greenland Ice Sheet, USA CRREL Research Report 70.

5. Charlesworth, J. K. (1957) The Quaternary Era, vol. 1. London: Arnold.

6. Cornish, V. (1914) Waves of sand and snow and the eddies which make them. London: Unwin.

7. Dingle, W. R. J. and Radok, U. (1960) "Antarctic snow drift and mass transport," in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55.

8. Kotlyakov, V. M. (1961) Snow cover in the Antarctic and its role in modern glaciation of the continent. Moscow: Akad. Nauk SSSR (text in Russian).

9. Lister, H. (1960) Glaciology — Solid precipitation and drift snow, Trans-Antarctic Expedition Scientific Reports, no. 5. London.

10. Loewe, F. (1956) Études de glaciologie en Terre Adélie 1951-52 (Glaciological studies in Adélie Land 1951-52), Expéditions Polaires Francaises, Paris.

11. ________ (1957) "Precipitation and evaporation in the Antarctic" in Meteorology of the Antarctic. South Africa Weather Bureau, Pretoria.

12. Mather, K. B. (1962) Further observations on sastrugi, snow dunes and the pattern of surface winds in Antarctica, Polar Record, vol. 11, no. 71.

13. Mellor, M. (1962) "Polar snow — a summary of engineering properties," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

14. ________ (1959) A study of factors governing the mass balance of Antarctica, University of Melbourne, M. Sc. thesis.

15. ________ and Radok, U. (1959) "Some properties of drifting snow," in Antarctic meteorology. Pergamon Press.

16. Schytt, V. (1958) "Snow studies at Maudheim" and "Snow studies inland," in Glaciology II, Scientific results of the Norwegian-British-Swedish Antarctic Expedition, vol. IV. Oslo.

17. Sharp, R. P. (1956) "Glaciers in the Arctic," in The dynamic north, U. S. Navy.

18. Stephenson, P. J. and Lister, H. (1959) Preliminary results of the glaciological work on the Trans-Antarctic Expedition 1955-58, Journal of Glaciology, vol. III, no. 25.

19. Swithinbank, C. W. M. (1957; 1959) Glaciology I, parts B and E, Scientific results of the Norwegian-British-Swedish Antarctic Expedition, vol. III. Oslo.

20. Vickers, W. W. (1959) Wind transport of Antarctic snow, IGY Bulletin no. 22; Transactions of the American Geophysical Union, vol. 40.

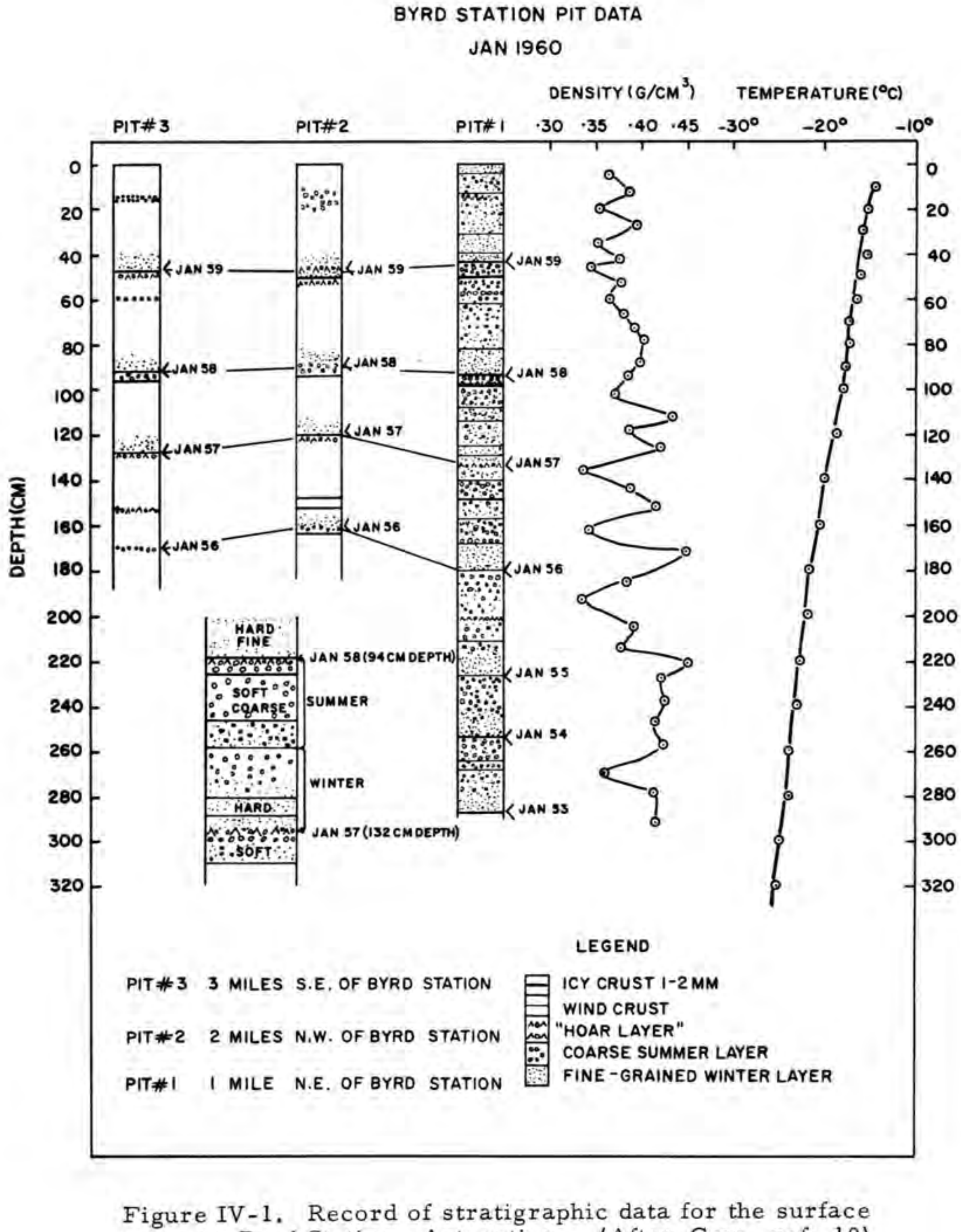


Figure IV-1. Record of stratigraphic data for the surface snow at Byrd Station, Antarctica. (After Gow, ref. 10)

CHAPTER IV. SNOW METAMORPHISM AND ICE FORMATION

Transformation of snow to ice

Soon after new snow is deposited on the surface of a glacier it loses its original crystalline form, either by a dry metamorphic process or by melting. If laid down in wind-free conditions the initial density may be low — about 0.1 g/cm^3 — but this density increases as a result of wind-packing, settlement under gravity, melt water percolation, or a combination of these things.

As time goes by, the snow is buried by later falls and it densifies under the weight of the snow above, with gradual grain growth proceeding. At first the densification of dry snow involves rearrangement of the snow grains in a way that permits closer packing; but after this process has reached its limit the grains themselves, and the bonds between them, have to change shape to achieve higher densities.[11] Eventually the air spaces between the individual grains of the cellular snow mass close off to form separate bubbles, changing the material from a permeable to an impermeable one.

Once the air spaces cease to be interconnecting and the material becomes impermeable it is known as ice instead of snow. Its density goes on increasing, however, as the snow above continues to build up. The densification, or compression, of the deeply buried glacier ice goes on largely at the expense of air bubble volume; the bubbles are squeezed ever smaller, and the pressure inside them consequently increases.

The stress condition deep within the ice mass is essentially hydrostatic, and the crystals appear to remain randomly oriented in this state.[10] Where the ice undergoes shear deformation, however, the crystals become re-oriented to permit glide along their basal planes.[18]

In areas where appreciable melting takes place, water soaks into the snow and turns it into dense impermeable ice while it is still close to the surface. This "wet" transformation to ice is much more direct and rapid than the "dry" process, and it tends to produce ice which is comparatively bubble-free.

Change of properties with depth

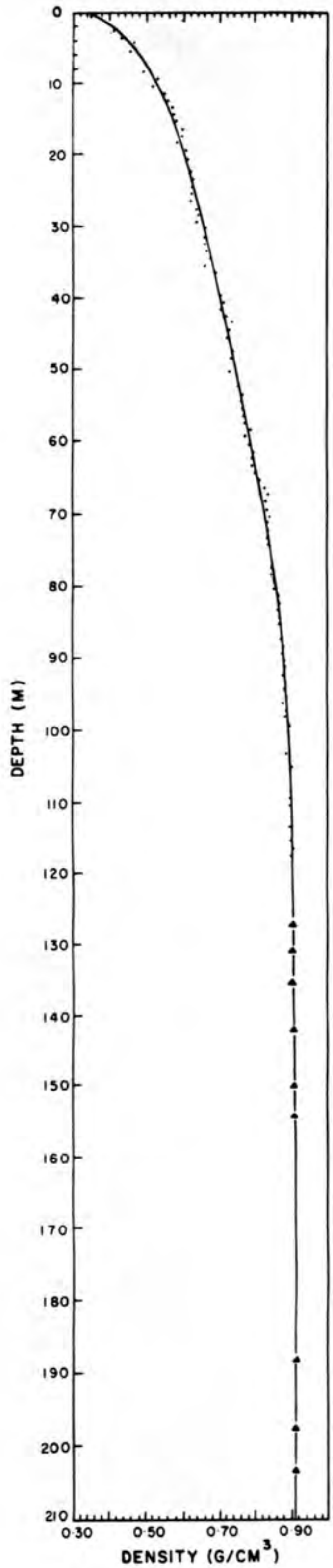
Two types of property variation beneath the surface can be distinguished: cyclic variations producing the seasonal stratification, and progressive variation with increasing depth. The first reflects recurring changes of deposition conditions, the second results from steady compression under ever-increasing overburden and from gradual changes of grain and crystal structure.

Stratification in dry polar snow is illustrated by Figure IV-1, which shows fluctuations of grain texture and density in snow close to the surface at Byrd Station, Antarctica. The character of each layer is determined by the weather conditions prevailing during its deposition and by the influence of ensuing weather conditions on the metamorphic processes. In areas where the snow surface is exposed to melt conditions in summer, the stratification is generally more obvious, and gross features can be photographed from the wall of a pit.

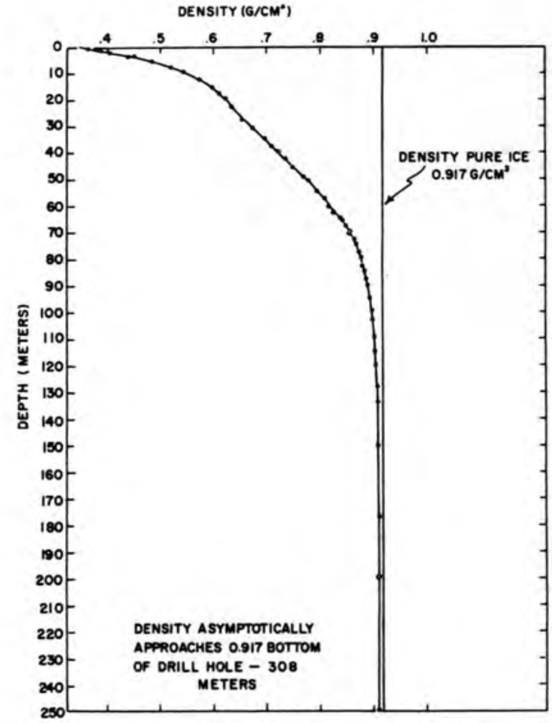
Density increase with depth is somewhat erratic at shallow depths, but overall the depth-density curve shows a regular form, with the rate of increase rapid in the upper layers but becoming more gradual as the density of solid ice is approached asymptotically at depth. Figure IV-2 gives some depth-density curves for sites in Greenland and Antarctica.

Grain structure and crystal form* change as a result of material transfer processes and mechanical deformation under pressure. In general, appreciable grain growth takes place near the surface (say in the first 2 meters), where temperature changes and convective air movements are felt most strongly. In the deeper snow no grain growth is apparent, but crystal size increases with depth.[20] In Figure IV-3 photomicrographs of

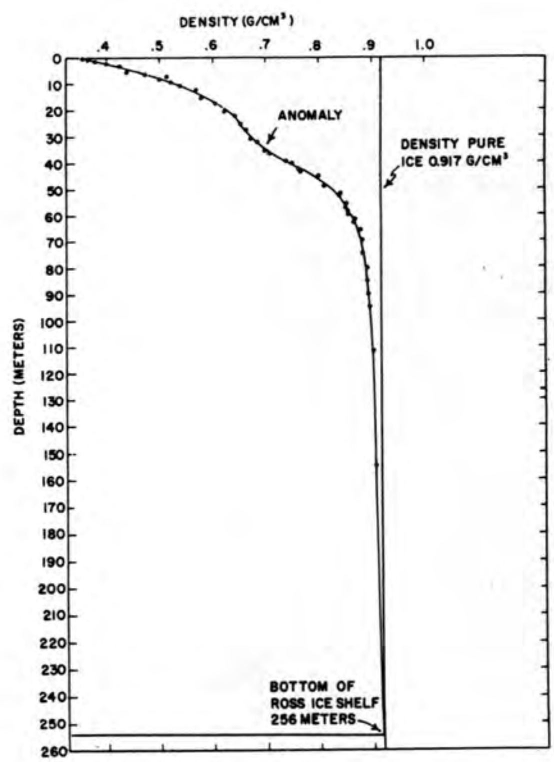
*A grain is a coherent particle which may consist of more than one crystal.



a. Site 2, Greenland.

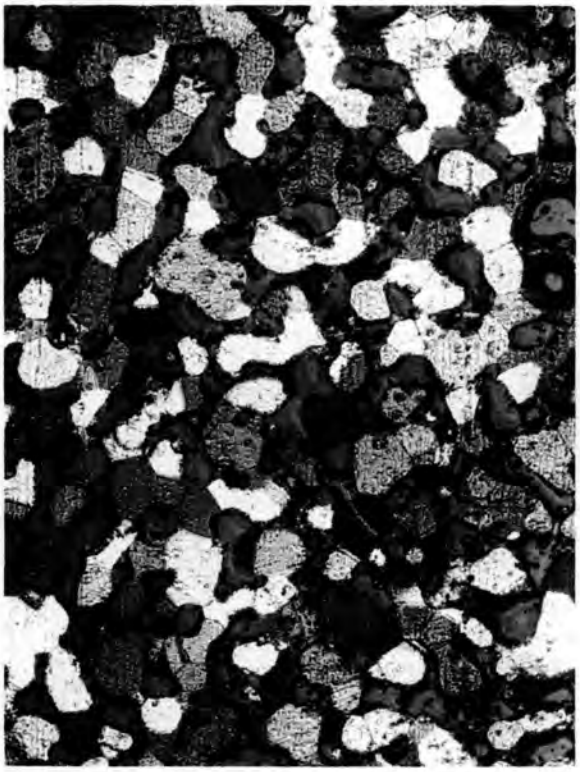


b. Byrd Station, Antarctica.

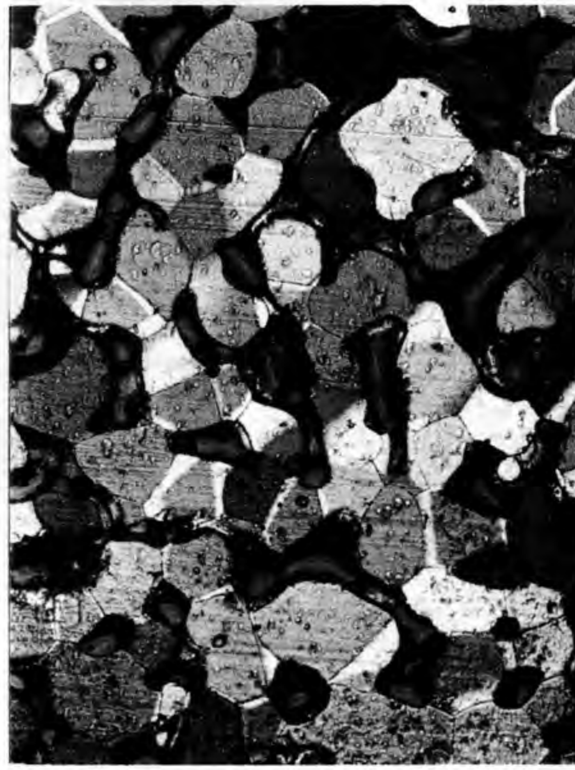


c. Little America, Ross Ice Shelf, Antarctica.

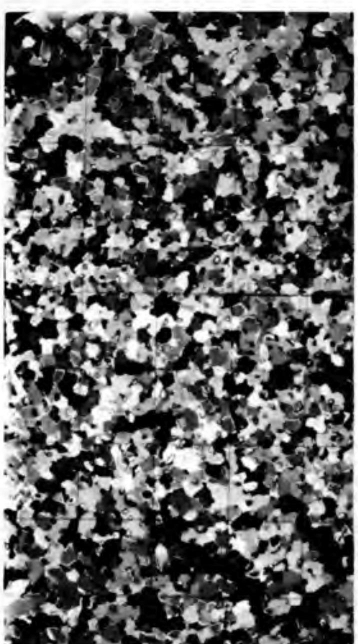
Figure IV-2. Depth-density curves.
(After Bender, Gow and Langway, CRREL data and ref. 26).



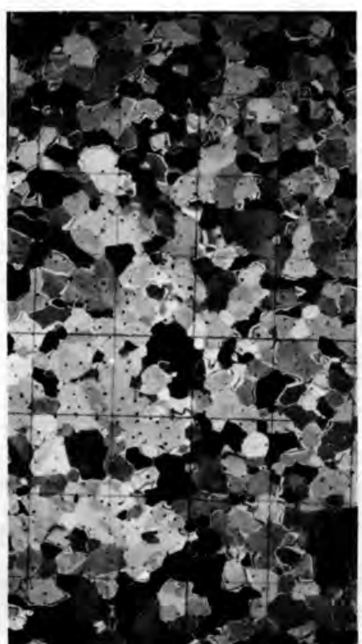
a. 10 meters (X4).



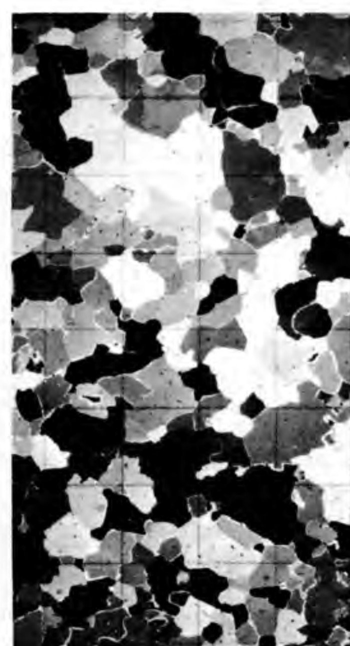
b. 35 meters (X4).



c. 65 meters (X1).



d. 193 meters (X1).



e. 306 meters (X1).

Figure IV-3. Photomicrographs showing the changes of snow and ice structure with depth at Byrd Station, Antarctica. Change of density with depth is shown in Figure IV-2. (Photomicrographs by A. J. Gow)

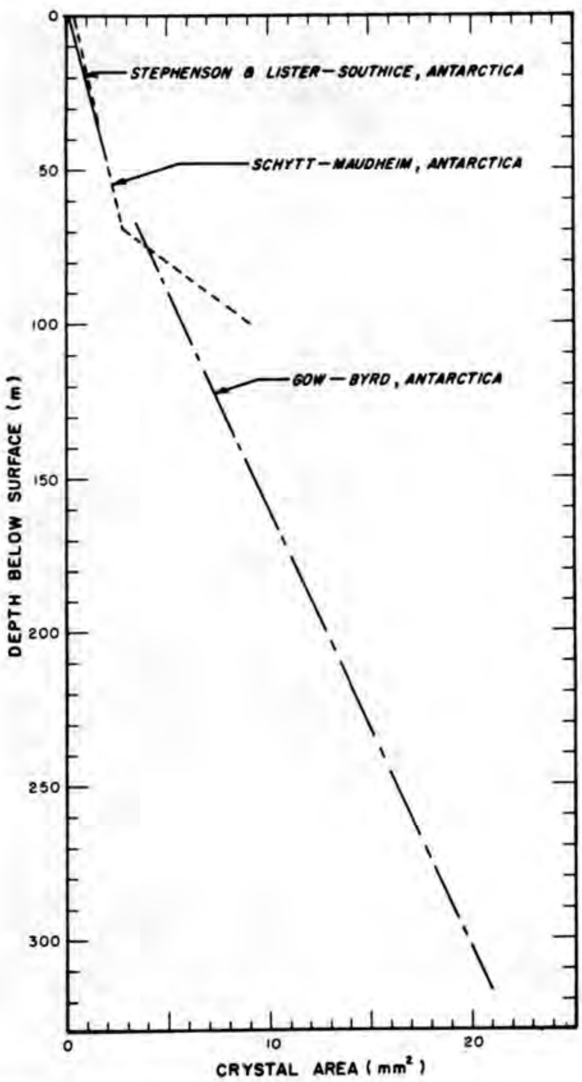


Figure IV-4. Increase of average crystal area with depth, from measurements made on thin sections. (After Gow, ref. 11; ref. 20; Stephenson & Lister, ref. 22)

snow and ice seen in thin section show changes of structure with depth. The gradual disappearance of fine grains and angular corners, and the development of intergranular bonds, can be seen (see also Fig. IV-5, 6); the reduction of pore space culminating in the formation of impermeable ice is also illustrated.

Permeability, which is related to porosity (or density) in snow decreases with depth, as shown in Figure IV-7, a plot of air permeability and porosity against depth. Air permeability is also affected by the grain structure of the snow, and in the upper layers the effect of decreasing porosity is offset to some extent by depletion of fine particles and the smoothing and rounding of larger grains as metamorphism proceeds.[23] The attainment of zero permeability marks the transition from snow to ice.

Mechanical properties change in accordance with the changes of density and grain structure. Elastic and visco-plastic properties (e.g. strength and deformation resistance) improve with depth, changes being rapid in the upper layers and tending towards a fixed limit after the snow-ice transition. Thermal and electrical properties follow the same trend.[16]

Types of metamorphism

Bader[5] distinguishes four types of metamorphic processes contributing to the transformation of snow to ice. They are (a) destructive metamorphism, (b) constructive metamorphism, (c) melt metamorphism, (d) pressure metamorphism. These are briefly described below.

Destructive metamorphism* is the disappearance of crystalline form in dry snow. Soon after snow is deposited in cold dry conditions, the original crystal shapes, e.g., star forms, are lost by sublimation and surface diffusion, and an aggregate of rounded or sub-angular particles results. The whole process can be regarded as a reduction of specific surface and corresponding decrease of surface free energy.

Constructive metamorphism* in dry snow is a process of grain growth depending chiefly on the transfer of material by evaporation and condensation. It is accelerated when vapor transfer is aided by convective air flow in the pores of the snow. Vigorous grain growth of this kind in snow of fairly low density ($< 0.4\ g/cm^3$) leads to the formation of a distinctive coarse-grained snow with low cohesion, known as "depth hoar" or "Schwimmschnee".

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*It is questionable whether these two processes should be separated, since both involve molecular migration.

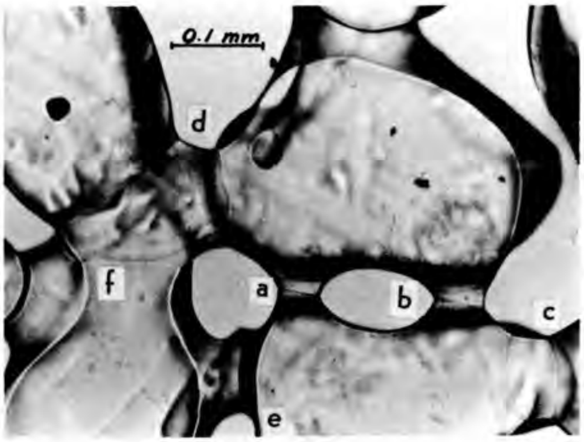

Figure IV-5. Thin section of snow showing the bonds formed between adjacent grains after 6 days of "age-hardening" at a temperature of -3C. (Photomicrograph by D. Kuroiwa, ref. 12)

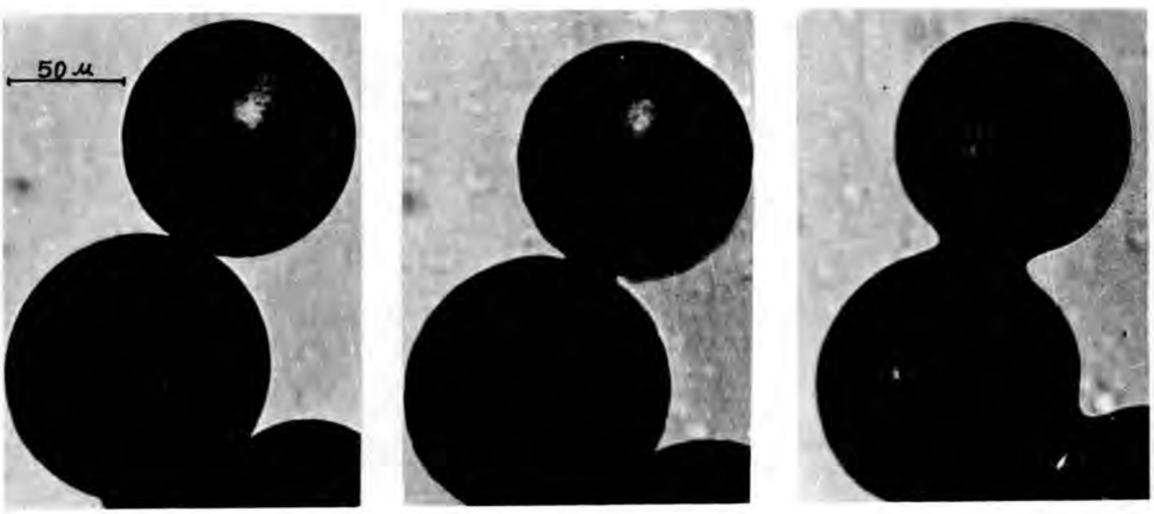

Figure IV-6. Development of a bond between two spheres of ice in ice-saturated air at a temperature of -5C. (a) shows the initial stage, (b) was taken after 33 minutes, and (c) was taken after 64 minutes. (Photomicrographs by D. Kuroiwa, ref. 12)

Melt metamorphism covers the changes resulting from rise of temperature to the melting point and percolation of surface melt water. In "warm" snow the crystals become rounded and are covered by a water film; surface tension effects tend to make water accumulate at intergrain contacts, leading to formation of strong bonds and composite grains on re-freezing. Melt-water percolation is an irregular process, and the water tends to re-freeze in pipes, layers and lenses, which are generally discontinuous. Melt-water retention in "warm" snow varies with the capillarity of the snow, surface slopes, and the presence of impermeable layers. Melt metamorphism produces rapid

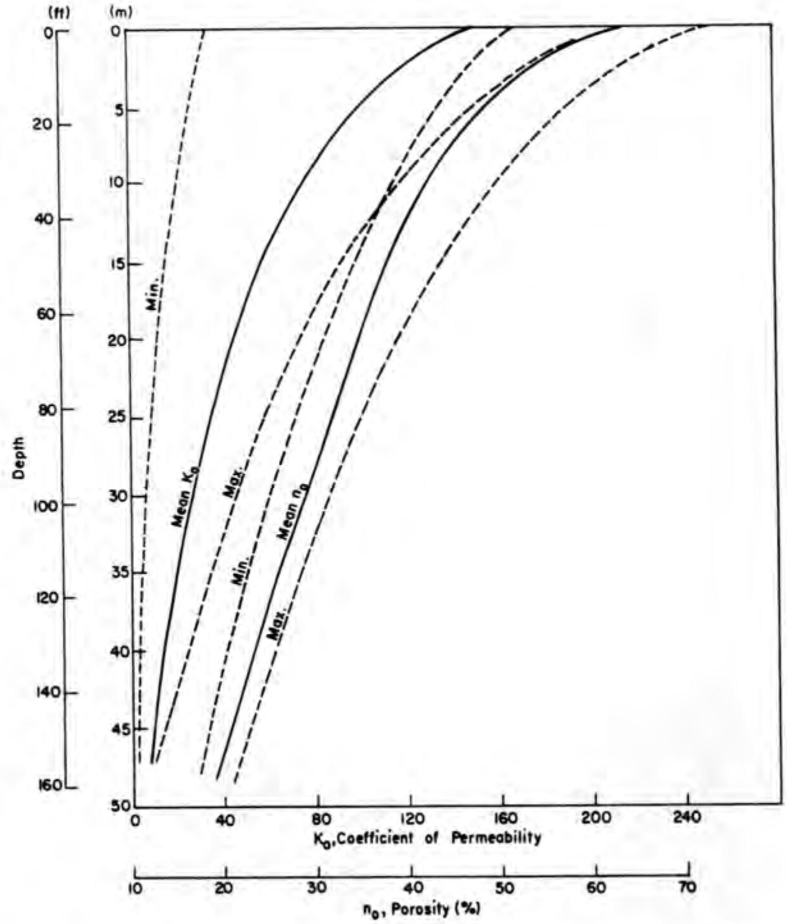

Figure IV-7. Curves showing how air permeability and porosity vary with depth below the surface of the ice cap (data from Site 2, Greenland). (After Bader et al., ref. 25)

increase of density towards the ice condition, both by plentiful material transfer and by acceleration of mechanical densification consequent upon the structural weakening at high temperature.

Pressure metamorphism is a term used to describe the mechanical densification of snow by steady compression, or compaction. It is particularly important in the absence of melting, which is the usual situation on high polar ice caps. The primary settlement, or densification, of dry snow is achieved by rearrangement of the individual grains to permit closer packing, but there is a limit to the density which can be reached in this way. Maximum densities reached by grain packing appear to be in the range 0.50-0.55 g/cm^3. Further densification proceeds by actual deformation of the grains and their bonds, with sublimation and surface diffusion processes also playing a part[1,11] The progressive reduction of pore space and specific surface finally leads to sealing-off of the pores to form closed bubbles; this happens when the density reaches 0.80-0.83 g/cm^3, corresponding to depths between about 40 and 150 m. Further compression of the impermeable ice squeezes-in the bubbles and builds up the pressure inside them.

The mechanics of snow densification

In its transition from surface snow to glacier ice, a dry snow layer is subjected to compressive creep* under ever-increasing load. Its horizontal continuity means that there is effective lateral restraint, and for the first few years, when it is near the surface, it is exposed to temperature cycles which accelerate the deformation.

If we had the time and patience, the densification process could be simulated experimentally by placing a sample of surface snow in a vertical open-end cylinder (with smooth walls), allowing it to settle under its own weight for the first year, and thereafter adding to the top of it a lead disk, equivalent in weight to one year's accumulation, every year. During any period when the weight on it is constant, the rate of densification would tend to fall off because deformation resistance increases with increasing density. The annual addition of more load, however, revives the settlement rate.

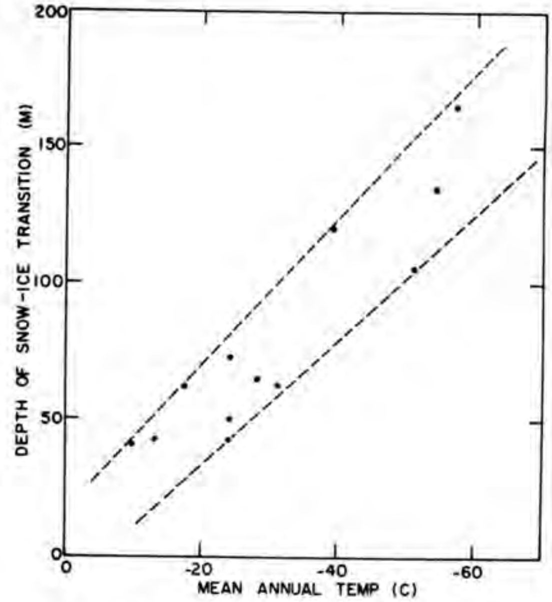

Figure IV-8. Effect of site temperature on the depth at which the transition from snow to ice occurs. The points represent observed transition depths for various sites in Greenland and Antarctica. (Compiled from various sources)

Depth-density curves give, in effect, a complete record of this compressive creep process by showing the snow layers at successive stages of deformation. Bader[2, 3, 4] has studied the densification problem in detail and has derived expressions relating depth, density, load, accumulation rate, and time, thus permitting depth-density profiles to be analyzed and yielding data on characteristic parameters which can be applied to various glaciological problems. Bader's theory is outlined below for the case where accumulation rate is assumed to be constant over a long period, i.e., the rate of load application is invariant with time.

The overburden pressure on any snow layer is equal to the weight of snow above it, which can be expressed as the weight of overlying layers or as the amount of snow accumulation since it was itself at the surface:

$$\sigma = \int_0^h \gamma \, dh = At \tag{1}$$

where σ is overburden pressure, γ is unit weight (density) of snow, $\underline{h}$ is depth of the layer, $\underline{t}$ the time since it was at the surface, and $\underline{A}$ is the snow accumulation rate (independent of time).

It follows that

$$\gamma = \frac{d\sigma}{dh} \quad \text{and} \quad \frac{dh}{dt} = \frac{A}{\gamma}.$$

Now, because the snow is densifying, its downward velocity with respect to the surface, $\frac{dh}{dt}$, is decreasing. The rate at which this velocity decreases with depth gives a "specific velocity of densification", $\underline{v}$. Thus,

* Fine-grained snow normally settles continuously, but under certain conditions layers of coarse depth-hoar resist deformation for a time and then suddenly collapse, causing the "firn quakes" sometimes reported by polar travelers.

$$v = \frac{d}{dh}\left(\frac{dh}{dt}\right) = \frac{-A}{\gamma^2} \cdot \frac{d\gamma}{dh} = \frac{-1}{\gamma}\ \frac{d\gamma}{dt}. \tag{2}$$

In the upper layers of the ice cap, where stresses are small ($\sigma < 1$ kg/cm^2), deformation is apparently governed by Newtonian viscous shear, i.e.

$$\dot{\varepsilon} = \frac{\tau}{2\eta} = -v \tag{3}$$

where $\dot{\varepsilon}$ is strain rate, τ is shear stress, and η is a viscosity coefficient (strongly dependent on density). Bader postulates, in the absence of data on the relationship between vertical and lateral stress components, that shear stress can be expressed as

$$\tau = \frac{\gamma_i - \gamma}{\gamma_i}\ \sigma \tag{4}$$

where γi is the density of hard impermeable ice.
From eq 2, 3, and 4

$$-v = \frac{A}{\gamma^2}\ \frac{d\gamma}{dh} = \frac{1}{\gamma}\ \frac{d\gamma}{dt} = \frac{\gamma_i - \gamma}{\gamma_i} \cdot \frac{\sigma}{2\eta}. \tag{5}$$

Substituting σ = At from eq 1, the following are given by eq 5

$$t = 2\ \sqrt{\frac{\gamma_i}{A}\ \int_{\gamma_0}^{\gamma} \frac{\eta\, d\gamma}{\gamma(\gamma_i - \gamma)}} \tag{6}$$

$$dh = \frac{2\ i\ \eta}{\gamma^2\ (\gamma_i - \gamma)\, t} \cdot d\gamma. \tag{7}$$

Integrating eq 7 after substitution for <u>t</u> from eq 6 gives an expression for <u>h</u> in terms of <u>A</u>, γ, γ_i and η. To replace the density-dependent viscosity coefficient η, Bader chooses

$$\eta = c\ \gamma\ e^{b\gamma} \tag{8}$$

and to allow for the acceleration of settlement caused by annual temperature cycles about the mean value he incorporates a factor $(1 - k\,e^{m\gamma_0}\ e^{-m\gamma})$. The <u>equation of the depth-density curve for the upper layers</u> then becomes

$$h = \sqrt{cA\gamma_i}\ \int_{\gamma_0}^{\gamma} \frac{e^{b\gamma}(1 - ke^{m\gamma_0}\ e^{-m\gamma})}{\gamma(\gamma_i - \gamma)\ \sqrt{\int_{\gamma_0}^{\gamma} \frac{\gamma\ e^{b\gamma}(1 - k\,e^{m\gamma_0}\,e^{-m\gamma})}{\gamma_i - \gamma} \cdot d\gamma}} \cdot d\gamma. \tag{9}$$

While this equation is somewhat ponderous it can be handled by a computer; for the range of densities $0.36 < \gamma < 0.60$ (g/cm^3) the value of $\gamma(\gamma_i - \gamma)$ is fairly constant and some simplification is possible.

In the deeper layers, where the stress condition tends towards hydrostatic, the assumption of Newtonian viscous deformation is not valid, and a hyperbolic sine relationship is adopted instead. Bader makes further assumptions to bridge the gaps in existing knowledge and, by a process of substitution similar to that outlined above and by integration, obtains another equation for the depth density curve, allowing a better fit to the data. The equation is even more ungainly than eq 9 (to which it reduces for small values of γ), but it can be programmed for a computer.

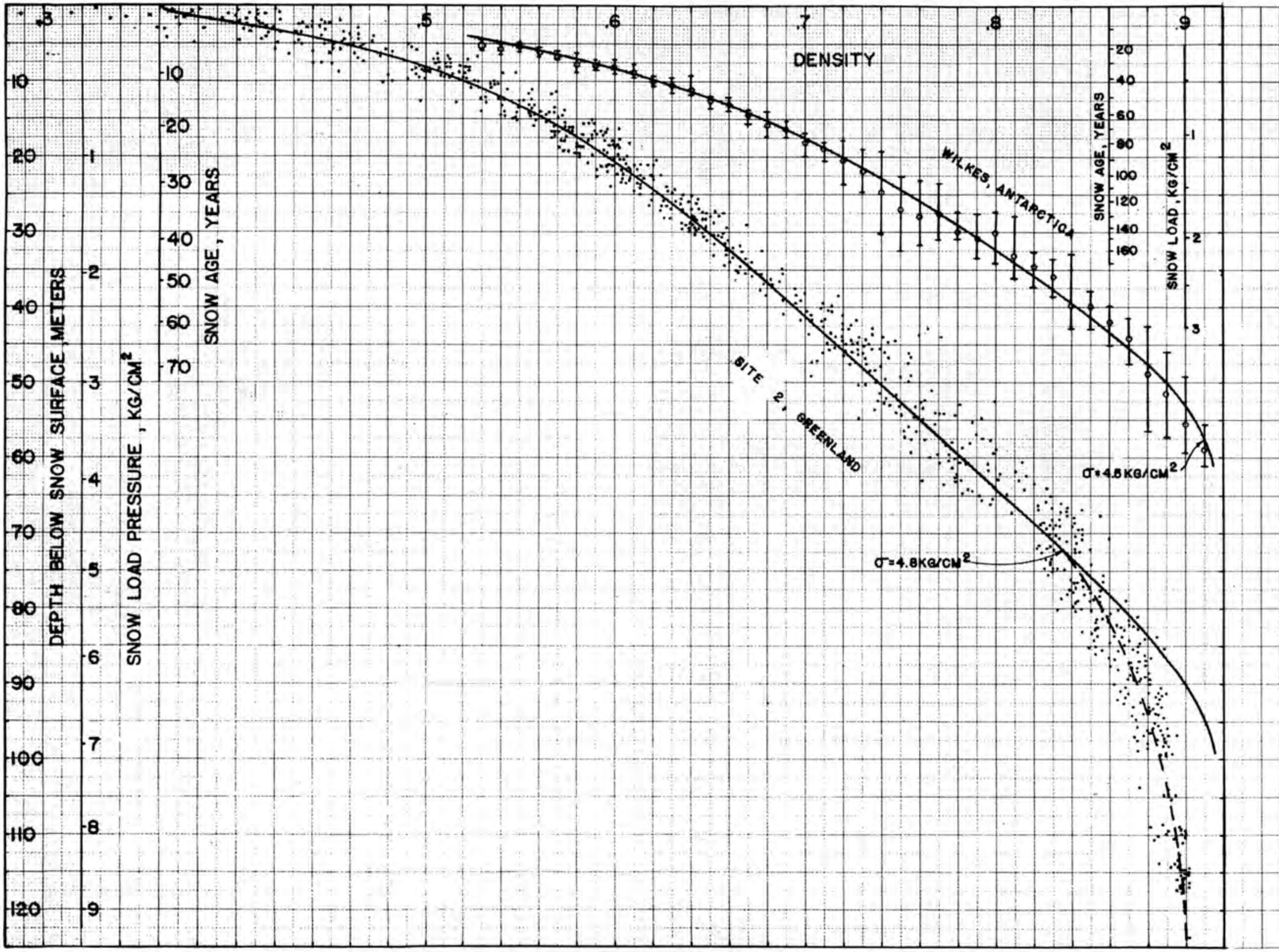

Figure IV-9. Depth-density relations for Site 2, Greenland, and Wilkes satellite station, Antarctica. The curves are calculated from Bader's theory of densification. (After Bader, ref. 5)

Bubbles in glacier ice

At the transition from snow to ice the bubbles, immediately after being sealed-off, have irregular tubular shapes, and they are randomly oriented.[15, 11] They are generally located along the intersecting edges of adjacent grains or crystals. They subsequently become more rounded by sustained hydrostatic pressure and by internal surface diffusion of molecules. In newly formed ice most of the bubbles are in the size range 0.1-1.0 mm; in older, deeper ice, however, they are compressed to smaller sizes and bubbles originally formed at grain and crystal boundaries become absorbed into the inside of crystals as a result of recrystallization and growth.

When ice is subjected to shear deformation the bubbles become drawn out into elongated tubular shapes. These tubular bubbles, each of which may consist of several of the original bubbles run together, are parallel, and their orientation corresponds with the direction of shear deformation. The effect is seen in deforming shelf ice and in the basal layers of non-melting land glaciers.

The concentration of air bubbles in glacier ice affects its color: bubbly polar ice appears white in reflected light, while layers of relatively bubble-free melt-water ice show a blue color. Melt-water ice which has been rendered almost entirely bubble-free by repeated freeze-thaw cycles often appears to be black when seen in large masses; the incident light is absorbed instead of being scattered and reflected as it is in bubbly ice.

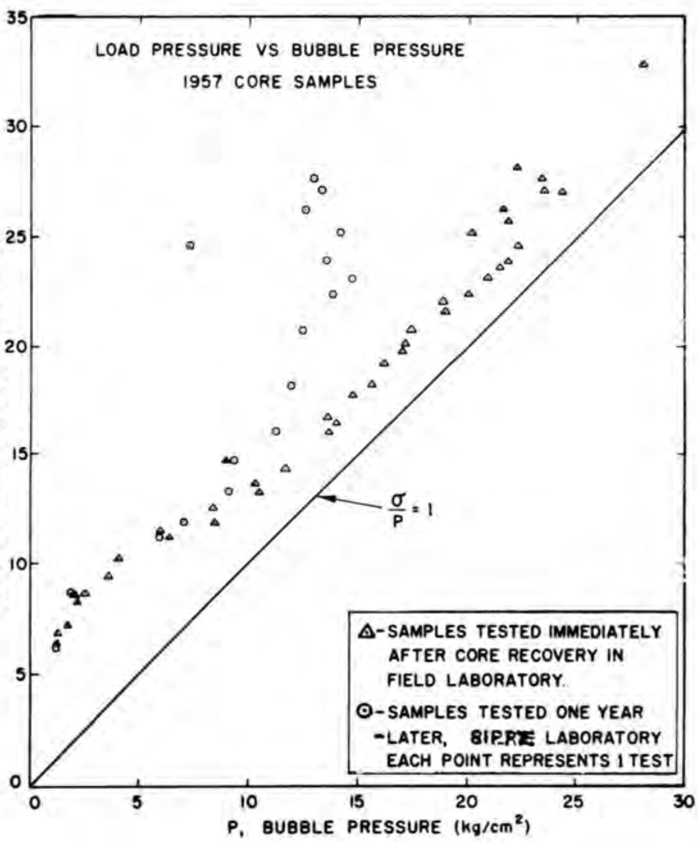

Figure IV-10. Relation between bubble pressure in polar glacier ice and the overburden pressure. (After Langway, ref. 15)

Bubble content also affects the strength and deformation resistance of glacier ice. Bubbly ice flows more easily than the denser clear ice, and it fractures more readily, particularly when the air in the bubbles is under high pressure.

When the bubbles first seal off, the air inside them is at the atmospheric pressure of the site, although the ice around them is subject to an overburden load of perhaps 3 or 4 atmospheres.[15] For this reason bubble pressure is less than the hydrostatic pressure of the ice just below the transition level, but deep down in the ice bubble pressure gradually "catches up" with the hydrostatic pressure. There is, of course, a simple relation between porosity of the ice and bubble pressure, the product of pressure and volume remaining constant if the temperature does not change (Boyle's Law). It has been found that appreciable variations of pressure can exist between different individual bubbles from the same layer;[19] this can probably be attributed to preferred directions of inter- and intra-crystalline deformation, and to the fact that all bubbles do not close off simultaneously.

There is virtually no diffusion of gas from bubbles in glacier ice, so that air samples are preserved for very long periods.[19] Careful analysis of the gases in glacier bubbles has permitted studies of the composition of ancient air to be made.

BIBLIOGRAPHY

1. Anderson, D. L. and Benson, C. S. (1963) "The densification of snow," in Ice and snow – processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

2. Bader, H. (1953) Sorge's law of densification of snow on high polar glaciers, U. S. Army Snow, Ice and Permafrost Research Establishment Research Report 2.

3. ________ (1960) Theory of densification of dry snow on high polar glaciers, USA SIPRE Research Report 69.

4. ________ (1963) "Theory of densification of dry snow on high polar glaciers, part II," in Ice and snow – processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

5. ________ (1962) "Snow as a material," in Cold Regions Science and Engineering (F. J. Sanger, Editor), U. S. Army Cold Regions Research and Engineering Laboratory monographs, Part II, Sect. B.

6. ________ ; Haefeli, R.; Bucher, E,; Neher, J.; Eckel, O., and Thams, Chr. (1939) Der Schnee und seine Metamorphose (Snow and its metamorphism), Beitrage zur Geologie der Schweiz, Geotechnische Serie, Hydrologie, Lieferung 3. USA SIPRE Translation 14, 1954, 313 p.

7. Bender, J. A. (1957) Air permeability of snow, USA SIPRE Research Report 37.

8. Benson, C. S. (1960) Stratigraphic studies of the snow and firn of the Greenland ice sheet, USA CRREL Research Report 70.

9. De Quervain, M. R. (1963) "On the metamorphism of snow," in Ice and snow – processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

10. Gow, A. J. (1961) Drill hole measurements and snow studies at Byrd Station, USA SIPRE Technical Report 78.

11. ________ (in press) "The Antarctic ice sheet" in Antarctica. London: Methuen.

12. Kuroiwa, D. (1962) A study of ice sintering, USA CRREL Research Report 86.

13. Landauer, J. K. (1957) On the deformation of excavation in the Greenland névé, USA SIPRE Research Report 30.

14. ________ (1958) Creep of snow under combined stress, USA SIPRE Research Report 41.

BIBLIOGRAPHY (Cont'd)

15. Langway, C. C. (1958) "Bubble pressures in Greenland glacier ice," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

16. Mellor, M. (1963) "Polar snow — a summary of engineering properties," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

17. Nakaya, U. (1959) Visco-elastic properties of snow and ice from the Greenland Ice Cap, USA SIPRE Research Report 46.

18. Rigsby, G. P. (1960) Crystal orientation in glacier and experimentally deformed ice, Journal of Glaciology, vol. 3, no. 27.

19. Scholander, P. F. and Nutt, D. C. (1960) Bubble pressures in Greenland icebergs, Journal of Glaciology, vol. 3, no. 28.

20. Schytt, V. (1958) Glaciology II, Scientific Results of the Norwegian-British-Swedish Antarctic Expedition, vol. IV, Oslo.

21. Shumskii, P. A. (1955) Osnovy strukturnogo ledovedeniia (Fundamentals of ice structure). Moscow: Izdatel'stvo Akademii Nauk.

22. Stephenson, P. J. and Lister, H. (1959) Preliminary results of the glaciological work on the Trans-Antarctic Expedition, Journal of Glaciology, vol. III, no. 25.

23. Waterhouse, R. W. (1962) "On the permeability of snow in the accumulation zone of polar regions," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

24. Yosida, Z. (1962) "Physical properties of snow," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

25. Bader, H., et al. (1955) Excavations and installations at SIPRE test site, Site 2, Greenland, USA SIPRE Report 20.

26. Bender, J. A. and Gow, A. J. (1960) "Deep drilling in Antarctica," in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55.

CHAPTER V. GLACIER FLOW

General facts

The movement of glaciers has intrigued philosophers and scientists for over a century, but it is only in the last decade that a firm quantitative theory of glacier flow has appeared. In the mid-nineteenth century it was not understood how an apparently solid mass of ice could flow, but many of the facts of glacier movement had been closely observed in the Alps, as is shown by Mark Twain's attempt to travel by glacier in Switzerland:*

"I marched the expedition down the steep and tedious mule-path and took up as good a position as I could upon the middle of the glacier — because Baedeker said the middle part travels the fastest. As a measure of economy, however, I put some of the heavier baggage on the shoreward parts, to go as slow freight".

After camping overnight waiting for the glacier to start, it was suspected that it might be aground and leaking, and finally the book was again consulted:

"Presently Baedeker was found again, and I hunted eagerly for the time-table. There was none. The book simply said the glacier was moving all the time. This was satisfactory, so I shut up the book and chose a good position to view the scenery as we passed along. I stood there some time enjoying the trip, but at last it occurred to me that we did not seem to be gaining any on the scenery. I soon found a sentence which threw a dazzling light on the matter. It said, 'The Gorner Glacier travels at an average rate of a little less than an inch a day'. I made a small calculation: 1 inch a day, say 30 feet a year; estimated distance to Zermatt, 3 1/18 miles. Time required to go by glacier, a little over five hundred years! I said to myself, 'I can walk it quicker — and before I will patronize such a fraud as this, I will do it!'"

Since that time glaciers have been closely studied in all parts of the world, and their speeds have been measured both on the surface and within the ice mass. At any cross-section of a valley glacier, surface velocity is greatest in the center as a result of shear resistance along the edges. Figure V-1 gives an example of velocity distribution along a transverse section.

Velocity varies along the length of a valley glacier, with shape, size and slope of the channel affecting speed (somewhat like a river). At a particular place it also varies with time. As the ice descends it is thinned by melting and its flow properties are changed by temperature rise; these factors also affect the longitudinal distribution of velocity. Figures V-2 and V-3 give longitudinal and transverse velocity distributions on the surfaces of valley glaciers.

In a valley glacier which terminates on land, velocity tends to increase with distance down-glacier in the accumulation zone, and further down still, in the ablation zone, it tends to decrease with distance. The condition where ice velocity increases with distance, and the ice is tending to thin, is referred to as 'extending flow'. The opposite situation, where velocity decreases with distance, is called 'compressive flow' (see Fig. V-5). Extending flow is often indicated by the presence of crevasses.

The distribution of velocity through a vertical section also must be considered. Figures V-6-10 show velocities measured at various depths in boreholes, and it can be seen that the ice moves fastest at the surface, differential movement resulting from shearing of the ice mass over the rock bed. The movement observed at the surface usually represents the sum of two separate mechanisms: shear deformation of the ice itself and slip of the ice over its bed. This is shown schematically in Figure V-11.

Many small valley glaciers move with surface velocities of only a few inches per day, while larger valley glaciers in mountain regions may have velocities of a few feet per day. In general the most rapid glaciers are the giant outflow glaciers of the Greenland ice sheet, which travel tens of feet per day. The big ice streams and valley glaciers of the Antarctic ice sheet are not so swift, partly due to the lower temperatures which keep the ice "stiffer". Table V-I gives a general impression of glacier speeds in various parts of the world.

* "Innocents Abroad" by Mark Twain

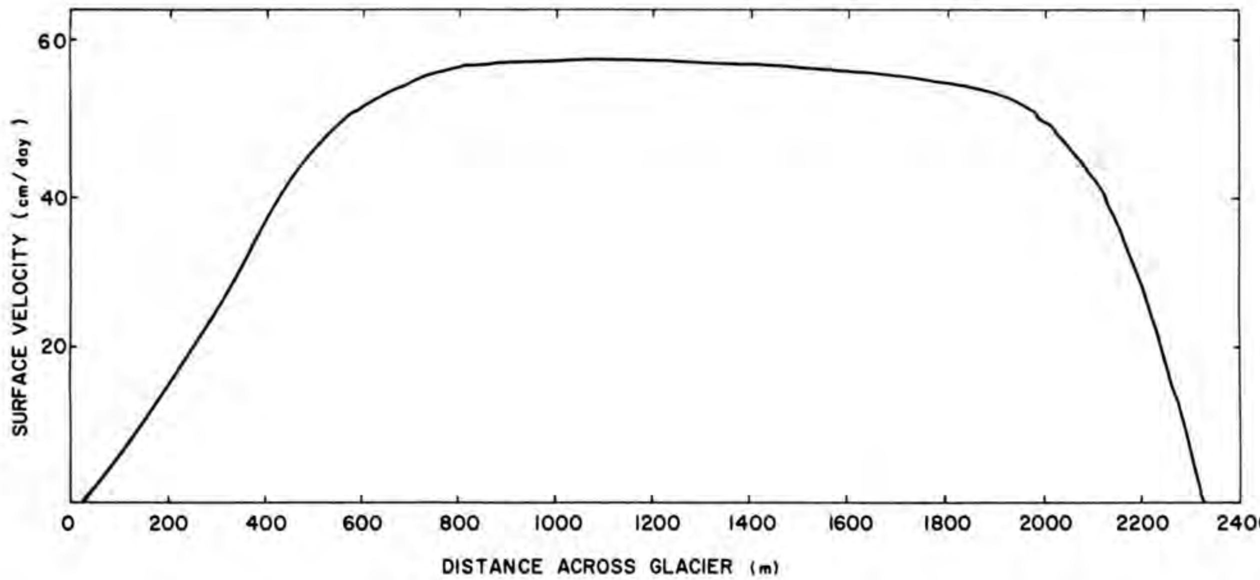

Figure V-1. Distribution of surface velocity across a section of the Fedchenko Glacier in the Pamir. The distribution of surface velocity across a glacier depends on the stress and temperature fields in the ice; in general, narrow glaciers exhibit U-shaped or parabolic patterns, while very wide glaciers tend to have the shear concentrated near the edges, with fairly uniform velocities across the central section. (After Souslov and Nozdryukhin, ref. 34)

In the preceding paragraphs, transverse, longitudinal, and vertical distribution of velocity in a glacier has been described, but no discussion has been made so far of variations of velocity with time, for a given point. Because of fluctuations in rates of accumulation and ablation, or temperature changes in the ice, ice velocity and the rate of discharge vary in a rather complex manner. A sudden increase of accumulation at the head of a glacier, for example, will cause a "wave" of increased discharge to travel down the glacier at a higher speed than the glacier itself is moving. These traveling-wave phenomena are believed to account for a number of previously unexplained features of glacier behavior, such as sudden advances and retreats of snouts, irregular changes of ice thickness, and certain types of transverse pressure waves. The theory of traveling waves, or "kinematic waves" is discussed later.

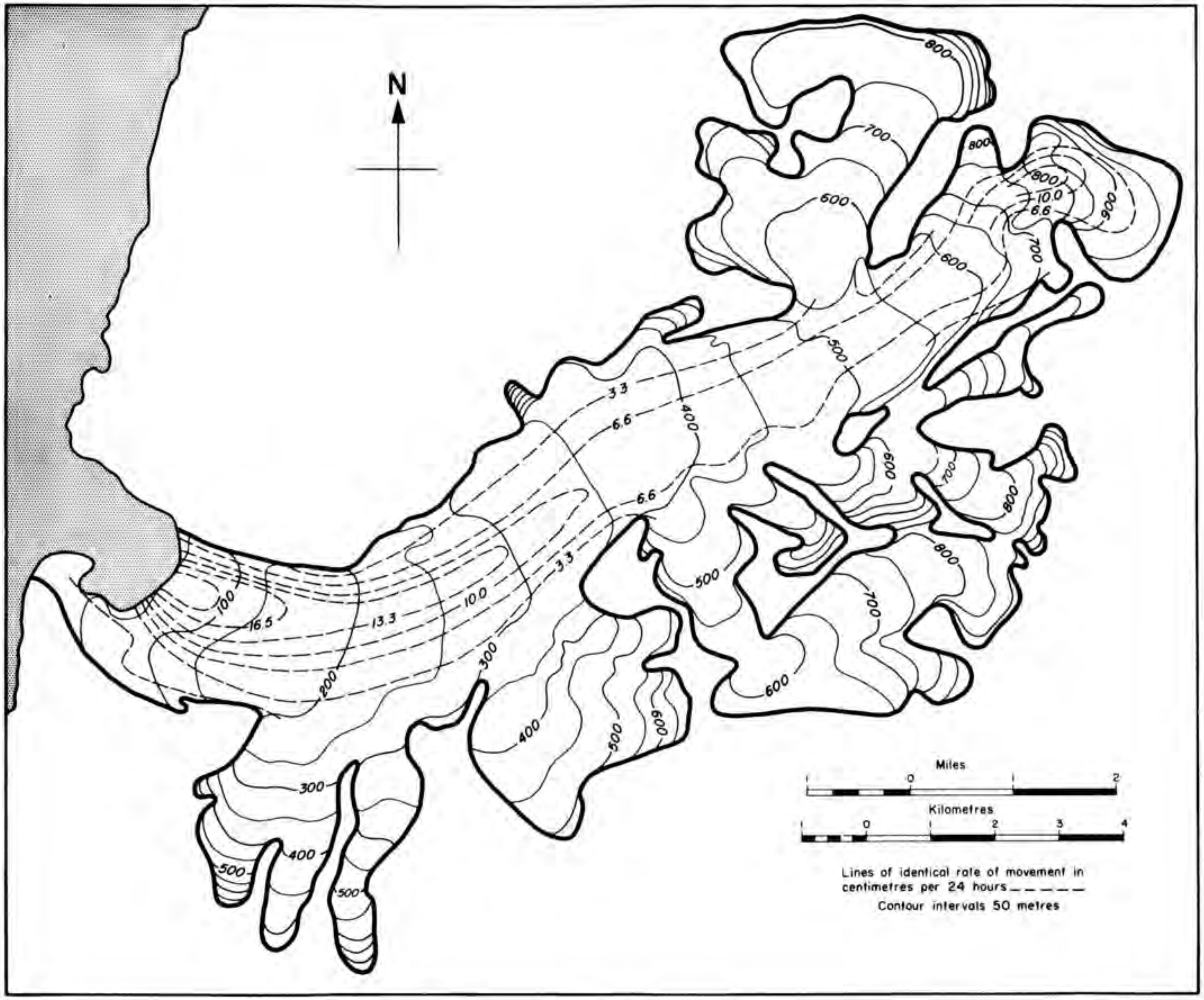

Figure V-2. Isotachytes (lines of equal velocity) showing distribution of surface velocity on the Fourteenth of July Glacier, Spitsbergen. Note the general increase of surface velocity towards the terminus; this indicates extending flow. (After Ahlmann, ref. 45)

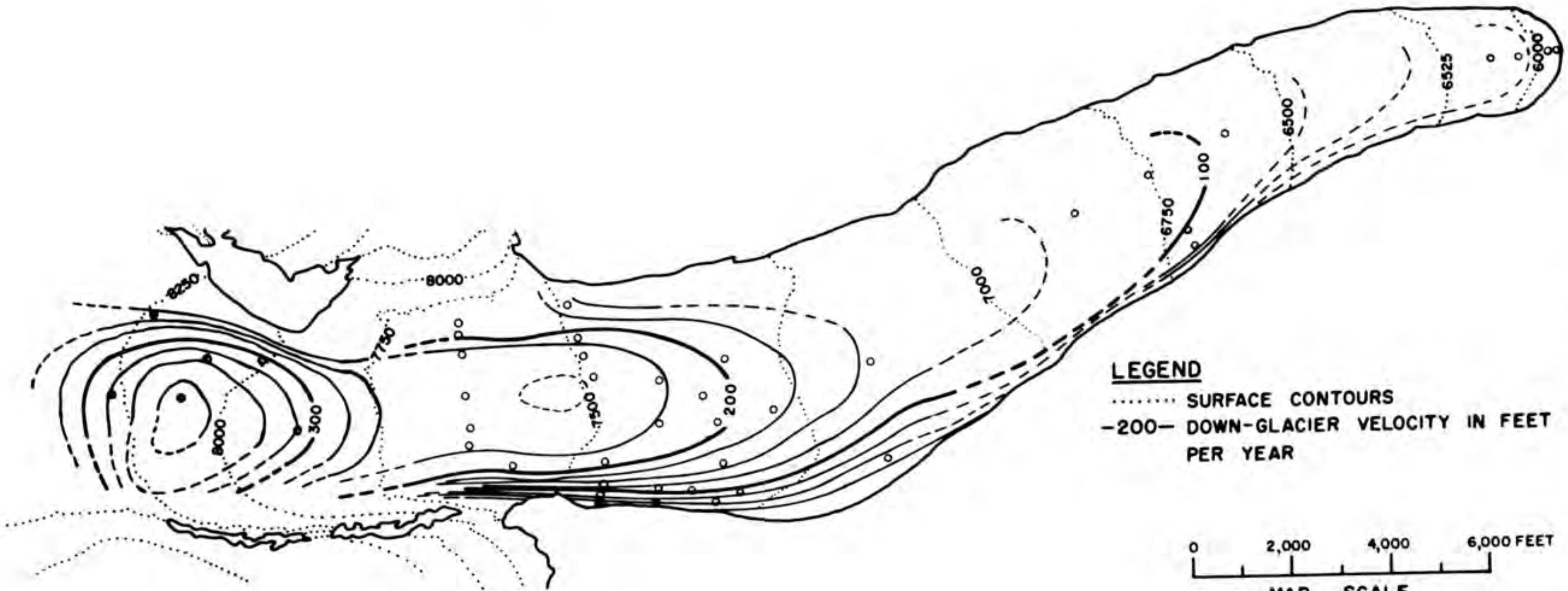

Figure V-3. Distribution of surface velocity on the Saskatchewan Glacier, Alberta. The decrease of velocity towards the snout indicates compressive flow (cf. Fig. V-2). (After Meier, ref. 15)

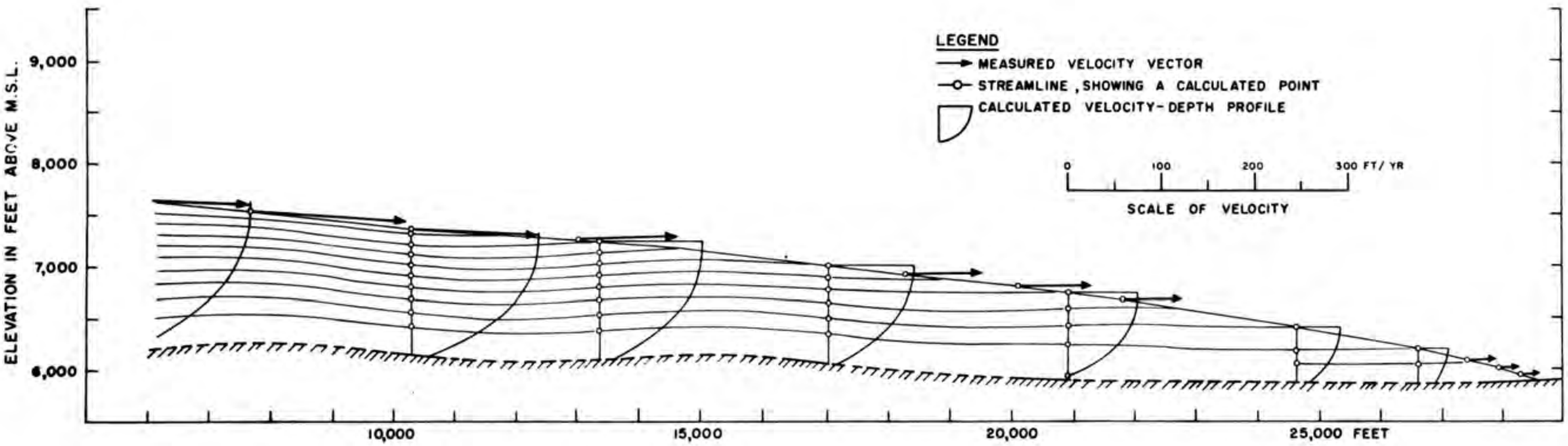

Figure V-4. A longitudinal section of the Saskatchewan Glacier showing surface velocity vectors with calculated flow lines and velocity-depth profiles in the body of the ice. (After Meier, ref. 15)

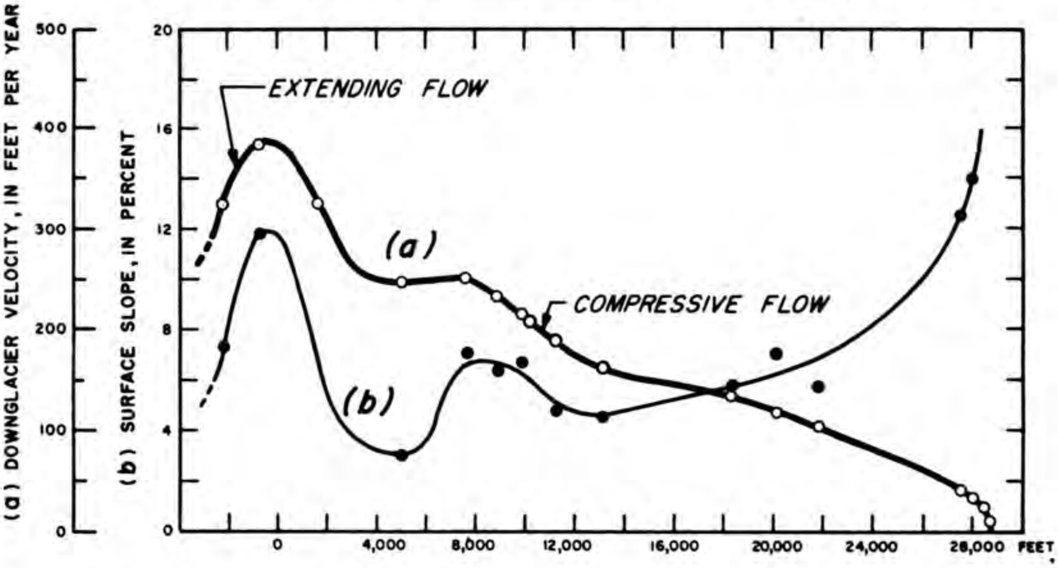

Figure V-5. Velocity variations along the length of a glacier. (After Meier, ref. 15)

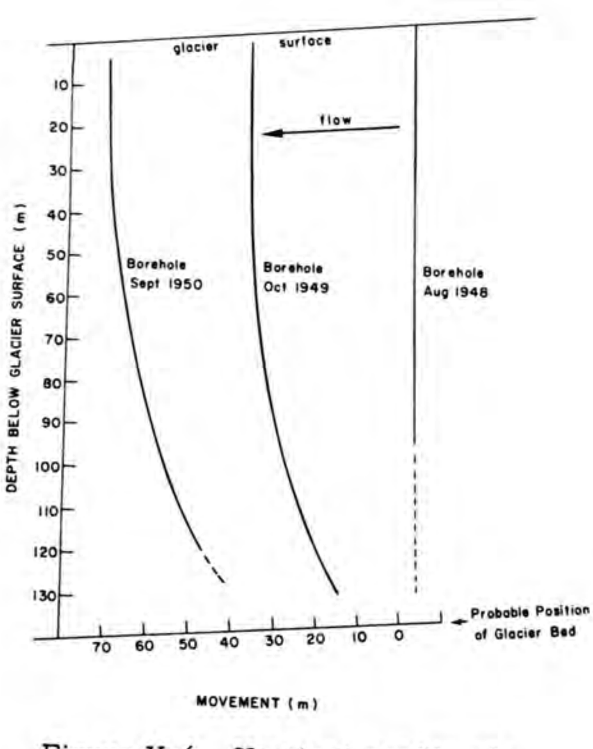

Figure V-6. Vertical profile of a borehole on the Jungfraujoch, Switzerland. Progressive deformation of the borehole gives the velocity distribution. (After Gerrard, Perutz and Roch, ref. 9)

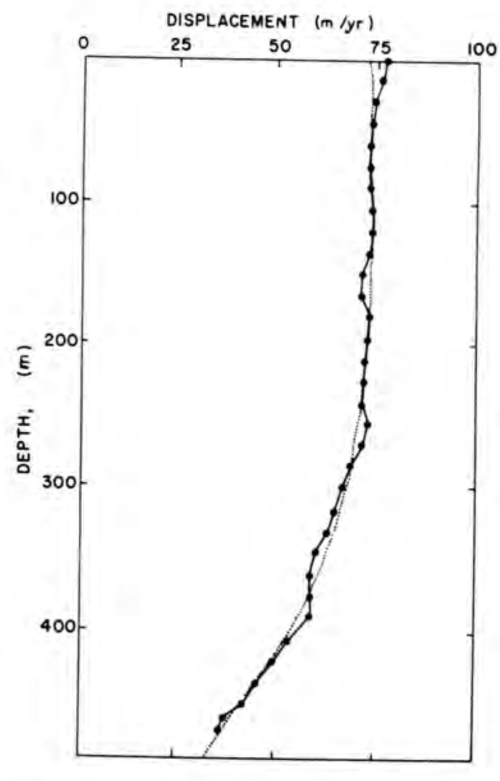

Figure V-7. Vertical profile of velocity from a borehole in the Salmon Glacier, B. C. (After Mathews, ref. 14)

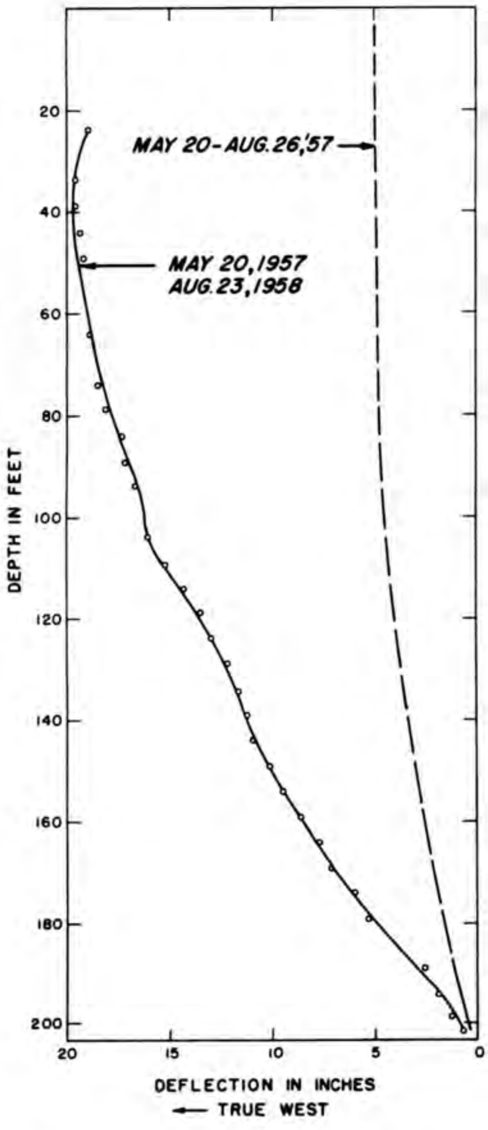

Figure V-8. Deformation of a borehole drilled into the Tuto Ramp, an ablating ice slope at the edge of the Greenland Ice Cap. (After Wilson, ref. 46)

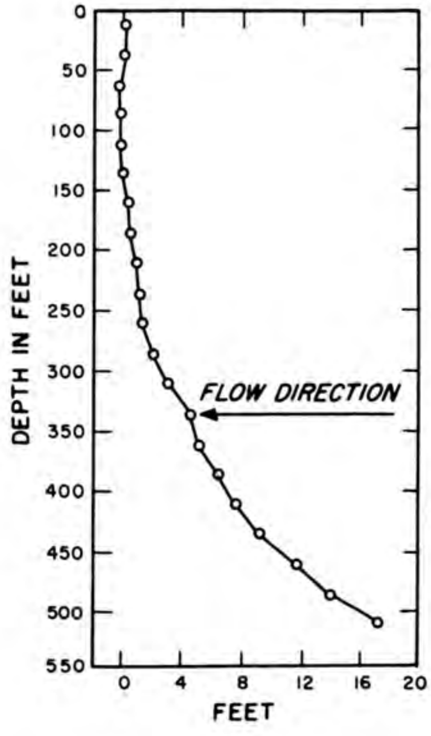

Figure V-9. Vertical profile of velocity in the Blue Glacier on Mount Olympus, Washington. (After Sharp, ref. 32)

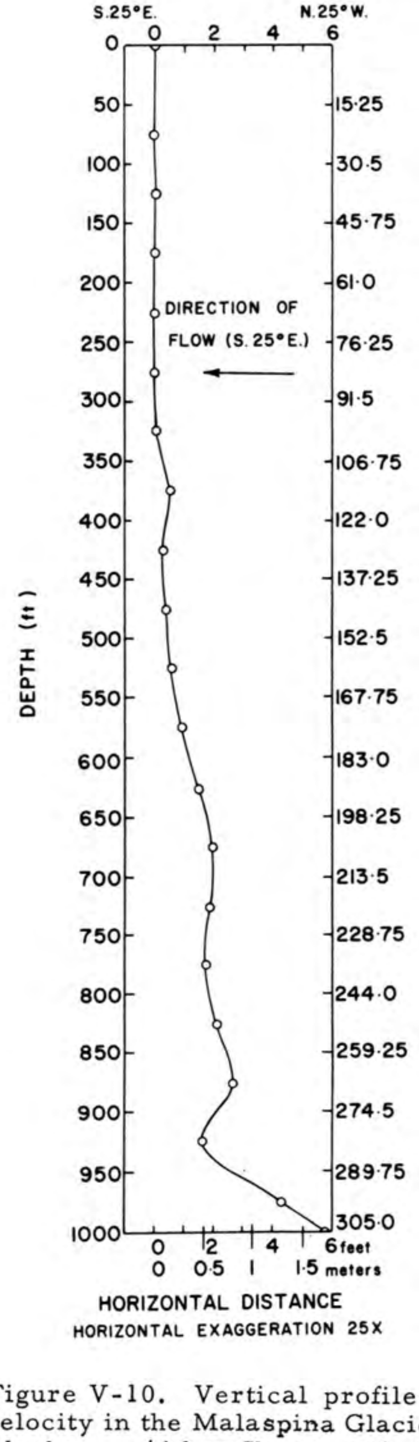

Figure V-10. Vertical profile of velocity in the Malaspina Glacier, Alaska. (After Sharp, ref. 31)

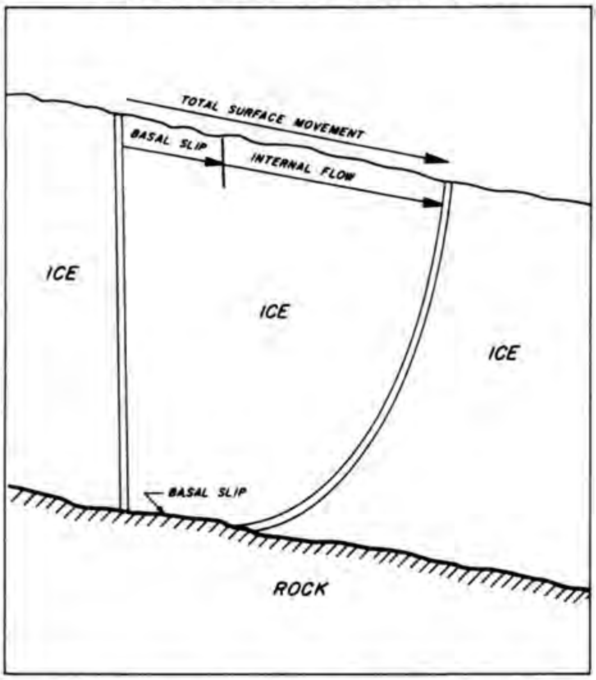

Figure V-11. Schematic illustration of the two flow components — sliding of the ice mass on the rock bed, and internal shearing of the ice mass. (After Sharp, ref. 32)

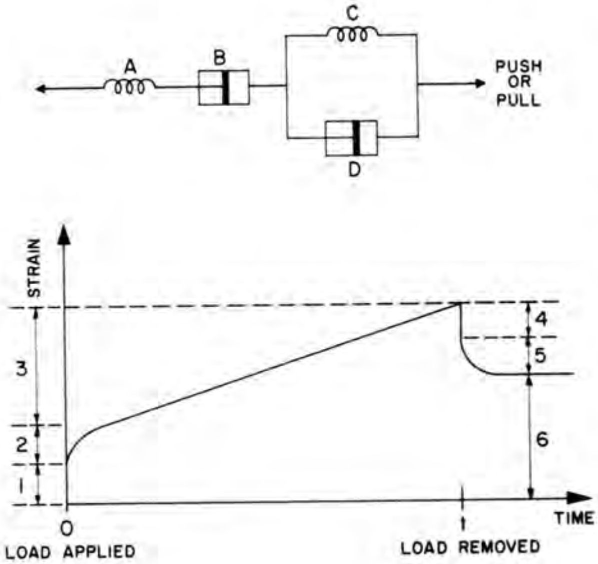

Figure V-12. Ice subjected to a given stress behaves as a visco-elastic material, whose response to loading can be illustrated by the spring and dashpot analog shown above (combination of Maxwell and Voigt rheological units in series).

The numbered stages of the strain time curve (creep curve) correspond to the action of the analog as follows:

Stage 1 - On application of load, instantaneous elastic response occurs (spring A extends or compresses).
Stage 2 - Creep begins, and the "delayed elastic" response occurs (dashpot D slides, but comes to a stop as spring C extends or compresses; dashpot B slides at a steady rate).
Stage 3 - Creep, or quasi-viscous flow, proceeds at a steady rate (dashpot B slides, A, C & D are immobile)
Stage 4 - On removal of load, instantaneous elastic recovery occurs (spring A unloads).
Stage 5 - Relaxation of the "delayed elastic" component of strain occurs (spring C relaxes, but return is damped by dashpot D).
Stage 6 - The permanent strain resulting from creep up to time t. (The amount of travel which occurred in dashpot B.)

Table V-1. Surface velocities on glaciers.
(Compiled from many sources)

Greenland	Type of flow	Approx. measured velocity
Thule Ramp, N. W. Greenland	Sheet flow	$\frac{1}{4}$-$\frac{1}{2}$ in/day
Nunatarssuaq ice cliff, N. W. Greenland	Sheet flow	$\frac{1}{2}$ in/day
Blue Ice Valley, N. W. Greenland	Sheet flow	$\frac{1}{4}$ in/day
Rink Glacier, W. Greenland	Outflow glacier	88 ft/day
Jakobshavn Glacier, W. Greenland	Outflow glacier	66 ft/day
Karajak Glacier, W. Greenland	Outflow glacier	59 ft/day
Upernavik Glacier, W. Greenland	Outflow glacier	59 ft/day
Itivdliarssuk Glacier, W. Greenland	Outflow glacier	46 ft/day
Sigssortartoq Glacier, W. Greenland	Outflow glacier	37 ft/day
Torssukatak Glacier, W. Greenland	Outflow glacier	26 ft/day
Kanigigdleq Glacier, W. Greenland	Outflow glacier	16 ft/day
Umiamako Glacier, W. Greenland	Outflow glacier	13 ft/day
Eqip Sermia, W. Greenland	Outflow glacier	10 ft/day
Frõya Glacier, E. Greenland	Small independent glacier	3 ft/day
Sefstrøms Glacier, N. E. Greenland	Valley glacier	1 ft/day
Antarctica		
Terre Adélie coast	Sheet flow	3 in/day
MacRobertson Land coast	Sheet flow	2 in/day
Cape Folger	Sheet flow	6 in/day
Dovers Glacier, MacRobertson Land	Ice stream	7 ft/day
Hoseason Glacier, MacRobertson Land	Ice stream	3 ft/day
Jelbart Glacier (west tongue), MacRobertson Land	Ice stream	3 ft/day
Jelbart Glacier (east tongue), MacRobertson Land	Ice stream	1 ft/day
Taylor Glacier, MacRobertson Land	Ice stream	1 ft/day
Helen Glacier, Queen Mary Land	Ice stream	5 ft/day
Denman Glacier, Queen Mary Land	Ice stream	9 ft/day
Scott Glacier, Queen Mary Land	Ice stream	4 ft/day
Reed Glacier, Queen Mary Land	Ice stream	4 ft/day
Obruchev Glacier, Queen Mary Land	Ice stream	3 ft/day
Apfel Glacier, Queen Mary Land	Ice stream	3 ft/day
Mirny, Queen Mary Land	Ice stream	2 ft/day
Gaussberg, Kaiser Wilhelm II Land	Ice stream	1 ft/day
Vanderford Glacier, Wilkes Land	Ice stream	6 ft/day
Glacier de Zélée, Terre Adélie	Ice stream	3 ft/day
Glacier de l'Astrolabe	Ice stream	$4\frac{1}{2}$ ft/day
Skelton Glacier	Ice stream	1 ft/day
Maudheim Ice Shelf, Dronning Maud Land	Shelf movement	3 ft/day
Ross Ice Shelf, Ross Dependency	Shelf movement	3-4 ft/day
Russell East Glacier, Trinity Peninsula	Valley glacier	21 in/day
Victory Glacier, Trinity Peninsula	Valley glacier	10 in/day
Depot Glacier, Trinity Peninsula	Valley glacier	3 ft/day
Spitsbergen		
Fourteenth of July Glacier	Valley glacier	3-6 in/day
Blomstrandbreen	Valley glacier	7 in/day
Kongsbreen (north section)	Valley glacier	17 in/day
Kongsbreen (south section)	Valley glacier	6 in/day

Table V-I. (Cont'd) Surface velocities on glaciers.

	Approx. measured velocity
Alaska	
Taku Glacier	2 ft/day
Yakatag Glacier	1 ft/day
Kennicott Glacier	1-2 in/day
Lemon Creek Glacier	2-5 in/day
Eldridge Glacier	9-18 in/day
Arctic and Sub-arctic islands	
Highway Glacier, Baffin Island	7 in/day
Weyprecht Glacier, Jan Mayen	10 in/day
Kjerulfbreen, Jan Mayen	8 in/day
Hoffell Glacier, Iceland	3-6 ft/day
Morsarjökull, Iceland	11 in/day
Svinafellsjökull, Iceland	16 in/day
Skaftafellsjökull, Iceland	14 in/day
Scandinavia	
Karsa Glacier, Swedish Lapland	3 in/day
Styggedal Glacier, Jotunheim, Norway	3 in/day
European Alps	
Aletsch Glacier	4-20 in/day
Unteraar Glacier	3-10 in/day
Upper Grindelwald Glacier	4-18 in/day
Mont Collon Glacier	6 in/day
Vernagtferner	1 in/day
Hintereisferner	4 in/day
Guslarferner	1 in/day
Pasterze Glacier	6 in/day
Rhone Glacier	2 ft/day
Glacier des Bossons	3 ft/day
U.S.S.R.	
Dzhilky-augan-dhiran, Caucasus	3 in/day
Shkhel'dy Glacier, Caucasus	5 in/day
Bashkara Glacier, Caucasus	3 in/day
Devdorak Glacier, Caucasus	1-4 in/day
Fedchenko Glacier, Pamir	10-30 in/day
Constitution Glacier, Tien Shan	1 in/day
Shokal'skiy Glacier, Novaya Zemlya	1 in/day
New Zealand	
Tasman Glacier	16 in/day
Franz Josef Glacier	5-6 ft/day
Canada & U.S.A.	
Salmon Glacier, British Columbia	1-8 in/day
Saskatchewan Glacier, Alberta	8 in/day
Blue Glacier, Washington	10 in/day
Sub-Antarctic	
Baudissin Glacier, Heard Island	1 ft/day
Hamberg Glacier, South Georgia	up to 7 ft/day
Karakorum, Kashmir	
Baltoro Glacier	9 in/day
Kuthiah Glacier	9-24 in/day

Flow properties of ice

Early theories of glacier flow were based on the assumption that ice behaves as a Newtonian fluid, in which strain rate is proportional to the applied stress. Coefficients of viscosity were deduced and quantitative theory was developed by use of the Navier-Stokes equations. Some glacier features, however, could not be explained by purely viscous flow.

More recently, glacier movement was analyzed on the assumption that ice is perfectly plastic, i.e., that it has a sharply defined yield point. This permitted adoption of established plasticity theory and more realistic considerations were possible. Although this theory was a big improvement on previous ideas it was not fully acceptable, since it has been shown that ice continuously deforms under the smallest stresses (below the formerly assumed yield stress) and cannot, therefore, be ideally plastic.

Current theory has been developed from that evolved for the ideal plastic case, the earlier treatment being adapted to conform with a non-linear visco-plastic flow law for ice.

Laboratory creep tests have been made on specimens of polycrystalline ice, and creep curves similar to those obtained from metals have been drawn. It is seen from the schematic curve of Figure V-12 that ice is a visco-elastic material which on being subjected to load has an immediate elastic deformation, followed by a "delayed elasticity" stage where strain rate decelerates, and eventually by a stage of creep at a steady rate. Ultimately, deformation may enter the tertiary creep stage, where the strain rate accelerates to the point of destruction. If the load is removed at any stage of the creep process the small elastic fraction of the total strain is recovered immediately, and the part of the strain due to "delayed elasticity" is slowly relaxed, but a residual deformation from viscous flow remains.

If a cold room is available (large refrigerator, cold store, unheated shed in subfreezing weather), the creep behavior of ice can be illustrated by a simple experiment. A neatly squared ice cube is placed on a firm table and a heavy block, e.g., a 5-lb weight, is carefully laid on top of it. An engineer's dial gage clamped to a retort stand is made to bear on the weight with its piston compressed, so that the dial unwinds as the ice is compressed by the weight. With dial readings plotted as ordinate and time as abcissa, a curve of the form described above will be obtained. The slope of the straight "secondary creep" portion of the curve gives the creep rate (the minimum strain rate) which is of most significance in problems concerning sustained loading.

If the size of the weight is varied, the creep rate changes. More refined laboratory experiments, in which samples are subjected to compression, tension, or shear, lead to relationships between strain rate and stress. The most frequently used expression is a simple (but not physically rigorous) power law:[4, 10, 11]

$$\dot{\gamma} = k\,\tau^{n} \tag{1}$$

where $\dot{\gamma}$ = strain rate

τ = shear stress

k = constant for a given temperature and ice type

n = an exponent, assumed constant for a limited range of stress, but actually stress-dependent.

As long as the stresses are less than about 1 kg/cm^2 (15 psi), the value of the exponent $\underline{n}$ is approximately equal to 1, i.e., the ice behaves like a Newtonian viscous fluid. With stresses above 1 kg/cm^2 this ceases to be true, and $\underline{n}$ takes on higher values (Fig. V-13). Experiments using stresses in the range 2-15 kg/cm^2 show that the power $\underline{n}$ ranges from 2.5 to 4.5 for these greater loadings, and for this reason the rheological behavior of ice is termed visco-plastic (for a Newtonian viscous fluid the value of $\underline{n}$ would be 1; for a perfectly plastic material $\underline{n}$ would be infinite).

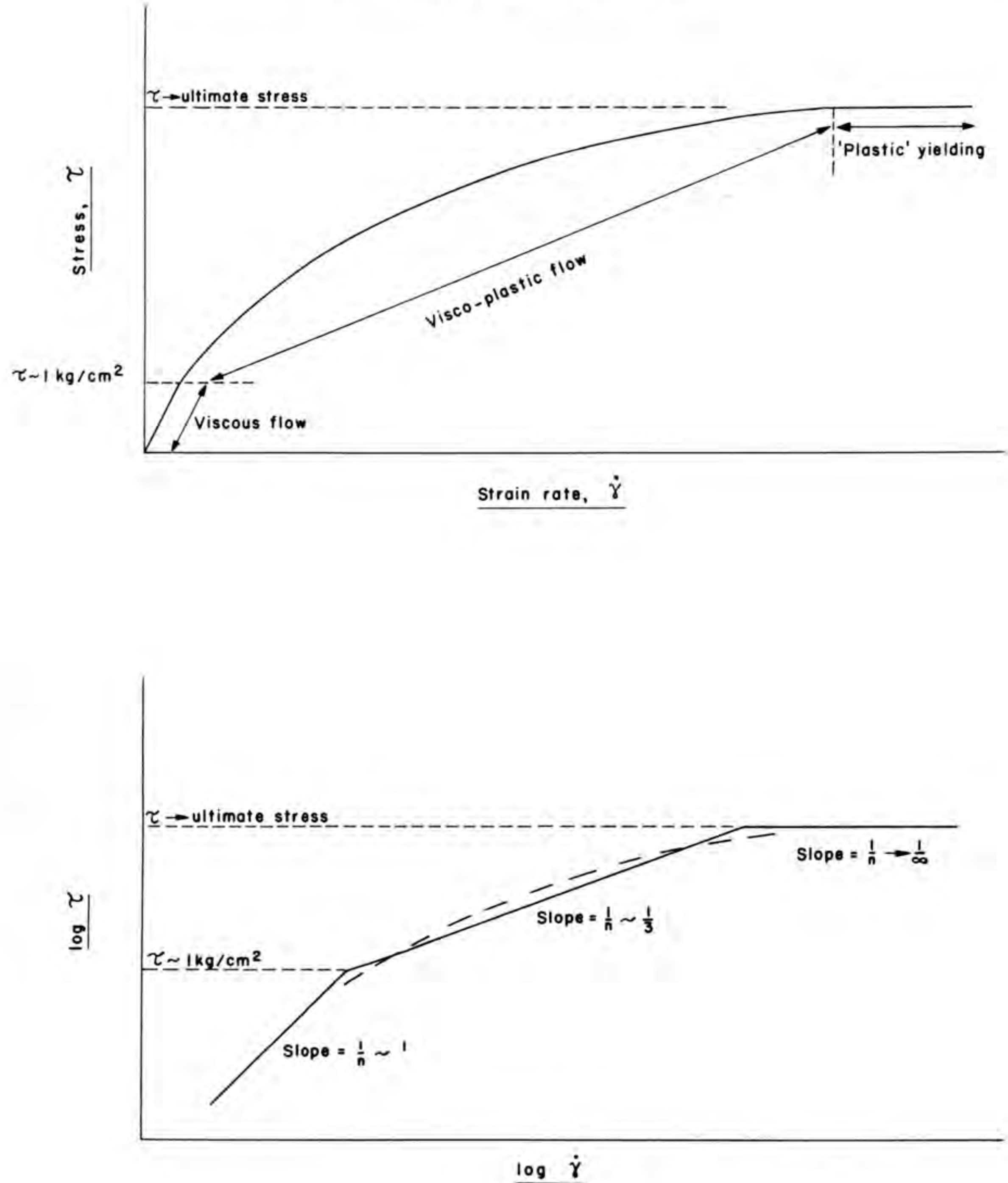

Figure V-13. Schematic relationships between stress and strain rate for polycrystalline ice. The figures illustrate how ice apparently changes from a viscous material at low stresses, to a plastic material at high stresses, thus demanding change of the exponent in a power flow law. On the upper figure a Newtonian fluid would be represented by a sloping straight line passing through the origin, and an ideal plastic would be represented by a straight line parallel to the strain rate axis.

A power law in which the exponent is stress-dependent is inelegant and inconvenient, and attempts have been made to establish other flow laws. The most promising alternative is a hyperbolic sine relationship of the form:[4,5]

$$\dot{\gamma} = A \sinh\left(\frac{\tau}{\tau_0}\right) \tag{2}$$

where $\underline{A}$ and τ_0 are constants for a given ice type at a given temperature. The experiments of Butkovich and Landauer (1959) at relatively high stresses gave results favoring the power law rather than the hyperbolic sine law, but their later (1960) work on creep at low stresses reversed this finding.

Going back to the simple experiment with the weight on an ice cube, it can be shown that the creep rate changes if the temperature of the cold room is changed, the ice deforming more rapidly at high temperatures than it does at lower temperatures. The temperature effect can be described by an exponential equation:[10,11]

$$\dot{\gamma} = B \exp\left(-\frac{Q}{RT}\right) \tag{3}$$

where B = constant for a given stress

R = the gas constant = 1.985 cal/mole °C(°K)

Q = the activation energy of ice (the best experimental values for glacier ice seem to be around 16,000 cal/mole)

T = absolute temperature (°K).

This expression cannot be applied when temperatures are close to the melting point, as it takes no account of the physical changes occurring near and at that temperature. Below about -5C, however, it appears to be satisfactory.

So far no attempt has been made to describe the ice which is being deformed, although experiments show that the character of the ice (bubble content, "dirt" content, crystal orientation) influences its rate of deformation at any given stress and temperature.

If ice samples with strongly preferred crystal orientation (c-axes parallel) are sheared in an apparatus of the type seen in Figure V-14, it is to be expected that deformation will be more rapid with the c-axes perpendicular to the shear direction than with them parallel to the shear direction.[29,33] This is mainly because the basal planes of the hexagonal ice crystals are the only glide, or cleavage, planes in a mass of polycrystalline ice. A corollary to this is the development of preferred crystal orientation in ice which is sheared over a long period. As a mass of ice is stressed, the crystal lattices are distorted, and recrystallization takes place, lightly stressed crystals growing at the expense of their more highly stressed neighbors. The lightly stressed crystals are those oriented most favorably for glide along basal planes, and thus there is a gradual reorientation until most of the crystals have their basal planes parallel to the direction of flow. If samples are taken from ice at the base of a glacier, where shear stresses are greatest, it is found that the crystals have a strongly preferred orientation, the c-axes being perpendicular to the plane of shear (i.e., perpendicular to the glacier bed).

Inclusions of rock particles (e.g., moraine) in ice affect the rate of deformation. Moderate concentrations of "dirt" appear to increase the mobility of the ice and allow it to deform faster. However, there must be a limit to this effect, and when the dirt concentration is very high the mixture of ice and rock will be "stiffer" than pure ice (the mixture would, of course, then be called frozen soil rather than "dirty ice").

The air bubbles in ice from very cold glaciers also permit that ice to deform more easily than solid, bubble-free ice.

Stresses and velocities

The deformation of a large ice mass depends on the rheological properties of the ice, the shape and dimensions of the ice mass, the nature of the bed, and the temperature distribution. In general, the surface motion of a glacier is the result of two separate movements: differential shear within the ice, and sliding of the ice over its bed. In thick polar ice sheets, differential motion between succeeding "horizontal" layers in the ice is small, most of the movement taking place by shear concentrated in a relatively thin basal layer, and by sliding on the bed (if the base temperature is close to the pressure melting point). Thinner and warmer glaciers are subject to appreciable internal shear throughout their entire depth, but sliding over the bed is probably responsible for much of the gross movement.

In recent years much has been done to clarify thinking on the distributions of stress and velocity in glaciers by theoretical analysis of stress fields and application of a flow law to obtain velocity fields. These treatments (primarily due to Nye, with substantial contributions by Weertman) have been applied to greatly simplified (generally two-dimensional) models, but nevertheless they yield relationships which reveal the significance of the various factors quantitatively. The derivation and solutions need not be given here, but the general approach is briefly outlined.[19,21]

At a general point in the ice mass, expressions for the components of stress and strain rate are set up, it being postulated that the ratios of strain rate components depend only on the ratios of deviator stress* components. The flow law is introduced by relating "effective shear stress" and "effective strain rate" which are defined in terms of deviator stress and strain rate components. Equations for plane strain are developed, and particular solutions for stress and velocity distributions in simple methods are derived by introducing appropriate boundary conditions. In addition to the simplifications inherent in the models, it is assumed that the flow law is known at every point (implying a complete knowledge of temperature and ice property distributions), that hydrostatic pressure has no effect on the flow law, and that the material is isotropic and incompressible.

The basis of Nye's theory of flow by ice deformation[19,21] (as distinct from sliding motion) can be outlined concisely if tensor notation is used:

At a general point in the ice, the equations of slow, non-accelerating flow are

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \rho g_i = 0 \tag{4}$$

and the strain rate components are, from standard theory:

$$\dot{\epsilon}_{ij} = \frac{1}{2}\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i}\right). \tag{5}$$

Strain rate is assumed to depend only on deviator stress, which has components

$$\sigma'_{ij} = \sigma_{ij} - \frac{1}{3}\,\delta_{ij}\,\sigma_{kk} \tag{6}$$

and, in the isotropic incompressible material, it is postulated that ratios of strain rate components depend only on the ratios of deviator stress components, so that

$$\dot{\epsilon}_{ij} = \lambda\,\sigma'_{ij}. \tag{7}$$

Effective shear stress and effective strain rate are defined by Nye in terms of second invariants, as

---

*Deviator stress, $\sigma'_1 = \sigma_1 - \frac{1}{3}(\sigma_1 + \sigma_2 + \sigma_3)$

Effective shear stress, $\tau = \sqrt{\frac{1}{2}(\sigma'_1\sigma'_2 + \sigma'_2\sigma'_3 + \sigma'_3\sigma'_1)}$

Effective strain rate, $\dot{\epsilon} = \sqrt{\frac{1}{2}(\dot{\epsilon}_1\dot{\epsilon}_2 + \dot{\epsilon}_2\dot{\epsilon}_3 + \dot{\epsilon}_3\dot{\epsilon}_1)}$ where σ_1, σ_2, σ_3 are principal stress components, ϵ_1, ϵ_2, and ϵ_3 are principal strain components, and the prime indicates a deviator.

$$\tau = \sqrt{\tfrac{1}{2}\,\sigma'_{ij}\,\sigma'_{ij}} \tag{8}$$

$$\dot{\epsilon} = \sqrt{\tfrac{1}{2}\,\dot{\epsilon}_{ij}\,\dot{\epsilon}_{ij}} \tag{9}$$

and they are related by a general flow law

$$\dot{\epsilon} = f(\tau). \tag{10}$$

In the above,

i, j = 1, 2, 3 and the summation convention for repeated suffixes is used.

x_j = coordinate system directions.

σ_{ij} = stress components.

$\dot{\epsilon}_{ij}$ = strain rate components.

u_i = velocity components in directions x_i.

σ'_{ij} = deviator stress components.

δ_{ij} = 1 if $i = j$, = 0 if $i \neq j$.

λ = a scalar factor.

ρ = ice density.

g_i = components of gravitational acceleration.

The dot convention indicates differentiation with time.

Particular solutions have been derived for a uniformly thick slab, of infinite lateral extent, resting on a rough inclined plane, for an ice shelf floating on the sea, and for two-dimensional models of idealized ice caps. It is necessary to make secondary assumptions when deriving particular solutions, and so the solutions mentioned above are not quoted here in view of the qualification required. The original references should be studied for details

Using simpler methods, Bader[1] derives a relationship for bed shear stress in the case of ice (not necessarily of uniform thickness) lying on an inclined plane.

Figure V-15 is a vertical section through a homogeneous ice mass in a plane parallel to the direction of flow. It is assumed that motion is parallel to the bed, and that at every point in the mass there is a hydrostatic stress corresponding to the weight of ice overburden. The vertical element in Figure V-15 is in motion at constant velocity under the action of the forces shown. Resolving in a direction parallel to the bed for the condition of no acceleration, and neglecting second order small quantities, gives

$$\tau = \rho g h \cos^2 \beta \tan \alpha \tag{11}$$

where τ = bed shear stress

ρ = ice density (assumed constant)

h = ice depth (measured vertically)

α = ice surface slope (usually small, so that $\tan \alpha \approx \alpha$)

β = bed slope (when small, $\cos \beta \approx 1$).

If the glacier bed is horizontal, the shear stress becomes

$$\tau = \rho g h \tan \alpha.$$

Some "wide" glaciers, e.g., the Greenland ice sheet, have surface profiles which are approximately parabolic. In such cases, the surface slope is proportional to the reciprocal of ice depth $\underline{h}$, and the bed shear stress is

$$\tau = a \rho g \cos^2 \beta$$

where $\underline{a}$ is a constant. Thus, if the bed slope is constant the shear stress is also constant.

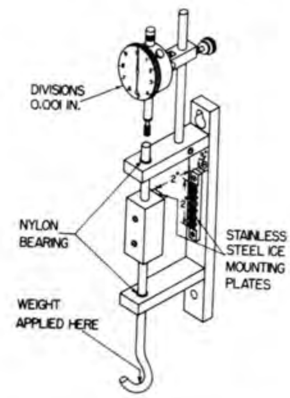

Figure V-14. Apparatus for making shear tests on ice specimens in the laboratory. (After Butkovich and Landauer, ref. 4)

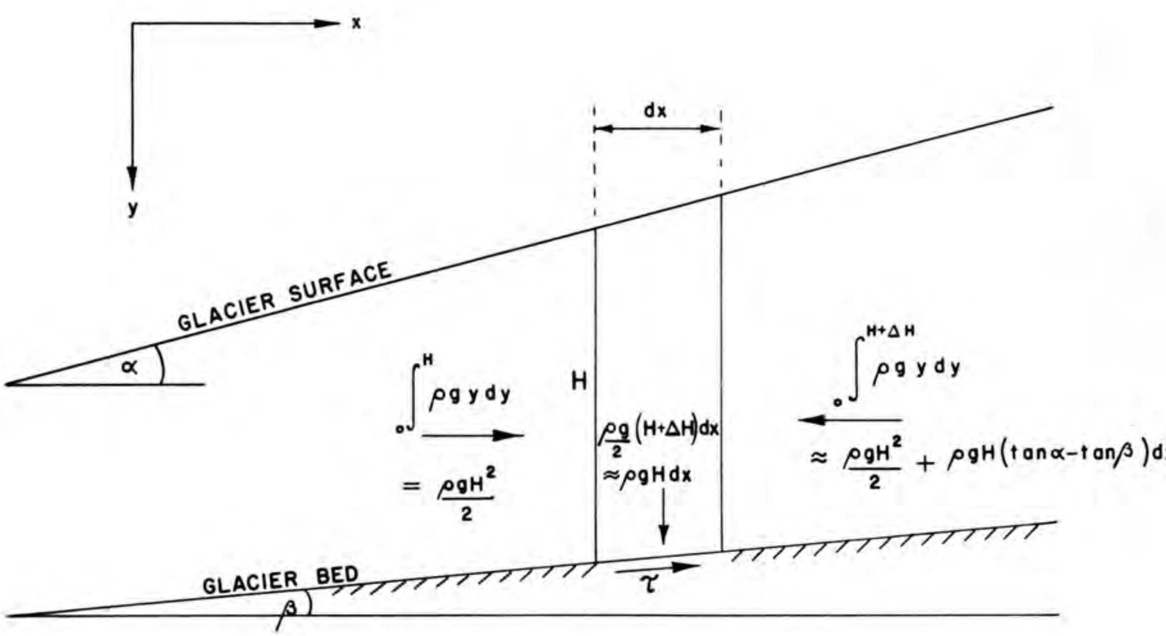

Figure V-15. Vertical section through a homogeneous ice mass in a plane parallel to the direction of flow. Forces producing motion — Bader's analysis. (After Bader, ref. 1)

The surface profile of East Antarctica appears to approximate an elliptic, rather than a parabolic, arc. In this case, therefore, differentiation of the shape equation to obtain slope, followed by substitution into eq 11, gives a linear relationship between bed stress and horizontal distance:

$$\tau = bx\rho g \cos^2 \beta$$

where b is a constant and x is the distance from the end of the glacier. If bed slope is constant there is a linear decrease of shear stress with distance from the glacier front.

The bed shear stress controls the rate at which a glacier slides over its bed, and in the case of a thick polar ice sheet, where shear deformation of the ice is thought to be concentrated in a very thin basal layer, it governs the total movement. Weertman[39] and Nye[22] have considered the theory of bed slip and its implications, working with an expression for sliding velocity of the form

$$v = C\tau^m \qquad (12)$$

where v = sliding velocity (constant with depth in the glacier)

τ = bed shear stress

m = a constant

C = a constant (probably dependent on bed roughness and ice temperature).

Weertman[39] explains the ability of ice to 'slide' over a bed studded with abrupt protuberances by invoking a combination of pressure melting and creep acceleration by stress concentration.

Flow and surface relief

Another application for bed stress equations is in the consideration of surface relief details on ice sheets. In Antarctica and Greenland, the apparently featureless expanses of the inland ice prove on closer acquaintance to have quite noticeable relief, which takes the form of rolling undulations or "terraces". Some of the more prominent "ice domes" are very obviously caused by ice disturbances over submerged hills, but where the ice is several thousand feet thick it is not clear whether minor surface perturbations can be caused by bed relief, particularly when the surface undulations are abrupt (a few kilometers or so between hilltops). However, it seems unlikely that they are caused by snowdrifting or other accumulation anomalies, even though these agencies may modify the undulations.

It can be reasoned that the degree to which bed topography influences ice surface relief depends upon the ratio p/h, where p is the height of a subglacial obstruction and h is the ice depth. As p/h approaches 1, the form of the ice surface will undoubtedly be affected (e.g., nunatak areas), but as p/h tends to infinity the surface ceases to be affected (a 10 ft high rock projection would not noticeably disturb a 10,000 ft thick ice sheet). Nye[22, 24] presents the expression

$$\frac{\Delta\alpha}{\alpha} \approx 1.4\frac{p}{h}$$

where α is surface slope, and $\Delta\alpha$ is the change of surface slope. Using more simple theory, in which h is inversely proportional to α if bed stress is assumed constant across the obstruction, Burgoin, Robin, and the writer have made quantitative correlations between surface and bed reliefs for parts of Greenland and Antarctica.

Kinematic waves

The final topic to be discussed before leaving flow theory is the propagation of kinematic waves, i.e., disturbances which travel down a glacier faster than the ice mass is moving.

In basic glacier theory, steady-state conditions are assumed; the rates of accumulation and ablation and the ice temperatures do not change with time, and therefore ice depth, velocity, and thus discharge, stay constant at any given cross-section. In nature, however, the glacier-controlling conditions do vary with time; there are seasonal changes, irregular year-to-year changes, and long-term variations due to climatic change. As a result, there are surges in the flow of a glacier, and "kinematic waves" of increased flow rate (but not necessarily visibly increased thickness) travel down it more rapidly than the ice itself is travelling. A similar phenomenon occurs in rivers: if a raft were launched in the headwaters at the same time as heavy runoff increased the flow there, the increased flow rate would be felt at a point far downstream before the raft had drifted down that far.

The theory of kinematic waves on glaciers was investigated in 1895 by de Marchi[6] and in 1907 by Finsterwalder,[8] but it was not until the phenomenon was rediscovered in 1958 by Nye[25,26] and Weertman[41] that glaciologists generally were made aware of it. Nye has since studied several facets of unsteady glacier flow by this theory, bringing to light many interesting features and possibilities.

The basic condition of continuity for flow of a stream of ice is

$$\frac{\partial Q}{\partial x} + \frac{\partial S}{\partial t} = aB \tag{14}$$

where $Q_{(x,\ t)}$ is the flow rate (volume/unit time) at a point distance $\underline{x}$ down the glacier and at time $\underline{t}$; $S_{(x,\ t)}$ is the corresponding cross-sectional area; $B_{(x,\ t)}$ the corresponding surface breadth; $a_{(x,\ t)}$ the average accumulation rate (ice thickness gained per unit time — negative value indicates ablation). For the steady state, when accumulation rate and glacier profile do not vary with time, eq 14 is simply

$$\frac{dQ_0}{dx} = a_0 \ B_0$$

the '0' suffixes denoting steady state conditions.

When the steady flow is disturbed by a change of accumulation, an expression relating the perturbations* (suffix '1') can be obtained from eq 14:

$$\frac{\partial Q_1}{\partial x} + \frac{\partial S_1}{\partial t} = (aB)_1 \tag{15}$$

since $Q = Q_0 + Q_1$, $S = S_0 + S_1$, $aB = a_0 B_0 + (aB)_1$.

It will be convenient to work with ice thickness $\underline{h}$ rather than cross-sectional area $\underline{S}$, so B_1 is neglected and $S_1 = B_0 h_1$. This is reasonable for confined flow, since a small height change will not affect the stream width greatly. Equation 15 then becomes

$$\frac{\partial Q_1}{\partial x} + B_0 \ \frac{\partial h_1}{\partial t} = B_0 h_1. \tag{16}$$

It is then assumed that $\underline{Q}$ at a point $\underline{x}$ is a function of ice thickness $\underline{h}$ and surface slope α. Since $\underline{Q}$ is the product of cross-sectional area $\underline{S}$ and average cross-section velocity $\underline{U}$, this implies that both $\underline{S}$ and $\underline{U}$ are functions of $\underline{h}$, and that $\underline{U}$ is also a function of α. $\underline{S}$ is a function of $\underline{h}$ if the stream width is constant ($S \propto h$), and also if the cross-section maintains geometrical similarity, e.g., if it is triangular ($S \propto h^2$). $\underline{U}$ is a function of $\underline{h}$ and α if the earlier discussion on stress and velocity is accepted, and if bed slope is constant; this follows from the conclusion that both sliding and shearing components of velocity depend on bed shear stress, which can be expressed in terms of $\underline{h}$ and α. If, then, $\underline{Q}$ is a function of $\underline{h}$ and α, the following relationship holds for small disturbances from the steady state:

$$Q_1 = \left(\frac{\partial Q}{\partial h}\right)_0 h_1 + \left(\frac{\partial Q}{\partial \alpha}\right)_0 \alpha_1. \tag{17}$$

Equations 16 and 17 are the basic equations of the kinematic wave theory.

*Perturbation — incremental change from the steady state.

Equation 17 can be rewritten,

$$Q_1 = c_0 h_1 - D_0 \left(\frac{\partial h_1}{\partial x}\right)$$

by putting

$$c_0 = \left(\frac{\partial Q}{\partial h}\right)_0, \quad D_0 = \left(\frac{\partial Q}{\partial \alpha}\right)_0, \quad \text{and} \; - \frac{\partial h_1}{\partial x} = \alpha_1$$

if the concentration is defined as 'quantity per unit distance'.

The velocity of a kinematic wave at point $\underline{x}$ is proportional to $\underline{c_0}$, and the diffusion coefficient at $\underline{x}$ is proportional to D_0. By making secondary assumptions, the basic equations 16 and 17 can be solved for particular cases, such as valley glaciers of given channel properties or ice sheets of the Greenland and Antarctic type. Suitable variations of accumulation rate $\underline{a}$ can be made to simulate annual fluctuations, climatic change, or other related effects. The solutions finally give kinematic wave velocities, rates of wave diffusion, response lag and stability of the glacier end.

Solutions obtained by Nye[28] indicate that kinematic waves produced by seasonal or secular changes in accumulation rate travel down the glacier at 2 to 5 times the average ice velocity $\underline{U}$. The characteristic response times of typical valley glaciers are calculated to be between 3 and 30 years, and the response time of the Antarctic ice sheet as a whole to be about 5,000 years. When long-period climatic changes occur, it is found that the upper regions of a valley glacier tend to respond in phase with the climate, thickness becoming greatest when accumulation rate is at a maximum. In the lower portions of a valley glacier, however, there are both direct responses and delayed responses caused by arrival of kinematic waves, and the net effect on the extent and thickness of the snout will depend on the interaction between the two. The Antarctic ice sheet shows an appreciable response lag for long-period changes, maximum thickness being reached about one-quarter period after accumulation rate reaches a maximum (a 250 year lag for a 1,000 year climatic cycle).

The stability of a disturbed glacier changes along its length; in sections where the velocity of a kinematic wave increases with distance, the glacier is stable, but where kinematic wave velocity declines with distance, the glacier is unstable, as the wave becomes a shockwave in principle. Since kinematic wave velocity is approximately proportional to average ice velocity, kinematic waves increase their speed with distance down glacier in regions of compressive flow. Thus, regions of compressive flow tend to be stable, while regions of extending flow are unstable.

Crevasses

Since the earliest days of glacier travel, crevasses have been a serious menace, and even with modern equipment they are still the major hazard for vehicles on exploratory ice sheet journeys. When concealed by snow bridges they are very difficult to detect, and the first indication of their presence is often break-through by a vehicle.

Crevasses are deep narrow fissures which form on glacier surfaces when tensile or shear strains are relieved by fracture. They are from a few inches to tens of feet wide (occasionally even wider) and may extend for many miles across the surface. The width decreases with depth and usually reduces to zero within 120 feet of the surface.

There are two common patterns in crevasse areas. The first is a series of parallel crevasses, often fairly equidistant, and the second consists of two intersecting systems of parallel crevasses. Crevasse patterns are governed by local flow characteristics, and highly complex systems can develop in mountainous areas.

In ablation areas, crevasses remain open and can be clearly seen, but in accumulation areas they frequently become bridged over and concealed by wind-blown snow, which builds out cornices until the gap is covered. In ablation areas where melting occurs, water drains into crevasses, re-freezing on the walls and sometimes completely filling the crevasse with clear ice.

Since crevasses are flow features, their formation can be studied by application of the stress field theory outlined above. Nye[19] has examined the case of transverse crevasse formation by longitudinal strains in a simple glacier model. The model, shown in Figure V-16 is a uniformly thick slab of ice resting on a rough slope, and the stress field in a vertical plane parallel to the flow direction is considered. The longitudinal stress in the direction of flow, σ_x, governs the formation of transverse crevasses, and for a general point in the plane (x, y) the solution of the basic equations gives

$$\sigma_x = \sigma_y \pm \sqrt{\tau^2 - \tau_{xy}^2} \tag{18}$$

where $\sigma_y = \rho g y \cos \alpha \approx$ hydrostatic pressure (head of ice + air pressure)

$\tau_{xy} = \rho g y \sin \alpha \approx$ (hydrostatic pressure) x(sin α)

τ = the "effective shear stress" (defined earlier), a measure of the octahedral shear stress.

ρ = mean ice density above the element.

Thus σ_x is composed of two terms, a compressive stress σ_y and another stress which can be either tensile or compressive depending on whether the square root is positive or negative. The positive root (tensile) corresponds to extending flow in a glacier, while the negative root represents compressive flow. These two conditions are analogous respectively to the Rankine active and passive states in soil mechanics.

If the variation of density with depth is neglected, σ_y and τ_{xy} increase linearly with depth. The variation of τ with depth is determined by the flow law and it is such that the magnitude of the square root term in σ_x becomes smaller relative to σ_y with increasing depth, i.e., as the depth increases $\sigma_x \rightarrow \sigma_y$ (hydrostatic). Hence, σ_x can be tensile only in the upper layers, where σ_y is relatively small, and so crevasses can form only near the surface. The maximum depth to which tensile stresses can occur is given by the condition, $\sigma_x = 0$, i.e.,

$$\sigma_y = 2\sqrt{\tau^2 - \tau_{xy}^2}$$

or,

$$y = \frac{2\tau}{\rho g \sqrt{1 + 3\sin^2 \alpha}}. \tag{19}$$

In the simple model taken here, the maximum tensile stress which can be developed is at the surface, where

$$\sigma_x = 2\tau.$$

τ_0 is given by the flow law already referred to (eq 1), viz.

$$\tau = \left(\frac{\dot{\gamma}}{k}\right)^{\frac{1}{n}}.$$

Since crevasses in ice are fractures which occur when the ice is "pulled", it is interesting to inquire into the magnitudes of stresses and strain rates which produce them. Laboratory tests on small specimens show that ice subjected to sustained tensile loads up to 10 kg/cm² can flow without fracturing. Expressed in another way, strain rates up to 10^{-5} sec^{-1} can be sustained. In glaciers, however, observations show that crevasses frequently occur when strain rates approach 10^{-9} sec^{-1}. The stresses corresponding to strain rates of this order are around 1 kg/cm² at temperatures down to -20C, and this stress is much smaller than the "ultimate tensile strength" found by standard laboratory tests. It seems, therefore, that a large mass of glacier ice ruptures more readily than a small laboratory specimen. This can probably be attributed to imperfections and stress concentrations produced by thermal cracking, inclusions, and irregularities.

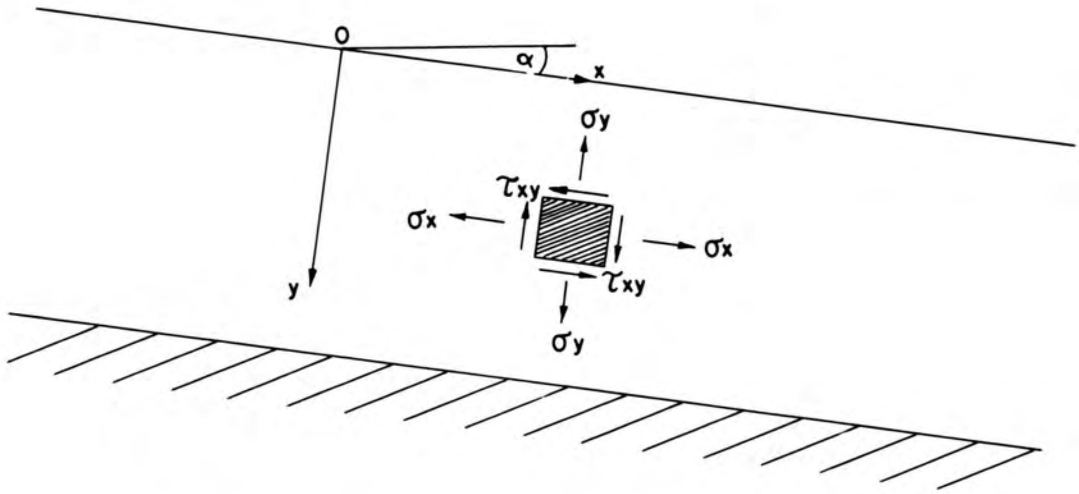

Figure V-16. Stresses on an element in a uniformly thick slab of ice resting on a rough slope — Nye's analysis of transverse crevasse formation. (After Nye, ref. 19)

For a marginal zone in N.W. Greenland and for the Saskatchewan Glacier, Canada, it has been suggested that crevasses occur when a critical tensile strain rate of 1% per year (3.2×10^{-10} sec^{-1}) is reached.[15, 17] A critical strain rate of this magnitude may be applicable to most glaciers, although crevasses might be expected to occur at smaller strain rates on very cold glaciers, since the stress/strain rate relationship is affected by temperature (see eq 3 above).

The formation of thermal cracks in glacier ice must also depend on the rate of straining. Slow cooling allows the ice to contract at a slow rate, and no cracking occurs. When the surface cools rapidly, however, rapid contraction develops high stresses and cracks appear. In periods of rapid cooling, with little wind, the noise produced by cracking can be quite alarming. It can also hinder seismic work when cracking sounds are picked up by the geophones.

When crevasses form, intercrevasse blocks are relieved of extending stress and the overall tensile strain is taken up by crevasse widening. As a result, the overall strain rate can be related to crevasse spacing and opening rate at the surface by the expression[17]

$$w = \dot{\epsilon}_x s \tag{20}$$

where $\dot{w}$ = crevasse widening rate

$\dot{\epsilon}_x$ = overall strain rate in the direction of flow

s = distance between consecutive crevasses at $y = 0$.

Observations show that there is a linear decrease of opening rate with depth.

Expressions for the maximum depth of crevasses have been derived, using a condition similar to eq 19. These expressions may require slight modification, since crevasse depth will depend to some extent on the relative magnitudes of widening rate and wall deformation under residual hydrostatic pressure in intercrevasse blocks.

The complexity and diversity of natural conditions produces an almost infinite variety of crevasse combinations, many of which defy theoretical analysis, but nevertheless an acquaintance with the theoretical mechanics for a simple model is an aid to the understanding of complicated local systems.

BIBLIOGRAPHY

1. Bader, H. Personal communications — unpublished note on glacier flow.
2. Burgoin, J. P. (1956) Quelques caractères analytiques de la surface et du socle de l'inlandsis Groenlandais (Some analytical characteristics of the surface and base of the Greenland ice sheet), Annales de Géophysique, Fom. 12, no. 1.
3. Bodvardsson, G. (1955) On the flow of ice sheets and glaciers, Jökull, vol. 5.
4. Butkovich, T. R. and Landauer, J. K. (1959) The flow law for ice, U. S. Army Snow, Ice and Permafrost Research Establishment Research Report 56.
5. ______ (1960) Creep of ice at low stresses, USA SIPRE Research Report 72.
6. de Marchi, L. (1895) Le variazioni periodiche dei ghiacciai (Periodic glacier variations), Reale Instituto Lombardo di Scienze e Lettere, Rendiconti, Series 2, vol 28.
7. Finsterwalder, R. (1958) "Brief notes about development of the physical aspects of ice movement," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics, International Association of Scientific Hydrology, Pub. no. 47.
8. Finsterwalder, S. (1907) Die Theorie der Gletscherschwankungen (The theory of glacier oscillations), Zeitschrift fur Gletscherkunde, vol. 2, p. 81.
9. Gerrard, J. A. F.; Perutz, M. F., and Roch, A. (1952) Measurement of the velocity distribution along a vertical line through a glacier. Proceedings of the Royal Society (London), Series A, vol. 213.

BIBLIOGRAPHY (Cont'd)

10. Glen, J. W. (1955) The creep of polycrystalline ice, Proceedings of the Royal Society (London), Series A, vol. 228.

11. ________ (1958) The mechanical properties of ice — I. The plastic properties of ice, Advances in Physics, vol. 7, no. 26.

12. ________ (1958) "The flow law of ice," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics.

13. Gold, L. W. (1962) "Deformation mechanisms in ice," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

14. Mathews, W. H. (1959) Vertical distribution of velocity in Salmon Glacier, British Columbia, Journal of Glaciology, vol. 3, no. 26.

15. Meier, M. F. (1957) Mode of flow of Saskatchewan Glacier, Alberta, Canada, California Institute of Technology, Ph.D. thesis.

16. ________ (1958) "Vertical profiles of velocity and the flow law of glacier ice," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics.

17. ________, et al. (1957) Preliminary study of crevasse formation, USA SIPRE Report 38.

18. Mellor, M. (1959) Analysis of ANARE seismic sounding records for MacRobertson Land (Unpublished).

19. Nye, J. F. (1951) The flow of glaciers and ice sheets as a problem in plasticity, Proceedings of the Royal Society (London), Series A, vol. 207, no. 1091.

20. ________ (1953) The flow law of ice from measurements in glacier tunnels, laboratory experiments and the Jungfraufirn borehole experiment, Proceedings of the Royal Society, Series A, vol. 219, no. 1139.

21. ________ (1957) The distribution of stress and velocity in glaciers and ice sheets, Proceedings of the Royal Society, Series A, vol. 239, no. 1216.

22. ________ (1959) The motion of ice sheets and glaciers, Journal of Glaciology, vol. 3, no. 26.

23. ________ (1952) A method of calculating the thickness of the ice sheets, Nature, vol. 169, no. 4300.

24. ________ (1959) Surface topography of the Antarctic ice sheet, Nature, vol. 184, no. 4689.

25. ________ (1958) Surges in glaciers, Nature, vol. 181, no. 7.

26. ________ (1958) "A theory of wave formation in glaciers," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics.

27. ________ (1962) "Theory of glacier variations," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

28. ________ (1960) The response of glaciers and ice-sheets to seasonal and climatic changes, Proceedings of the Royal Society, Series A, vol. 256.

29. Rigsby, G. P. (1960) Crystal orientation in glacier and in experimentally deformed ice, Journal of Glaciology, vol. 3, no. 27.

30. Robin, G. de Q. (1958) Seismic shooting and related investigations, Scientific Results of the Norwegian-British-Swedish Antarctic Expedition, vol. 5, Oslo.

31. Sharp, R. P. (1953) Deformation of a vertical borehole in a piedmont glacier, Journal of Glaciology, vol. 2, no. 13.

32. ___________ (1960) Glaciers. University of Oregon Press.

33. Shumskii, P. A. (1958) "The mechanism of ice straining and its recrystallization," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

34. Souslov, V. F. and Nozdryukhin, V. K. (1960) The movement of the Fedchenko Glacier in the firn zone, Moscow: Akad. Nauk SSSR, Glaciological Researches, no. 5 (text in Russian).

35. Steinemann, S. (1958) "Résultats expérimentaux sur la dynamique de la glace et leurs corrélations avec le movement et la pétrographie des glaciers," (Results of experiments on the dynamics of ice and their correlation with movement and petrography of glaciers), in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

36. Vialov, S. S. (1958) "Regularities of glacial shields movement and the theory of plastic viscous flow," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

37. ___________ (1958) "Regularities of ice deformation," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

38. Weertman, J. (1957) Steady-state creep of crystals, and Steady-state creep through dislocation climb, Journal of Applied Physics, vol. 28.

39. ___________ (1957) On the sliding of glaciers, Journal of Glaciology, vol. 3, no. 33

40. ___________ (1957) Deformation of floating ice shelves, Journal of Glaciology, vol. 3, no. 38.

41. ___________ (1958) "Traveling waves on glaciers," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

42. ___________ (1961) Equilibrium profiles of ice caps, U. S. Army Cold Regions Research and Engineering Laboratory Research Report 84.

43. ___________ (1962) Stability of ice-age ice caps, USA CRREL Research Report 97.

44. ___________ (1962) "The Eshelby-Schoeck viscous dislocation damping mechanism applied to the steady-state creep of ice," in Ice and snow — processes, properties and applications, (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

45. Ahlman, H. W. (1948) Glaciological research on the North Atlantic coasts, Royal Geographical Society (London), Research Series no. 1.

46. Wilson, S. D. (1959) Report on the 1958 Tuto ramp slope indicator measurements, Shannon and Wilson, for U. S. Army Corps of Engineers.

CHAPTER VI. WASTAGE OF GLACIER ICE.

In the previous chapters accumulation, metamorphism, and movement of glacier ice were discussed; we now turn to a consideration of how ice is lost from glaciers.

Ice can be removed from a glacier in the solid, liquid, and vapor phases by the following processes: iceberg calving and blowing snow (solid), surface and subglacial melting (liquid), and evaporation, either directly from ice or from melt water (vapor).

The relative effectiveness of each process will vary from glacier to glacier, and from place to place on a single glacier. Where a swiftly moving glacier debouches into the sea, iceberg calving may remove far more material than any other process, but at the snout of a glacier which terminates on land in the temperate regions, surface melting will almost certainly predominate. At high altitudes on polar glaciers, the only local loss processes are evaporation and deflation (wind erosion of the snow).

Iceberg calving

Glacier ice flowing down to the sea reaches the coast in a variety of forms, and for convenience we distinguish here three different types of flowing ice which produce icebergs. They are:

(a) Ice streams — distinct streams of relatively fast-moving ice, with or without lateral boundaries of visible rock.

(b) Ice shelves — extensive areas of floating ice fed by flow from the land and by local snow accumulation.

(c) Ice sheet edge — the undifferentiated edge of a land-based ice sheet, which flows down to the sea over a very wide frontage at more or less uniform low speed.

Ice streams produce almost all the icebergs of the northern hemisphere, and make a large contribution to the berg production of Antarctica. The kind of icebergs formed range from relatively small fragments, which break off from different parts of the snout at frequent intervals, to bergs many square miles in area, produced when a large part of the floating glacier tongue parts from the parent mass. Local conditions govern the details of the calving process. The valley glaciers of Alaska and Spitsbergen, which flow into constricted fjords, generally crumble into small pieces (which should probably be called "bergy bits" rather than bergs) at their termini.* Some of the big Antarctic outflow glaciers, on the other hand, push long floating tongues out into the ocean, and when these finally split off under tidal action they become large bergs.

Ice shelves are almost exclusively Antarctic features. The major ones, e.g., the Ross, Filchner, Amery, Shackleton, are vigorously active, the twofold nourishment (inflow and accumulation) leading to quite rapid horizontal motion towards the ice front (the unrestrained seaward edge). The wide frontage of free-floating ice permits enormous flat-topped bergs, commonly hundreds of square miles in area, to break loose. Ice shelves are the principal source of Antarctic bergs.

Along much of the Antarctic coastline the inland ice meets the sea directly, without being channeled into ice streams and without intervening ice shelves. The rate of flow of the ice along these stretches is usually very slow but, because of the wide frontages involved, the total contribution to the overall berg production is significant. At any one place, though, major calving is infrequent.

To estimate the rate of discharge of a glacier, the area of a vertical cross-section near the snout is multiplied by the mean velocity at that section:

$$m = \gamma AV$$

where m is the rate of discharge (mass per unit time)
γ is the mean ice density
A is the area of the cross-section
V is the mean velocity of the ice.

The mean ice density γ may vary from about 0.84 g/cm^3 for shelf ice to about 0.91 for relatively bubble-free "warm" ice.

* Similar attrition of the terminal ice cliffs occurs on most coastal glaciers, but is negligible compared to calving of giant bergs.

The cross-sectional area of a confined valley glacier can be calculated if depth soundings are available, otherwise it is obtained from an inspired guess. When a very wide glacier front is being studied, the average depth is multiplied by the surface width. If the ice is free-floating (ice shelf or floating glacier tongue) the depth may be estimated from buoyancy considerations. When this procedure is followed, it should be remembered that oceanic melting and lateral spreading have thinned the ice at the seaward extremity.

The average flow velocity $\underline{V}$ is usually estimated from measured surface velocity. In thick ice (> 500 m deep) it is usually permissible to assume the surface velocity as an average for the entire depth. In thin ice, however, some reduction of the surface velocity should be made to compensate for internal shear (see Chapter V).

Wind-blown snow

Considerable quantities of snow can be carried off glaciers by the wind, and in some areas wind transport is very significant. The effect is often strongly marked where katabatic winds flow down steep slopes or funnel through valleys.

Some typical mass transport rates for the Antarctic coast, and height distributions for different wind conditions, are shown in Figure VI-1. The total mass transport of snow through a vertical strip, of unit width, perpendicular to the wind direction is given by the product of drift concentration and wind velocity integrated through the height of the strip, from ground level up to the top of the drift cloud:*

$$Q = \int_0^H v \cdot n \cdot \rho \, dz$$

where $\underline{Q}$ is snow transport (g/cm-sec), $\underline{v}$ is wind velocity at height $\underline{z}$ (cm/sec), $\underline{n}$ is drift concentration at height $\underline{z}$ (non-dimensional), ρ is air density (g/cm^3) and $\underline{H}$ is the height of the top of the drift cloud (cm).

Total transport is difficult to estimate on the basis of shallow surface observations in view of the uncertainty regarding vertical extent of the drift cloud and distribution of wind velocity with height. From studies made in Antarctica, however, it appears that strong blizzard winds (say about 60 mph or 25 m/sec) can carry snow at rates up to 100 g/cm-sec. Excessively high winds, such as those of Terre Adélie, can move snow much more rapidly than this. Figure VI-2 gives rates of snow transport as a function of wind speed for Wilkes Station, a moderately windy place on the coast of East Antarctica.

Surface melting and evaporation (ablation)

Melting and evaporation at the surface is usually termed ablation, although there are some who use the word to mean removal of ice by any process.

Melting is the principal wastage process for all glaciers which terminate on land; melt water is channeled into definite drainage courses and runs off directly, either by flow along the surface or by passage through subglacial channels. Where surface slopes are gentle, particularly on high snowfields, melt water may be unable to escape by direct runoff and in such cases refreezing occurs, either in the snow pack or in surface depressions. Evaporation from a glacier surface may be either evaporation from melt water or direct evaporation from the solid snow or ice. When melting, followed by runoff, is occurring evaporation is relatively unimportant; since evaporation can take place at any altitude and any time of year, however, it may be a significant loss process.

Melting and evaporation are dependent upon, and in turn influence, the heat balance at the glacier surface. The main items entering into this heat balance are: (a) transfer by convection and conduction, (b) the net short-wave radiation, (c) the net long-wave radiation, (d) the intake or output of latent heat as melting, freezing, evaporation, or condensation takes place. Rainfall may also be an important heat source, even though amounting to only a few inches a year. Figure VI-5 gives the heat balance items and the factors controlling them. An understanding of these things explains the variations of ablation which occur seasonally and diurnally, and the variations with latitude, altitude, and general location.

*See also Chapter III

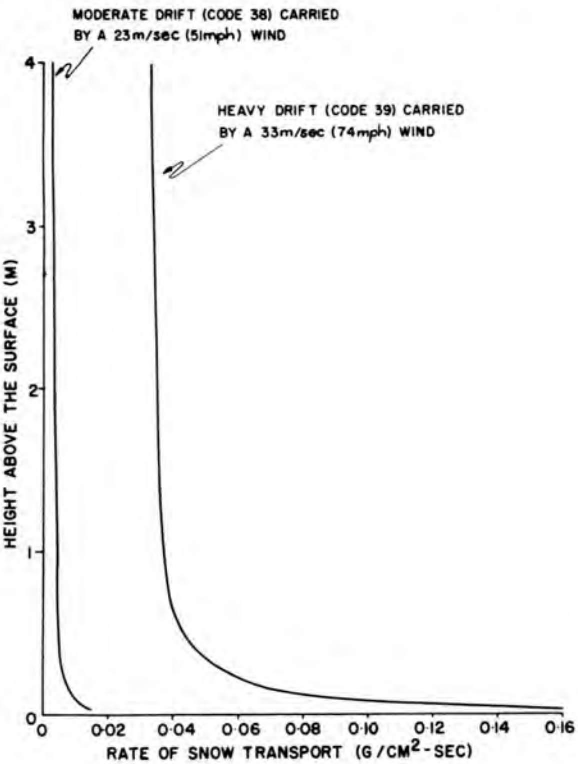

Figure VI-1. Rate of snow transport as a function of height above the surface. (ref. 19)

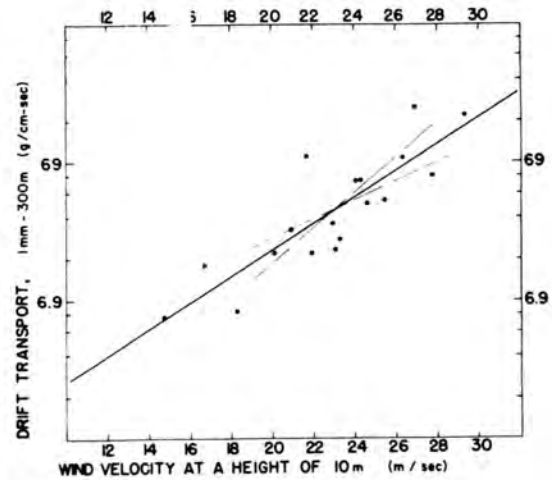

Figure VI-2. Rate of snow transport related to the wind velocity at a height of 10 m. (After Dingle and Radok, ref. 4)

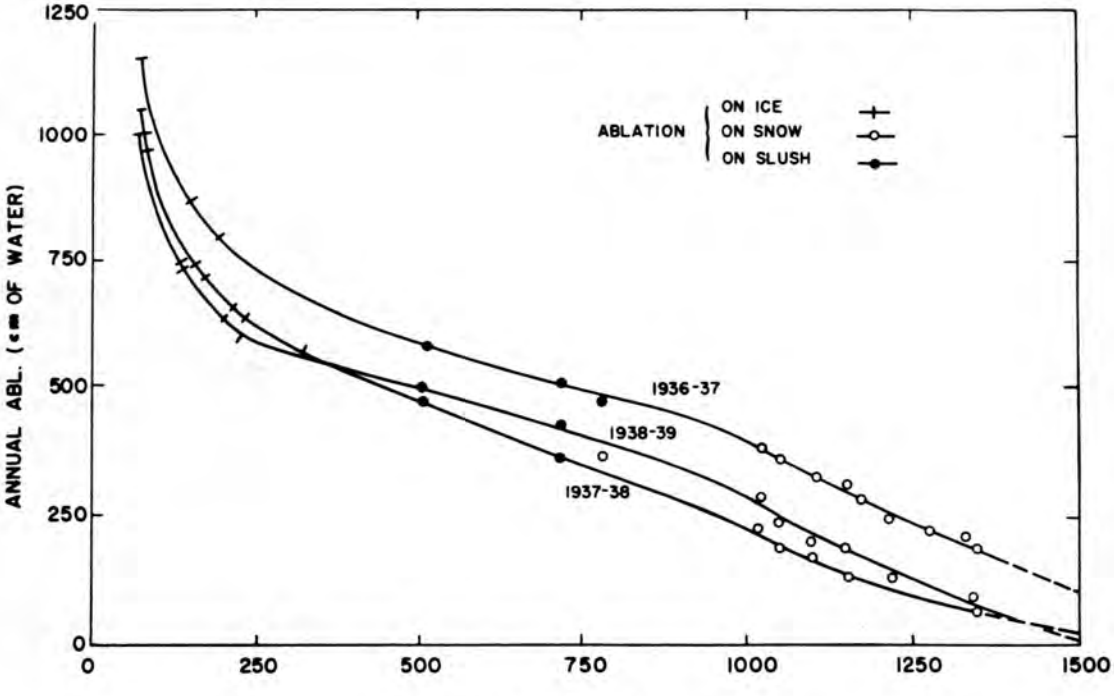

Figure VI-3. Annual ablation on the Hoffell Glacier, Iceland, as a function of surface altitude. (After Ahlmann, ref. 1)

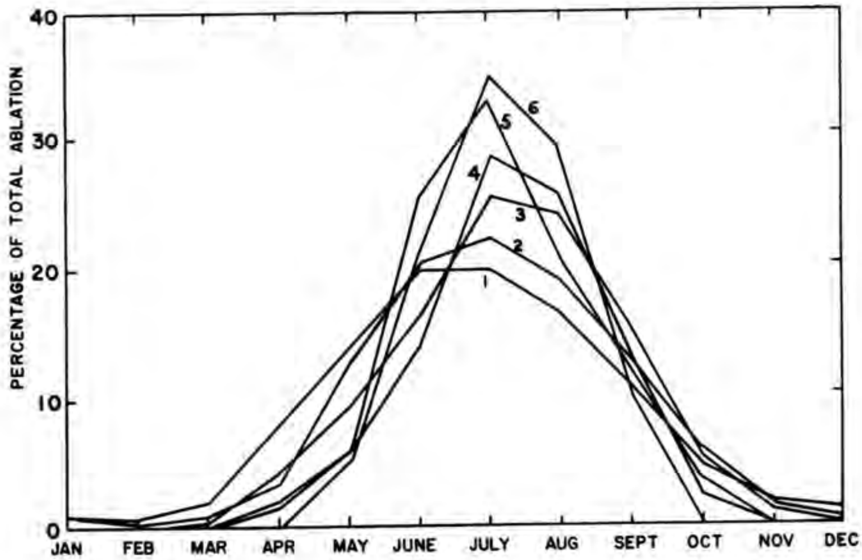

Figure VI-4. Mean monthly ablation values expressed as percentages of the mean annual ablation for the Hoffell and Heinaberg Glaciers, Iceland. The curves refer to the following surface altitudes: 1 - 72 m, 2 - 185 m, 3 - 715 m (all on Hoffell Glacier); 4 - 1025 m, 5 - 1115 m, 6 - 1345 m (4 - 6 on Heinaberg Glacier). As the altitude of the site goes up, so does the percentage of the total ablation occurring at the period of maximum solar radiation (June-July). (After Ahlmann, ref. 1)

The sources of heat for ablation can be put under three headings: convection, condensation and radiation. The relative importance of each will vary with time and from place to place, but their overall contributions to the ablation process can be compared for different glaciers, as is done in Table I. Although not immediately evident from the table, radiation usually becomes relatively more important with increasing altitude and with increasing distance from the sea ("cold and sunny" weather). Convective heating becomes more important in areas where high air temperatures are experienced.

An important factor governing the balance of short-wave radiation is the albedo of the surface, which expresses the ratio of reflected radiation to incident radiation. Albedo is actually the reflectance of the snow or ice, which is a function of radiation wavelength, integrated between the somewhat arbitrary limits which define the "short-wave" band (about $0.3\,\mu$ to $4\,\mu$ for common measuring instruments). Only smooth fresh snow acts as a diffuse reflector; old and crusted snow, and ice, give some specular reflection, and surfaces commonly have directional irregularities such as sastrugi, dunes, and radiation pits. Albedo therefore varies with the elevation and azimuth of the sun when the sky is clear. The most significant factors which control albedo, however, are those which affect the condition or color of the surface — grain form, free water content, solid impurities. Clean, non-melting snow on high polar snowfields commonly has maximum albedo values in the range 0.85 to 0.9, while clean but old snow has average values from 0.7 to 0.8. Albedos for clean, dry ice vary with the bubble content, with average values up to about 0.7. When wet and dirty, snow and ice may have values in the range 0.2 to 0.4.

The drastic acceleration of ablation with lowering of the albedo can easily be demonstrated by scattering dark-colored dust over the snow on a sunny day. The rate of ablation near a gravel road on the ice ramp at Tuto, Greenland was doubled by the effect of dust blown off the road.

Having discussed in general terms some of the factors affecting ablation, we can express the heat balance of an ice surface, which is the essence of the problem, in the form of an equation. It should be noted that the amount of ice melted or frozen corresponds directly to a quantity of heat. Similarly, evaporation and condensation represent

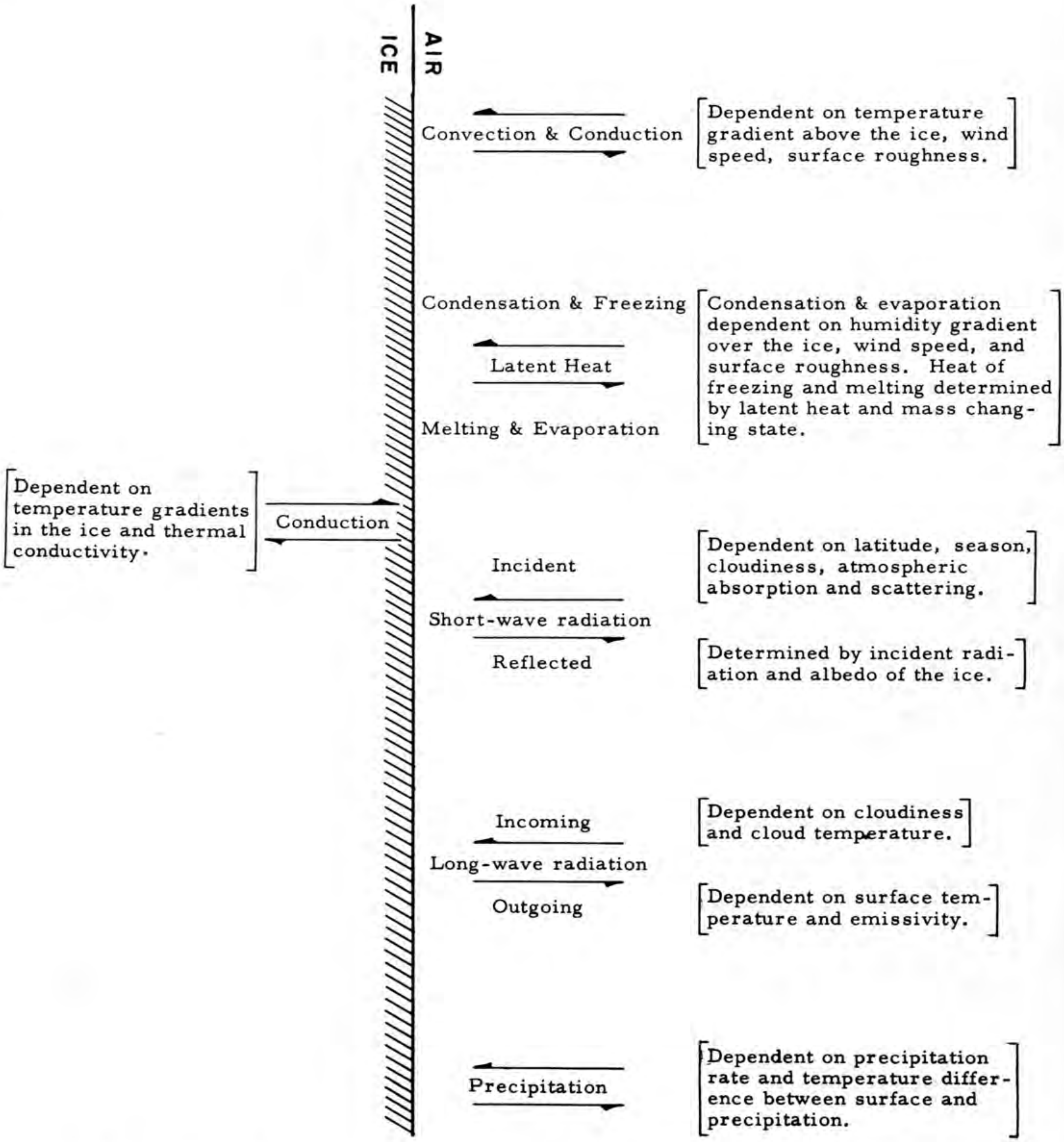

Figure VI-5. Factors controlling the heat balance of a horizontal ice surface.

heat loss or gain. The heat balance at an ice surface can be stated as follows:*

$$q_1 + q_2 + q_3 + q_4 + q_5 + q_6 + q_7 = 0$$

q_1 is the net heat flow into the ice in unit time due to short-wave radiation. It is the incoming short-wave solar radiation (direct and diffuse) minus the reflected short-wave radiation, which is given by multiplying the incoming radiation by $(1 - \beta)$, where β is the albedo of the surface. Short-wave radiation can be measured directly, or it can be estimated from standard charts when the date, latitude, altitude, cloud cover, and surface albedo are known.

q_2 is the net rate of heat flow into the ice due to long-wave radiation. It is the incoming long-wave radiation, which depends mainly on cloud amount and cloud temperature, minus the outgoing long-wave radiation, which depends on the temperature of the ice surface. Since snow and ice behave as a "black body" for long-wave radiation, the outgoing can be expressed as σT^4, where σ is the Stefan-Boltzmann constant (0.826×10^{-4} ly min^{-1} deg^{-4}) and $\underline{T}$ is absolute temperature of the surface. The net heat flow due to long-wave radiation is usually negative (net outflow), although it can be positive with a cover of clouds which are warmer than the ice.

q_3 is the net rate of flow of sensible heat due to convection. Heat is carried into, or away from, the surface as a result of vertical exchange of air caused by turbulence. The heat flow depends on the density and specific heat of the air, on the temperature gradient above the ice, and on the coefficient of eddy conductivity, which itself varies with wind speed and surface roughness.

q_4 is the rate of flow of latent heat connected with evaporation (negative sign) or condensation (positive sign). The vertical exchange of water vapor over the ice is analogous to the convective transfer of sensible heat. It depends on air density, the humidity (vapor pressure) gradient, and the coefficient of eddy diffusion for water vapor, which is often assumed to equal the coefficient of eddy conductivity mentioned above. The rate of mass transfer by evaporation multiplied by the latent heat of vaporization gives the evaporative heat flow.

q_5 is the rate of heat flow to or from the surface by conduction through the ice. It depends upon the temperature distribution within the ice and on the thermal conducticity of the ice. During winter, heat flows from the ice to the surface, while in spring and summer heat flows from the surface into the ice, conduction ceasing when the subsurface layers become isothermal at the melting point.

q_6 is the rate of supply of sensible heat by precipitation. It is given by the product of specific heat, precipitation rate (in mass units), and temperature difference between the precipitation and the ice surface.

q_7 is heat flow corresponding to melting (negative sign) or freezing (positive sign) at the surface. The flow of available heat can be equated to the mass of ice melted multiplied by the latent heat of fusion. In an ablation study the heat available for melting can be regarded as the surplus of the surface heat budget.

Subglacial melting

Subglacial melting can occur at the ice-rock interface beneath land glaciers, or at the ice-sea interface below floating ice shelves. Because these zones are virtually inaccessible, very little observational evidence on melting rates is available.

The amount of subglacial melting beneath land glaciers can be calculated with some confidence when the vertical distribution of temperature in the glacier is known, since the flow of heat from the earth's interior is fairly constant. In regions where the geothermal flux is about average, i.e., 1.65×10^{-6} cal/cm^2-sec for continents, subglacial melting must be slight, since the heat available is only enough to melt about 7 mm of ice per year.† In volcanic areas, however, subglacial melting may be locally intense.

The water flowing out from beneath the snouts of mountain glaciers should not be equated with subglacial melting; much of this is surface melt water which has found its way down cracks and sink holes to the base of the glacier.

* Molecular conduction in the air is neglected here

† Although small in itself, this basal melt water may lead indirectly to high ice losses, as it will greatly facilitate sliding of the glacier on its bed.

Table VI-I. Relative contributions of convection, condensation, and radiation to ablation. (After Ahlmann, ref. 1; Orvig, ref. 14; Lister & Taylor, ref. 11)

Glacier	Year	Position	Elev. (meters)	Dist. from ocean (km)	Conv. (%)	Condens. (%)	Rad. (%)	Surface
Sveanor snowfield, N. E. Land	1931	79°56'N 18°18'E	5	6	58	18	24	snow
Isachsens Plateau, W. Spitsbergen	1934	79°09'N 12°56'E	870	30	29	15	56	snow
Fourteenth of July Glacier, W. Spitsbergen	1934	79°08'N 12°E	600	8	53*	-	47	snow (ice later)
Britannia Gletscher, N. E. Greenland	1953	77°14'N 23°49'W	620	150	32	1	67	ice (snow earlier)
Britannia Gletscher, N. E. Greenland	1953	77°12'N 23°48'W	470	150	20	5	75	ice
Frejagletscher, N. E. Greenland	1939	74°24'N 20°50'W	453	10	83	9	8	snow
Barnes Ice Cap, Baffin Island	1950	70°N 72°W	865	150	32*	-	68	snow (ice later)
Penny Ice Cap, Baffin Island	1953	66°59'N 65°28'W	2050	90	9	30	61	snow
Karsa Glacier, Sweden	1942-8	68°20'N 18°20'E	>1100	80	44	24	32	snow
Karsa Glacier, Sweden	1942-8	68°20'N 18°20'E	<1100	80	29	16	55	ice
Hoffellsjökull, Iceland	1936	64°30'N 15°30'W	1000	20	92*	-	8	ice
Salmon Glacier, British Columbia	1957	56°10'N 130°07'W	1700	320	15	10	75	snow
Vernagtferner, Austria	1950	46°50'N 10°45'E	3000	240	15	4	81	ice
Vernagtferner, Austria	1952	46°50'N 10°45'E	3000	240	16*	-	84	ice

*Includes condensation

The intensity of melting at the base of ice shelves is not so easy to predict, and the question of this basal melting is one of the outstanding unsolved problems of Antarctic glaciology. In the absence of direct observations, inferences have been drawn from the forms of vertical temperature profiles and from considerations of mass flux, but both data and interpretive theory are inadequate for definite conclusions to be arrived at. It is generally agreed that melting does occur, and that melting rates are significant near the ice front. Actual rates, and their relation to distance from the open ocean, are unknown.

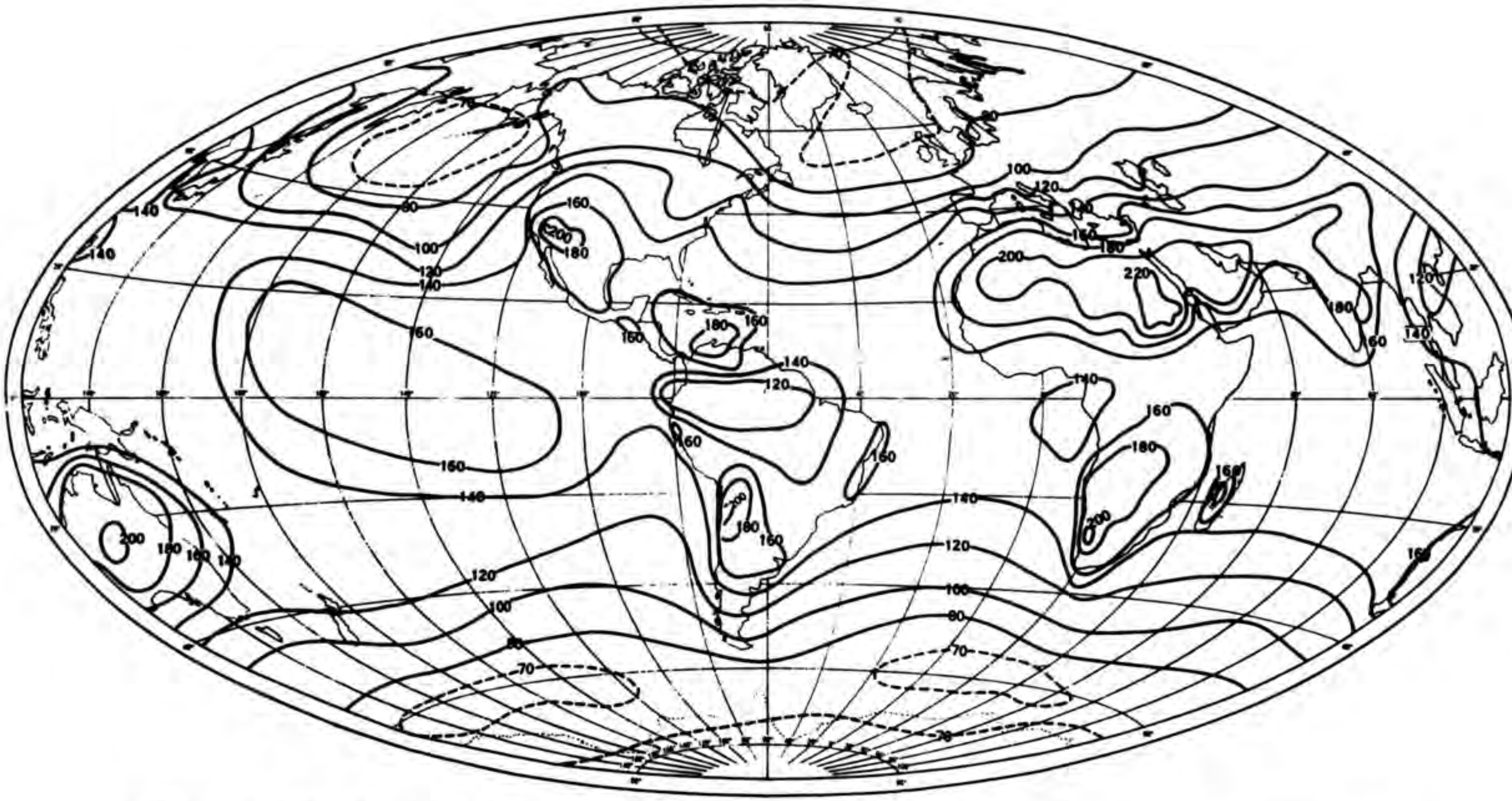

Figure VI-6. Annual global radiation received at the earth's surface — generalized distribution. Numbers on the isolines give kilo-langleys per year. (After Landsberg, ref. 18)

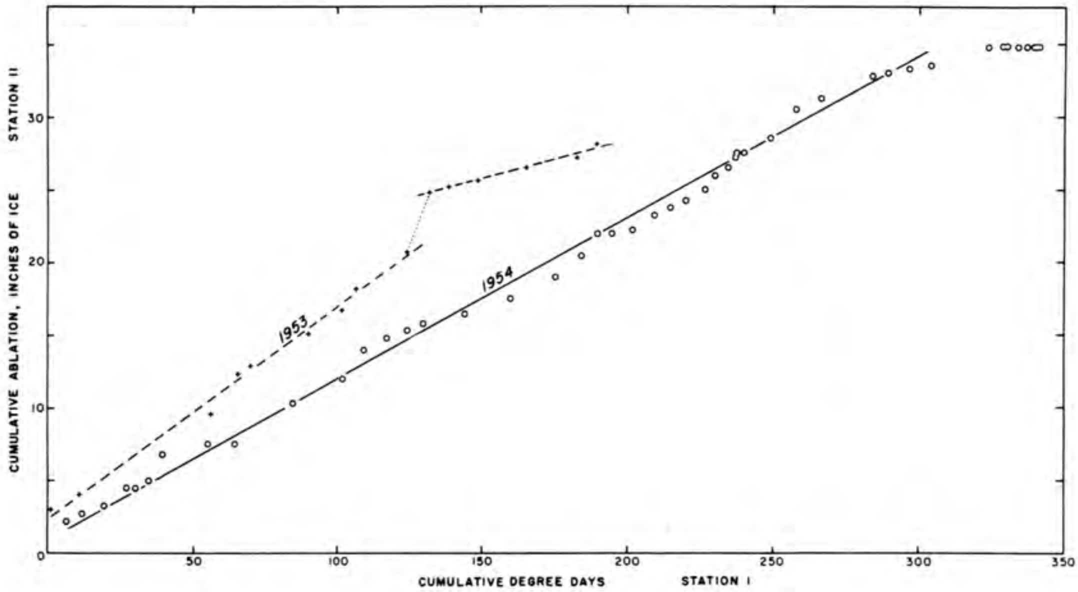

Figure VI-7. Empirical correlation between ablation and air temperature. Ablation quantity is plotted against a summation of above-freezing air temperatures. The sum of above-freezing temperatures for the whole ablation season is sometimes used to define a "thawing index". This approach to ablation studies may be expected to yield reasonable results at sites where convective heating is important, but it is not suitable for sites where radiation predominates. (After Nobles, ref. 13)

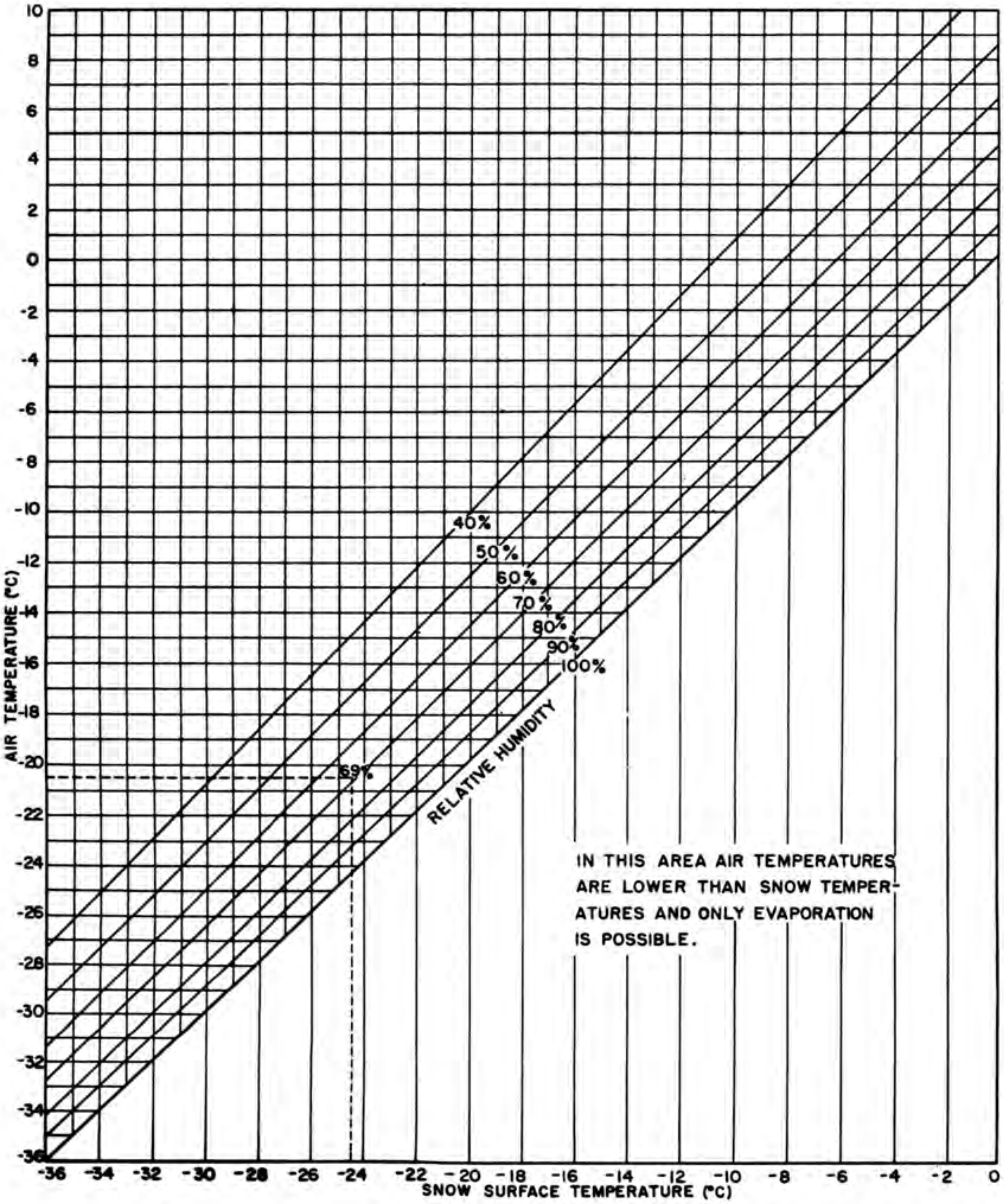

Figure VI-8. Conditions for evaporation or condensation from a snow surface. If measured relative humidity of air is greater than that shown on chart for given air and snow temperatures, condensation will occur; if lower, evaporation will occur. (After Diamond, ref. 3)

BIBLIOGRAPHY

1. Ahlmann, H. W. (1948) Glaciological research on the North Atlantic coasts, Royal Geographical Society (London), Research Series, no. 1.

2. Charlesworth, J. K. (1957) The Quaternary era. London: Arnold, vol. 1.

3. Diamond, M. (1953) Evaporation or melt of a snow cover, U. S. Army Snow, Ice and Permafrost Research Establishment (USA SIPRE), Research Report 6.

4. Dingle, W. R. J., and Radok, U. (1961) "Antarctic snow drift and mass transport," in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55

5. Gerdel, R. W., Diamond, M., and Walsh, K. J. (1954) Nomographs for computation of radiation heat supply, USA SIPRE, Research Paper 8.

6. Griffiths, T. M. (1960) Glaciological investigations in the TUTO area of Greenland, USA SIPRE, Technical Report 47.

7. Hoinkes, H. (1955) Measurements of ablation and heat balance on Alpine glaciers, Journal of Glaciology, vol. 2, no. 17.

8. Hubley, R. C. (1957) An analysis of surface energy during the ablation season on Lemon Creek Glacier, Alaska, Transactions, American Geophysical Union, vol. 38, no. 1.

9. ______ (1955) Measurements of diurnal variations in snow albedo on Lemon Creek Glacier, Alaska, Journal of Glaciology, vol. 2, no. 18.

10. Lister, H. (1958) "Glaciology: ablation or the heat balance" in Venture to the Arctic (R. A. Hamilton, Editor). Penguin Books.

11. ______ and Taylor, P. F. (1961) Heat balance and ablation on an Arctic glacier, Meddelelser om Grønland, bd. 158, no. 7.

12. Mellor, M. (1959) Mass balance studies in Antarctica, Journal of Glaciology, vol. 3, no. 26.

13. Nobles, L. H. (1960) Glaciological investigations on the Nunatarssuaq ice ramp, N. W. Greenland, USA SIPRE, Technical Report 66.

14. Orvig, S. (1954) Glacial-meteorological observations on ice caps in Baffin Island, Geografiska Annaler, vol. 36, pt. 3, 4.

15. Schytt, V. (1955) Glaciological investigations in the Thule Ramp area, USA SIPRE, Technical Report 28.

16. Sverdrup, H. U. (1935) Ablation of Isachsen's Plateau and on the Fourteenth of July Glacier in relation to radiation and meteorological conditions, Geografiska Annaler, vol. 17.

17. Wright, C. S. and Priestley, R. E. (1922) Glaciology, British (Terra Nova) Antarctic Expedition 1910-13, London.

18. Landsberg, H. E. (1961) Solar radiation at the earth's surface, Solar Energy, vol. V, no. 3.

19. Mellor, M. (1961) "The Antarctic ice sheet" in Cold Regions Science and Engineering (F. J. Sanger, Editor), U. S. Army Cold Regions Research and Engineering Laboratory (USA CRREL) monographs, Part I, Sect. BI.

20. ______ (1963) A brief review of the thermal properties and radiation characteristics of snow, Karlsruhe Conference of the German Society for Polar Research. (To be published in "Polarforschung".)

BIBLIOGRAPHY (Cont'd)

21. Scott, R. F. (1963) "Heat exchange at the ground surface" in Cold Regions Science and Engineering (F. J. Sanger, Editor), USA CRREL monographs, Part II, Sect. A1.

CHAPTER VII. MASS ECONOMY AND GLACIER FLUCTUATIONS

Surplus and deficit

Mass economy studies on glaciers aim at quantitative determination of the ice income (accumulation) and the ice expenditure (wastage), and hence the state of balance. A surplus of accumulation over wastage (a "positive balance") implies that the total ice volume is increasing and the glacier expanding. A deficit ("negative balance") indicates decreasing ice volume and shrinking of the glacier.

The various factors which influence total accumulation and total wastage have already been discussed; in estimating the mass balance, however, it is frequently only possible to measure net accumulation and net wastage. The balance between net accumulation and net wastage still gives the magnitude of the surplus or deficit, but additional data on gain and loss processes permit a more searching appraisal of the glacier's economy.

A simple case to consider is an independent mountain glacier which terminates on land, for here the mass balance is determined essentially by the relation of net accumulation to net ablation. If a network of stakes were planted on such a glacier so as to give a representative sampling of surface gains and losses, it would be found after one year that the stakes below the annual snowline show net lowering of the surface, and stakes above the annual snowline show net increase of snow cover. By suitable summation of losses and gains over the areas of the ablation and accumulation zones the mass balance would be arrived at.

In the example above, no information on flow of the glacier is required to arrive at a mass balance, but this becomes necessary when a glacier which calves into the sea is considered. If we look at an independent valley glacier which flows down into a fjord, accumulation and ablation enter the balance as before, but there is also ice loss at the terminus to be taken into account. This loss can be estimated by calculating the discharge of ice at the snout, i.e., the mean velocity multiplied by the mean ice thickness and frontal width.

In the case of a major ice sheet such as the Antarctic ice sheet, there are three net loss processes to be balanced against the net accumulation: iceberg production, surface melting and evaporation in zones of net ablation, and oceanic melting beneath ice shelves and floating glacier tongues. The sheer size of the Antarctic and Greenland ice sheets makes mass balance studies extremely difficult, since not even massive international efforts such as the IGY give a representative sampling of the economy of the whole ice mass.

Mass turnover

Balancing net gains against net losses shows whether the glacier is gaining or losing substance; the general activity of the economy is also of interest. To express the vigor of a glacier the term "mass turnover" is introduced here.* It is intended to describe the rate at which glacial processes operate, and an indicator of mass turnover might be the total accumulation (or total wastage) divided by the total glacier mass (mass = volume x average ice density).

A glacier with a low mass turnover is one which maintains a balance with small accumulation and small losses; the Antarctic ice sheet is a thrifty glacier of this kind, since its accumulation per unit of residual mass is very small. A glacier with a high mass turnover balances large accumulations against heavy losses; vigorously active glaciers of this kind often exist on coastal mountains, e.g., Vatnajökull in Iceland, or the west-coast glaciers of New Zealand.

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*Sharp[15] and Flint[7] use the term "metabolism" in a similar sense.

The mass budgets of Antarctica and Greenland

The prime targets for mass economy investigations are the ice sheets of Antarctica and Greenland, since there are far-reaching implications if these big glaciers have unbalanced budgets. Even after years of study, however, the available information is scanty and susceptible to diverse interpretations, so that few generally agreed conclusions have been reached. The best that can be done here is to outline mass budgets which give a reasonable impression of the magnitude of quantities involved, drawing on the more credible of published studies, and deliberately compromising between them.

Antarctica

The Antarctic mass economy has received considerable attention in recent years, but some of the estimates have to be discounted because of improper application of data, extrapolations not in accordance with geographical facts, or manipulation of the budget to fit foregone conclusions. The following figures, which are presented only for illustration, lie within the ranges of values established by modern investigators.[8,9,10,12,13,14,17,18]

Net gains

Accumulation $\approx 20 \times 10^{17}$ g/yr*

Net wastage

Iceberg production $\approx 12 \times 10^{17}$ g/yr
Oceanic melting $\approx 2 \times 10^{17}$ g/yr
Net ablation $\approx 1 \times 10^{17}$ g/yr

Other depletions

Blown snow $\approx 1 \times 10^{17}$ g/yr
Evaporation from accumulation zone $\approx 0.5 \times 10^{17}$ g/yr.

Comparison of net accumulation with net wastage shows an annual surplus of 5×10^{17} g; this is a huge surplus compared to the total budget (23%), but such a conclusion seems inescapable when analyzing available data (some published studies show larger surpluses — up to twice this amount).

The items listed under "other depletions" represent snow lost from the surface before net accumulation is measured, so that by adding them to the net accumulation a value is obtained for total (gross) accumulation, or precipitation. This value is nearly 22×10^{17} g/yr.

Greenland

The various mass economy estimates for Greenland lead to different conclusions regarding the state of balance. For the present purpose of illustration, a simplified budget is drawn up on the basis of the two most recent studies (by Benson[5] and by Bader[3]).

Net gains

Accumulation $\approx 5.5 \times 10^{17}$ g/yr

Net wastage

Net ablation $\approx 2.5 \times 10^{17}$ g/yr
Iceberg production $\approx 2.2 \times 10^{17}$ g/yr.

These figures indicate an appreciable surplus of 0.8×10^{17} g/yr, a value lying between the 3% surplus of Benson and the 19-43% surplus of Bader.

Glacier fluctuations

Fluctuations of the area and thickness of a glacier can often be followed or deduced from direct observations on the margins where ice meets rock. Old maps, sketches, or

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* 10^{17} g = 1.1×10^{11} short tons = 110 billion short tons = 100 km^3 of water.
This accumulation over a total area of about 13×10^6 km^2.

photographs can be compared with their modern equivalents; rock boundaries can be inspected for evidence of recession such as abandoned moraines, and scraped, vegetation-free surfaces.

These fluctuations can be linked with changes in the mass economy, but the relation between mass balance and glacier extent is, in general, not a simple and direct one. A very small mountain glacier might be expected to respond promptly to a changed regime: sudden increase in accumulation would cause expansion; sudden increase of ablation would cause retreat. The expansion or recession would be self-limiting, since expansion exposes more ice to wastage processes and recession withdraws ice from wastage zones. On a very long valley glacier, it is not to be expected that the snout would respond immediately to sudden changes of alimentation far up in the accumulation area, since the effects of these changes have somehow to be transmitted to the snout. This raises another complication, since the disturbance in the accumulation area may well change again during the period in which the lower part of the glacier is responding, and thereby complex interacting responses may be set up.

Glacier surges have long been recognized, and reasonable qualitative explanations, invoking ideas of delayed response, "inertia", and bed peculiarities, have been put forward from time to time. The inherent unsteadiness of flow and of glacier dimensions received mathematical attention more than half a century ago, but it is only recently that decisive theoretical studies of unsteady motion have been applied. The basis of this theory has already been outlined in Chapter V (under kinematic wave theory).

The margins of the Antarctic ice sheet appear to be stationary or in slow retreat, and nowhere is there evidence of a general advance of the ice fringes.[14] This is not, however, considered to be incompatible with the strongly positive mass balance shown by mass economy studies, since an appreciable time lag is to be expected before the periphery responds to build-up in the interior. In Greenland the edges of the ice sheet are receding quite rapidly in many areas, although again the more recent mass balance studies give no indication of any strong negative balance. If any reliance can be placed in mass balance data, then the inland ice in both Antarctica and Greenland should be building up at the present time, although no advance of the ice margins is yet apparent.

Effects of climatic changes

The popular notion that all glaciers expand if the climate gets colder, and recede if the climate gets warmer, is completely unjustified as a generalization. Change of a climatic parameter will affect the regime of a given glacier according to the relative importance of the mass economy items which it controls.

Consider a long valley glacier which has an extensive ablation area and terminates on land. Such a glacier could well be expected to expand under colder air temperatures and shrink under warmer temperatures if no other climatic elements changed, as surface heating of the ice by convection has a strong effect on the mass balance. However, climatic parameters rarely change singly, and in a real case it would be necessary to inquire whether the change of air temperature was accompanied by a change in cloudiness (affecting the radiation balance) or increased precipitation (i.e., accumulation).

Now consider a portion of the Antarctic ice sheet, where ablation is almost negligible and most of the ice is lost by iceberg calving. A small increase of air temperature would have virtually no effect on ice wastage: it would do little to enhance the already-slight ablation, since radiation is the main factor there, and it would have to be sustained over a very long time period to affect ice outflow (by raising the temperature of the basal ice). If the higher air temperature was accompanied by increased precipitation (the warmer air having more moisture to deposit as snow), the net effect might be to move the mass balance in a positive direction, i.e., increase the total ice volume.

These two examples are extreme cases, but differing responses to climatic change can occur in the same glacier region. The advances and retreats of adjacent glaciers in a mountain region may not be synchronous, and this can often be traced to different distributions of accumulation and ablation areas with altitude, or different directional exposures (to wind and sun). In the southern Andes, for example, glaciers on the dry eastern side of the range have shrunk more rapidly in recent time than those on the wetter western slope.[16]

BIBLIOGRAPHY

1. Ahlmann, H. W. (1948) Glaciological research on the North Atlantic coasts, Royal Geographical Society (London), Research Series, no. 1.
2. ______ (1953) Glacier variations and climatic fluctuations, American Geographical Society (New York), Bowman Memorial Lecture no. 3.
3. Bader, H. (1961) "The Greenland ice sheet," in Cold Regions Science and Engineering (F. J. Sanger, Editor). USA CRREL monographs, Part I, Sect. B2.
4. Bauer, A. (1955) The balance of the Greenland ice sheet, Journal of Glaciology, vol. 2, no. 17.
5. Benson, C. S. (1960) Stratigraphic studies of the snow and firn of the Greenland ice sheet, California Institute of Technology, Ph.D. thesis.
6. Charlesworth, J. K. (1957) The Quaternary Era, vol. 1. London: Arnold.
7. Flint, R. F. (1957) Glacial and Pleistocene geology. Wiley.
8. Kotlyakov, V. M. (1960) "The intensity of nourishment of the Antarctic ice sheet," in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960) International Union of Geodesy and Geophysics. International Association of Scientfic Hydrology, Pub. no. 55.
9. ______ (1961) Snow cover in the Antarctic and its role in modern glaciation of the continent, IGY glaciology results no. 7. Moscow: USSR Academy of Sciences (text in Russian).
10. Lister, H. (1959) in Geophysical investigations of the Commonwealth Trans-Antarctic Expedition, Geographical Journal, vol. 125, parts 3-4.
11. ______ (1958) "Glaciology: the balance sheet or the mass balance," in Venture to the Arctic (R. A. Hamilton, Editor). Penguin Books.
12. Loewe, F. (1960) "Notes concerning the mass balance of the Antarctic inland ice," in Antarctic meteorology. Pergamon Press.
13. Mellor, M. (1959) Mass balance studies in Antarctica, Journal of Glaciology, vol. 3, no. 26.
14. ______ (1961) "The Antarctic ice sheet," in Cold Regions Science and Engineering (F. J. Sanger, Editor). USA CRREL monographs, Part I, Sect. B1.
15. Sharp, R. P. (1960) Glaciers. University of Oregon Press.
16. Mercer, J. H. (1962) Glacier variations in the Andes, American Geographical Society, Glaciological Notes, no. 12.
17. Rubin, M. J. (1962) Atmospheric advection and the Antarctic mass and heat budget. Antarctic Research, Geophysical Monograph no. 7.
18. Mellor, M. (1963) Remarks concerning the Antarctic mass balance, Paper presented to German Society for Polar Research, Karlsruhe, 1963.

CHAPTER VIII. TEMPERATURES IN GLACIERS.

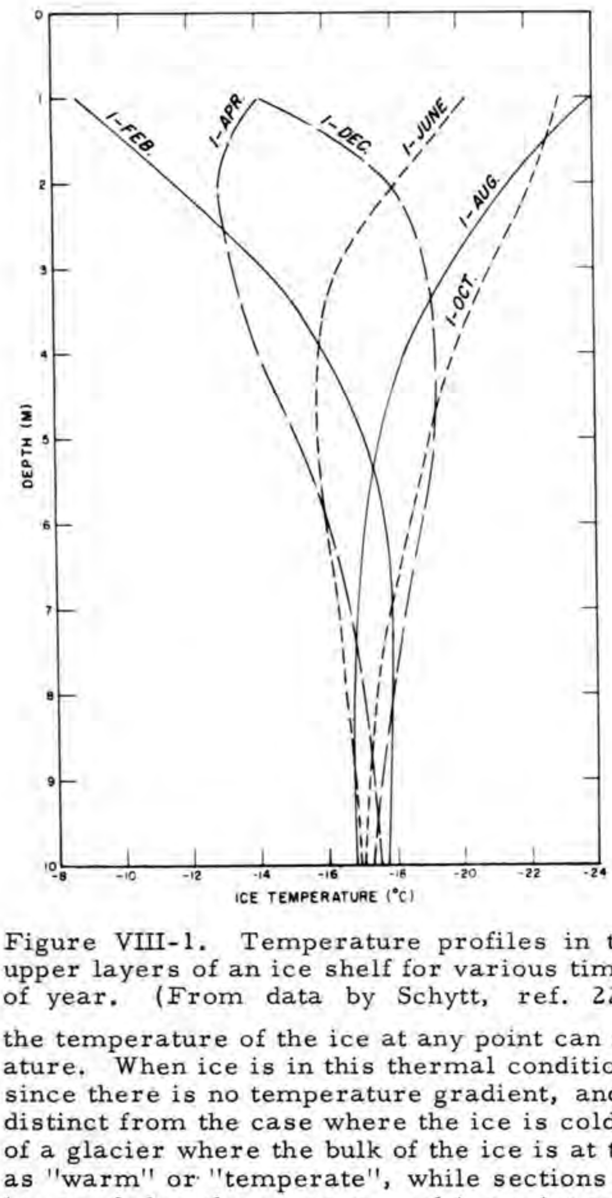

Figure VIII-1. Temperature profiles in the upper layers of an ice shelf for various times of year. (From data by Schytt, ref. 22)

When asked the naive question "How cold is a glacier?", the glaciologist finds himself hard-pressed to answer without going into lengthy explanations about the effects of location and climate, shape and dimensions, position and time, and so on. It is almost equally difficult to attempt a technical discussion of glacier temperatures in very general terms.

When glacier temperatures are measured in the field, it is found that they vary with position on the glacier and with depth in the ice. The ice temperature at shallow depths depends very largely on the surface heat balance, which is controlled by local climate and by the condition and exposure of the surface (see Chapter VI). Since air temperatures generally fall with increasing altitude, glaciers become colder at higher altitudes. The distribution of temperature through the thickness of a glacier, however, is also affected by geothermal heat flow from beneath and by heat liberated when the ice shears and slides. The vertical temperature profile is further complicated by time-dependent changes of temperature at the surface caused by the glacier's movement downhill or by climatic change.

There is, of course, an upper limit to the temperature of a glacier: the temperature of the ice at any point can never exceed the pressure melting temperature. When ice is in this thermal condition there can be no heat conduction through it, since there is no temperature gradient, and therefore the temperature problem is quite distinct from the case where the ice is colder than the pressure melting point. A section of a glacier where the bulk of the ice is at the pressure melting point is often referred to as "warm" or "temperate", while sections where temperatures through the bulk of the ice are below the pressure melting point are called "cold" or "polar". Some glaciologists classify entire glaciers as "warm" or "cold", but this is not good; many glaciers are "warm" in the lower (ablation) areas, and "cold" in the higher (accumulation) areas.

Temperatures at shallow depths

The snow and ice close to the surface of a glacier respond to cyclic fluctuations of surface temperature; the depth to which the fluctuations are felt depends on the period of the change and the thermal conductivity of the ice. In cold, dry areas free from melting, day-to-day temperature changes can be detected to depths of 2 or 3 feet, while annual temperature changes are noticeable to no more than 50 ft. The amplitude of a temperature cycle decreases with increasing depth below the surface, and there is a progressive time lag with depth.

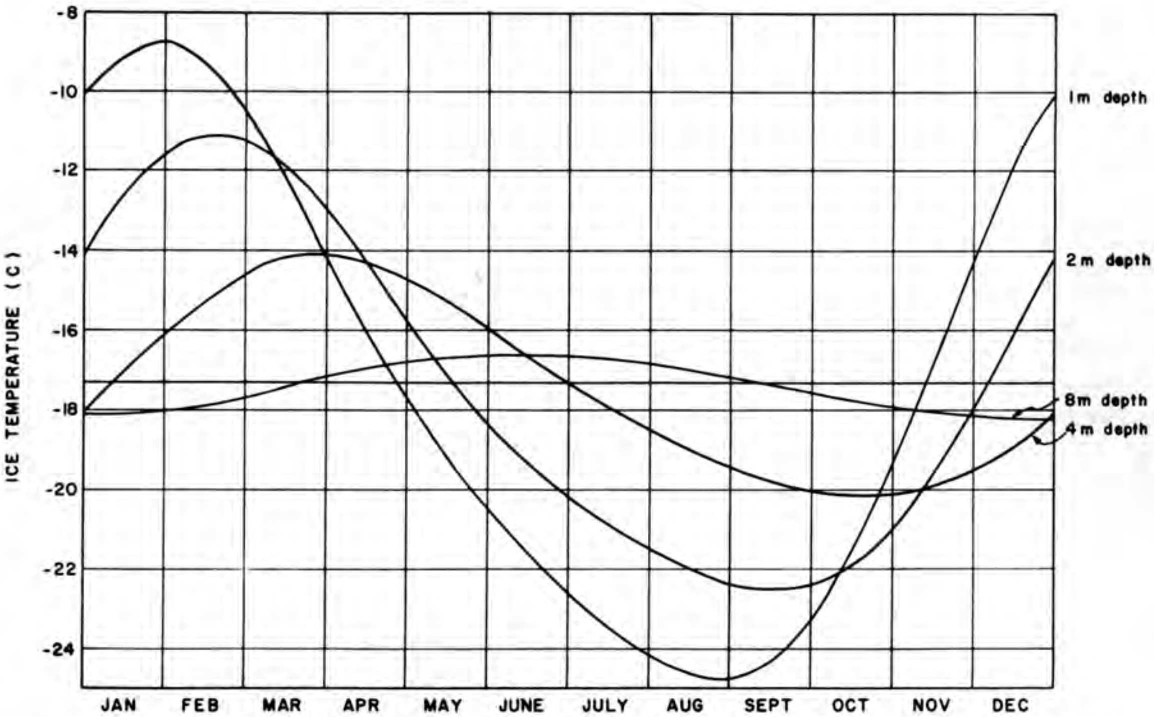

Figure VIII-2. Annual temperature waves at various depths in the upper layers of an ice shelf. Same data as those appearing in Figure VIII-1. (After Schytt, ref. 22)

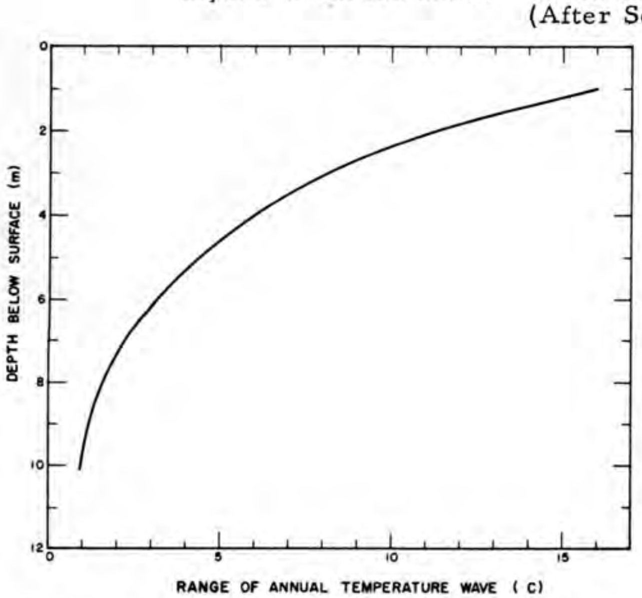

Figure VIII-3. Range of annual temperature waves plotted against depth. The exponential curve shows the attenuation of the waves drawn in Figure VIII-2. At 10 m depth the annual fluctuation of temperature is about 1C. (From data by Schytt, ref. 22)

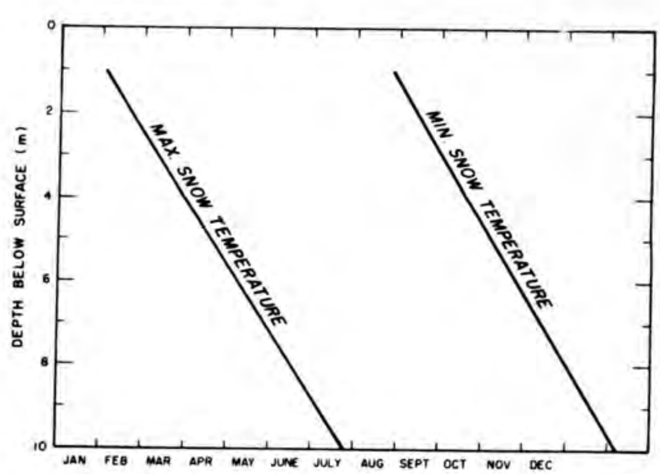

Figure VIII-4. Times of occurrence of maximum and minimum temperatures at various depths in the snow, from the data of Figure VIII-2. The slopes of the lines give the rate at which a temperature wave travels into the snow, and from this speed of penetration the effective thermal diffusivity of the snow can easily be calculated (see text). These lines show a rate of penetration of about 1.6 m per month. (From data by Schytt, ref. 22)

Figure VIII-2 shows annual temperature waves at depths of 1, 2, 4, and 8 m for a point on an Antarctic ice shelf.[22] The attenuation of the wave with depth is clearly seen: at 1 m depth there is a temperature range of about 16C, while at 8 m depth the wave range is less than 2C. In Figure VIII-3 temperature range is plotted against depth. The amplitude attenuation with depth means that, at depths below 10 m, the temperature is close to the mean annual temperature throughout the year. Thus it is possible, in a cold snow area, to obtain the mean annual temperature of the snow (which is very close to the mean annual temperature of the air above) simply by drilling a hole and measuring the temperature at 10-15 m depth. The time lag also shows clearly in Figure VIII-2: at 1 m depth the highest temperature occurs at the end of January, shortly after the warmest period of summer, but at 8 m depth the maximum temperature is not reached until the month of June, which is mid-winter in Antarctica. This also gives the average rate at which a temperature wave travels into snow of this density and type — about 5 cm/day (or 2 in./day) (Fig. VIII-4).

The effects of surface temperature change can be studied analytically by making the assumption that the ice is a semi-infinite solid with horizontal temperature gradients which are negligible compared to the vertical temperature gradients, and then applying the linear conduction equation from standard theory.[6] The basic heat conduction equation is

$$\frac{\partial^2 \theta}{\partial x^2} - \frac{1}{\alpha}\frac{\partial \theta}{\partial t} = 0 \tag{1}$$

where

θ = temperature
x = depth
α = thermal diffusivity ($\alpha = \frac{K}{\rho c}$, where K is thermal conductivity, ρ is density, and c is specific heat)
t = time.

This has to be solved for the case where the surface temperature is a harmonic function of time, and so an expression for the surface temperature is required. In an actual problem the observed surface temperature fluctuation, e.g., the annual measured surface temperature wave, would be represented by a suitable combination of Fourier components, but for illustration a simple sine wave can be used here:

$$\theta_{0,t} = \theta_m + A_0 \sin(2\pi nt) \tag{2}$$

where
$\theta_{0,t}$ = surface temperature at time t
θ_m = mean temperature (around which the fluctuations occur)
A_0 = amplitude of the surface temperature wave
n = frequency of the temperature wave.

The solution of eq 1 for this condition is

$$\theta_{x,t} = \theta_m + A_0 e^{-x\sqrt{\frac{\pi n}{\alpha}}} \sin\left(2\pi nt - x\sqrt{\frac{\pi n}{\alpha}}\right). \tag{3}$$

Details of the solution are given in standard texts.

Equation 3 is a sine wave equation similar to eq 2, but in it the wave amplitude is reduced by a factor $e^{-x\sqrt{\frac{\pi n}{\alpha}}}$, and the wave at time t lags behind the surface wave by an angle of $x\sqrt{\frac{\pi n}{\alpha}}$ radians. Thus, at depth x below the surface the wave amplitude and the lag are given by

Amplitude, $A_x = A_0 \, e^{-x\sqrt{\frac{\pi n}{\alpha}}}$

Time lag, $\Delta t = \frac{1}{2\pi n} \cdot x\sqrt{\frac{\pi n}{\alpha}} = \frac{x}{2\sqrt{\pi n \alpha}}$

and the temperature wave travels into the ice with an apparent velocity

Velocity, $V = 2\sqrt{\pi n \alpha}$.

From the above relationships it is possible to obtain values for the effective thermal diffusivity or conductivity from suitable field temperature data. Alternatively, if the thermal conductivity is calculated independently, surface temperature data can be used to predict the effects at depth. For most practical purposes the thermal conductivity of high-density dry snow (density > 0.35 g/cm^3) is given by the Kondrat'eva equation[18]

$$K = 0.0085\ \rho^2$$

where K = thermal conductivity (cal/cm-sec-C)
ρ = snow density (g/cm^3)

and the thermal diffusivity α is defined as

$$\alpha = \frac{K}{\rho C}$$

where c is specific heat of the snow ~ 0.5.

While the above expressions give good values for homogeneous snow samples, the effective conductivity of a thick snow layer under natural conditions may be affected by the presence of buried ice crusts or depth hoar layers.*

The penetration of surface temperature changes into warm ice does not follow the same pattern. Although air temperature above the ice ranges from below-freezing winter minimums to above-freezing summer maximums, ice temperature can rise only to a maximum of 0C. Temperature records in warm ice for a complete year are not readily available from the literature, but the general form of the time-temperature curves can be deduced.

In the height of summer, when air temperatures are above freezing, the ice stays at a constant temperature of 0C. In the autumn, when daily mean temperatures drop below freezing point, the winter "cold wave" begins to penetrate the ice, and the temperature at, say, 2 m depth starts to fall below 0C three weeks or so after the mean surface temperature drops below freezing. The cold wave in the ice lags behind the surface change, and the amplitude is attenuated with increasing depth. In spring the ice warms up again, the temperature at 2 m reaching 0C again two or three weeks after the surface has come to the melting point.

Figure VIII-7 shows temperature changes in permeable firn, where the form of the winter cold wave is modified by melt-water percolation from the surface. During the summer the firn stays at the melting point. In the autumn, daily mean air temperatures drop below freezing, and the snow surface temperature drops below freezing after a delay period, caused by absorption of latent heat as the free water in the surface snow freezes. The presence of dispersed free water in the pores of the snow also slows down the rate at which temperature falls at any given shallow depth. In the summer, the surface again melts and water percolates into the permeable snow, accelerating the rate of

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*In this chapter convective heat transfer by flow of air in the pores of the snow is neglected. Actually there is some motion of the interstitial air in the upper layers when surface pressure changes occur.

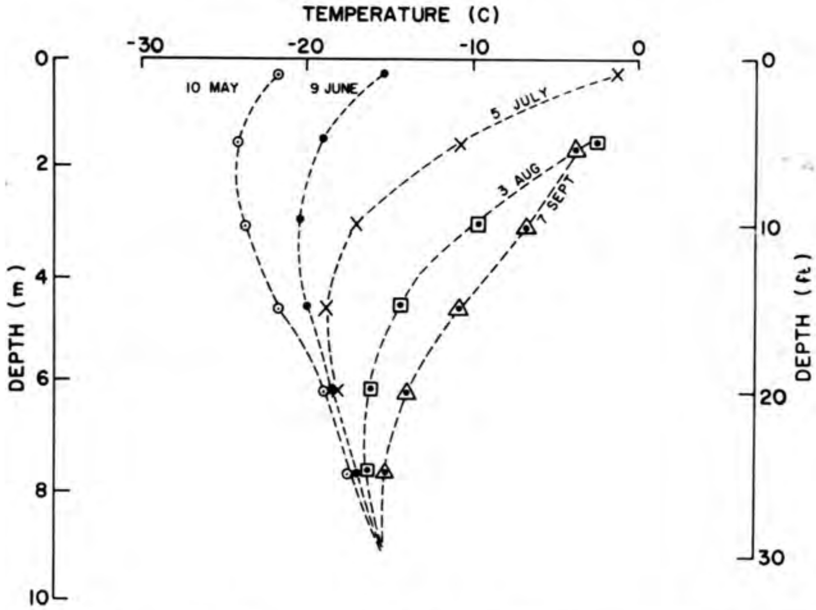

Figure VIII-5. Changes in the temperature distribution in the upper layers of the ice island T-3 during the summer months. (After Crary, ref. 8)

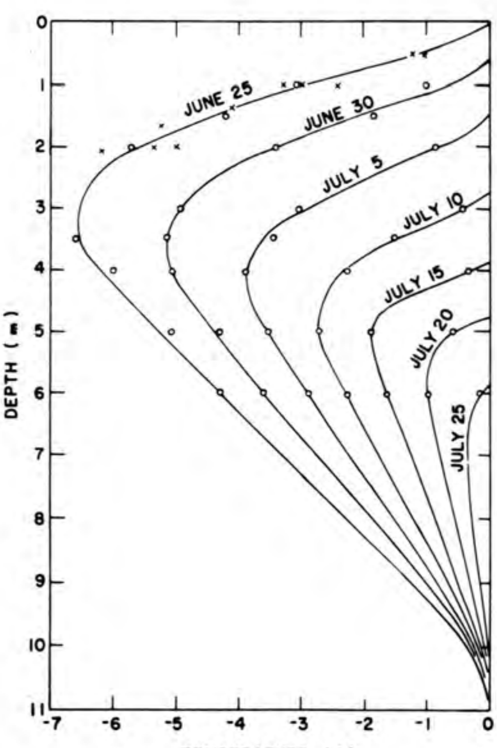

Figure VIII-6. Decay of the winter cold wave in the upper layers of "warm" glacier ice. (After Ahlmann, ref. 1)

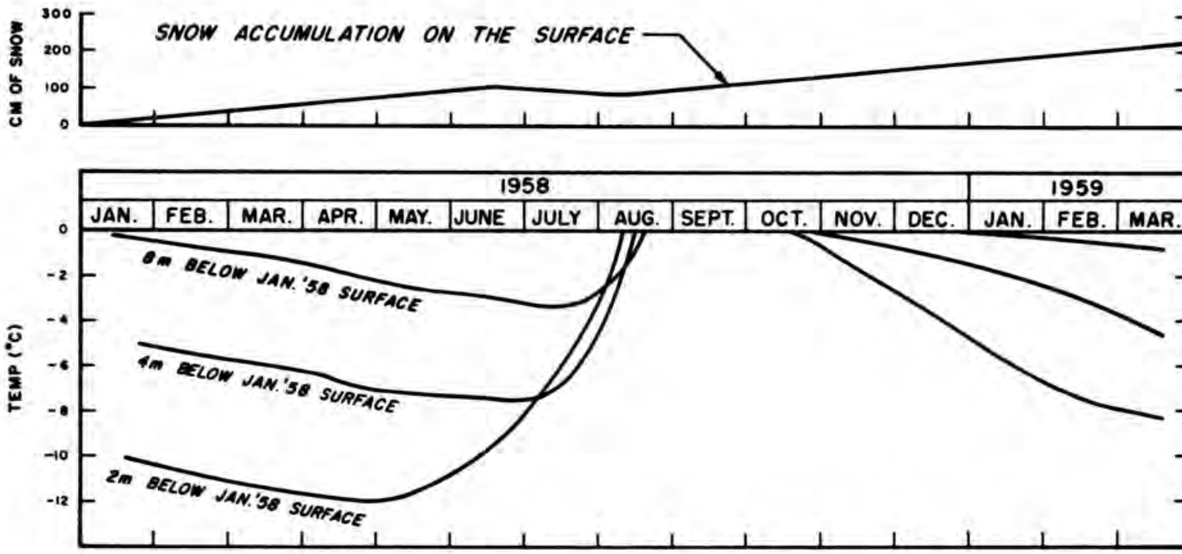

Figure VIII-7. Annual temperature changes in snow subject to summer melting at the surface and infiltration of melt water. (From data by Chizhov, ref. 7)

temperature rise at shallow depths. The temperature waves in Figure VIII-7 are progressively attenuated through the period shown, because of snow accumulation on the surface. In situations where melting is strong and thick layers and lenses of ice develop in the snow the temperature changes are more complicated.

Temperature variations with altitude and latitude

In parts of a glacier where the ice is thick and the upper layers are cold (non-melting), the temperature measured 10 to 15 m below the surface gives a good approximation to the mean annual surface temperature. This surface temperature depends upon solar radiation, which varies with latitude, and upon convection, which is influenced by movement of air over the surface.

In Greenland and Antarctica, measurement of "10-meter temperatures" in boreholes and pits has yielded a mass of data on the distribution of mean annual surface temperatures, and it has been possible to draw isotherms for mean annual surface temperature, even though very few places have received year-round meteorological study. Figures VIII-8 and -9 show these isotherms.

Inspection of Figures VIII-8 and -9 shows that mean annual temperatures are strongly dependent on surface altitude, and that they are only slowly affected by changes of latitude. In Figures VIII-10 and -11, mean annual temperature is plotted against surface altitude for various parts of Greenland and Antarctica. From these it can be shown (by isolating regional peculiarities and latitude effects) that the temperature lapse rate is about 1C/100 m in regions where katabatic winds predominate. This agrees with the dry adiabatic lapse rate. Regions exposed to maritime influence and to uplifting of cyclonic air masses appear to have smaller lapse rates; Langway[13] mentions lapse rates of about 0.6C/100 m in parts of north Greenland, and Schytt's curve for Dronning Maud Land[22] (Fig. VIII-11) shows a similar lapse rate below 1000 m. Both regions are subject to influx of maritime air; the lapse rates are close to that of the standard atmosphere. West Antarctica, which is subject to more cyclonic penetration than East Antarctica, appears to have generally smaller altitude lapse rates.

The effect of latitude on temperature is deduced by comparing temperatures for points which are at the same altitude but at different latitudes. For both Greenland and

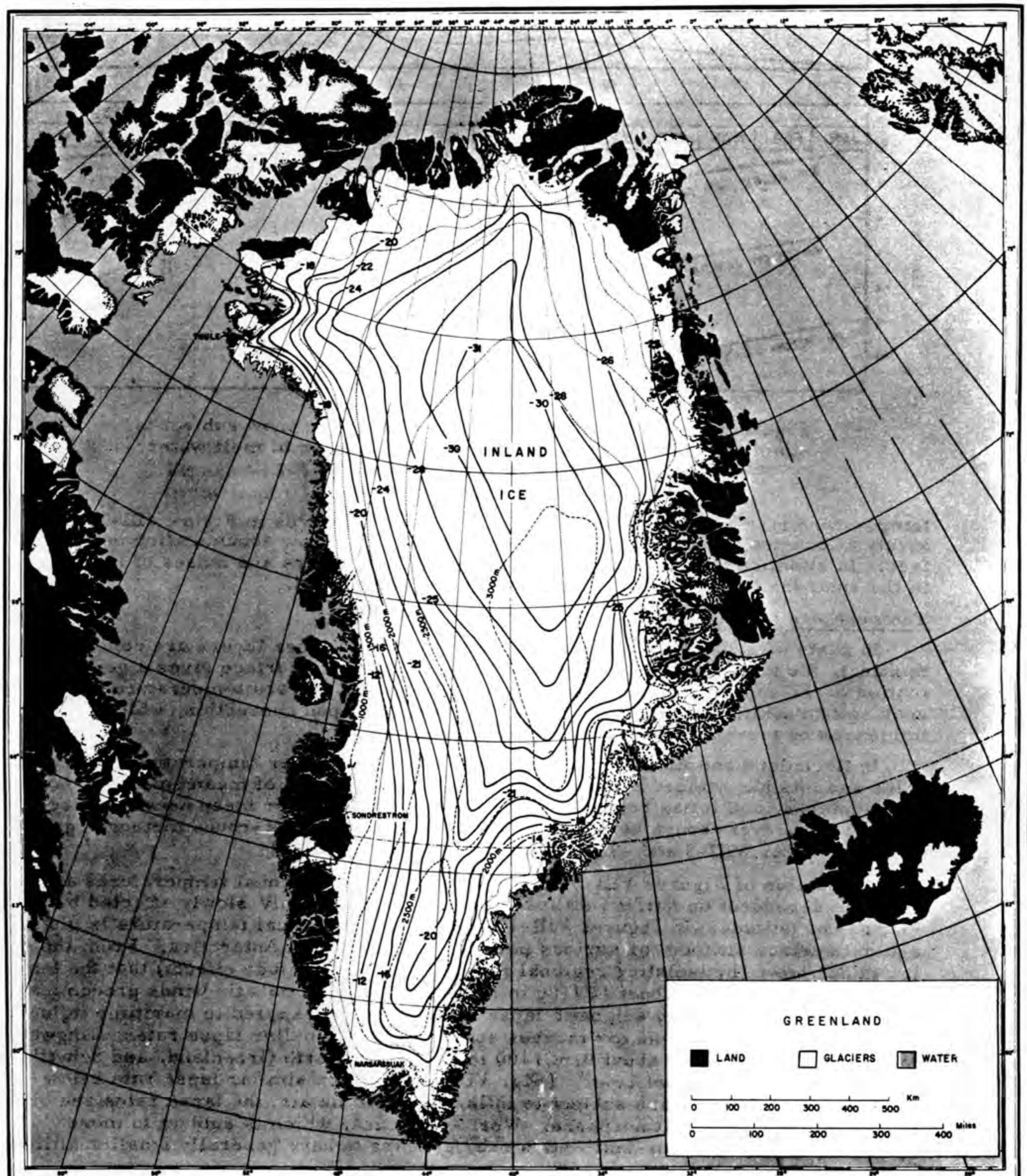

Figure VIII-8. Mean annual surface temperatures on the Greenland ice sheet. (After Bader, ref. 27)

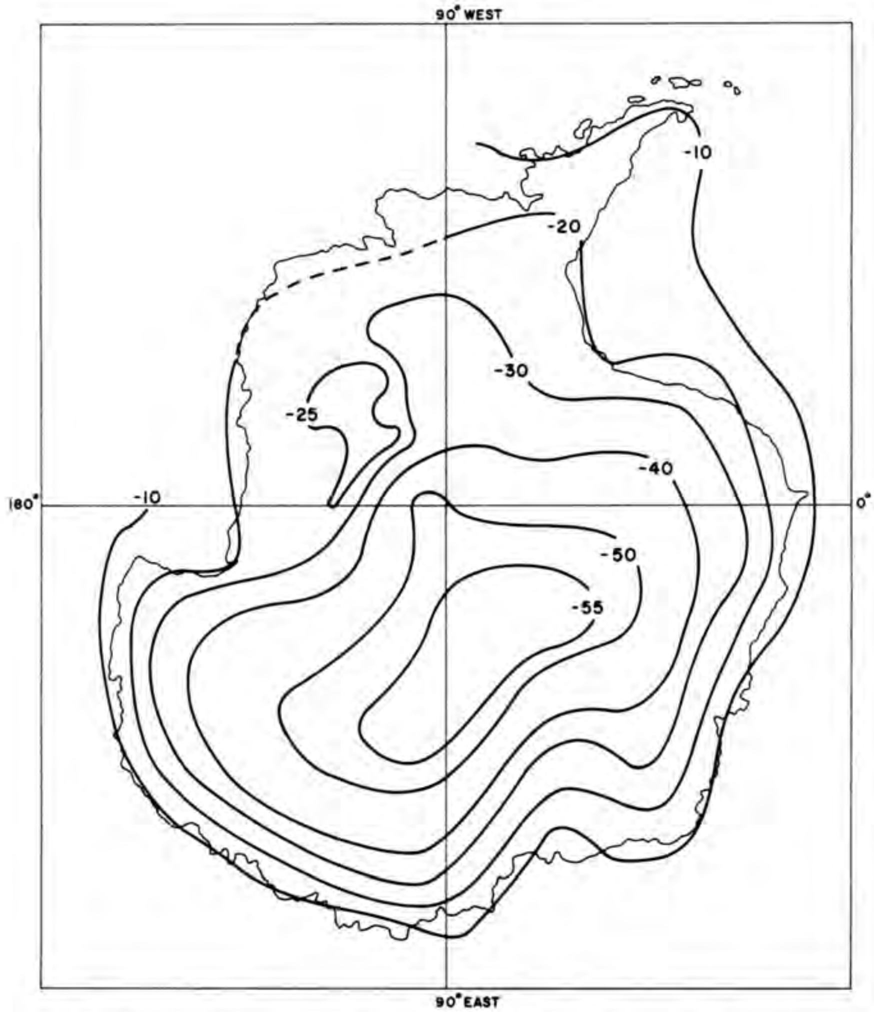

Figure VIII-9. Mean annual surface temperatures on the Antarctic ice sheet. (After Rubin, ref. 21)

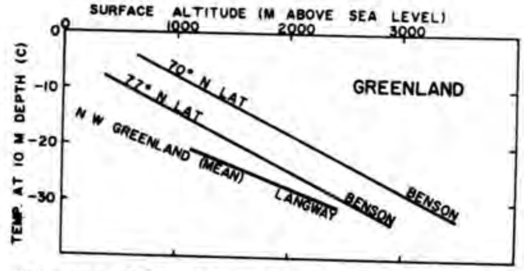

Figure VIII-10. Mean annual surface temperature (10 m temperature) plotted against surface elevation for the Greenland Ice Cap. (After Benson, ref. 3 and Langway, ref. 13)

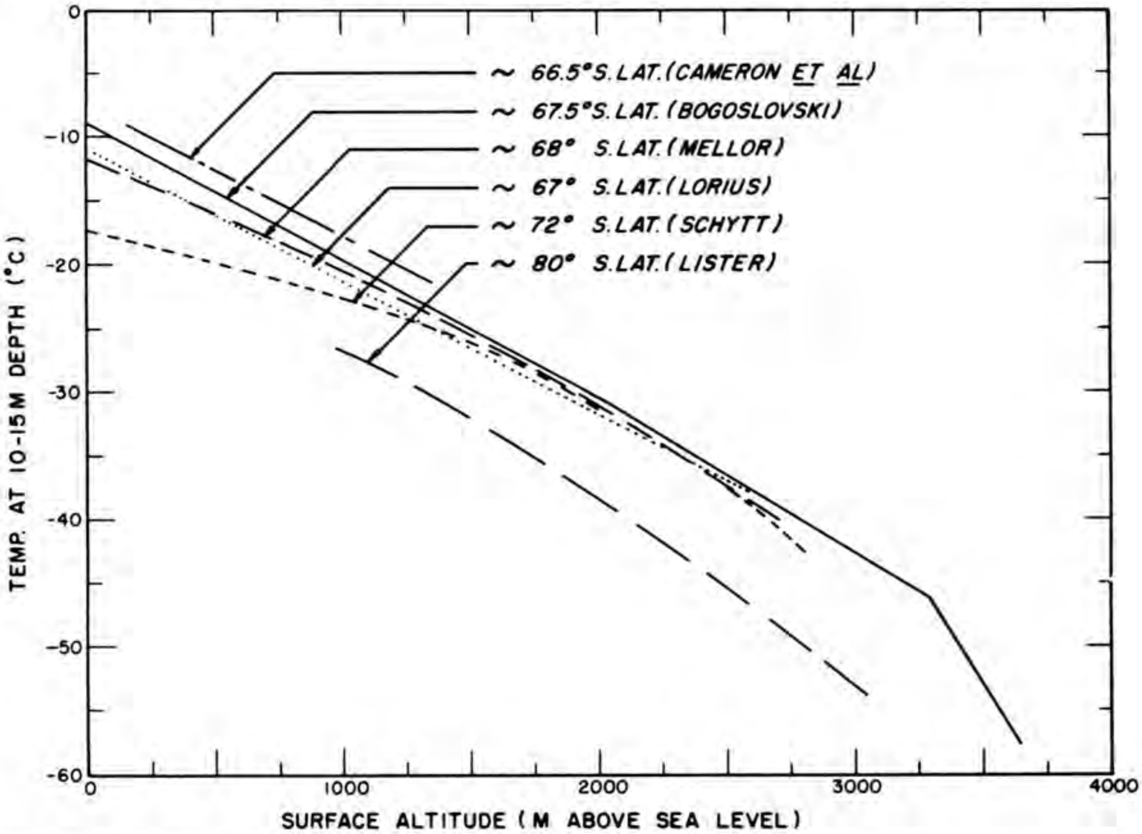

Figure VIII-11. Mean annual temperature (10-15 m temperature) plotted against surface elevation for the Antarctic ice sheet. (Refs. 4, 5, 14, 15, 17, and 22)

Antarctica it is found that temperature changes with latitude at a rate of roughly 1C per degree of latitude. In Figure VIII-11 the curves are noticeably influenced by the latitude effect at altitudes above 2500 m, since there is only a very small surface gradient on the ice sheet at these elevations.

Temperatures deep within the ice

On low-altitude glaciers of the temperate and sub-polar regions, the ice below the level of seasonal temperature variation (say below 15 m or 50 ft) is usually almost isothermal at the pressure melting point. If the ice is at pressure melting point throughout, there is a very slight temperature gradient of about -0.065 C/100 m depth, corresponding to the melting point depression by increasing overburden pressure (0.0075 C per atmosphere of pressure). Since there can be no conduction through ice which is at the melting point, virtually all the heat supplied to the surface of the glacier (by radiation, convection, and condensation) and to the base of the glacier (geothermal flux) is used in melting ice. The geothermal heat flow is very small, averaging about 1.2×10^{-6} cal/cm^2-sec, or enough heat to melt 5 mm of ice per year.

On cold glaciers and ice shelves, the ice below the level of seasonal variations is frequently found to have a "positive" temperature gradient, i.e., the temperature increases with depth from the mean annual surface temperature at 15 m, up to a maximum value (often the pressure melting temperature) at the glacier bed. In Figures VIII-12 and -13 some temperature profiles are shown for glaciers and ice shelves which have stable positive gradients. In the case of glaciers which have basal temperatures below the pressure melting point, and positive thermal gradients, heat can be

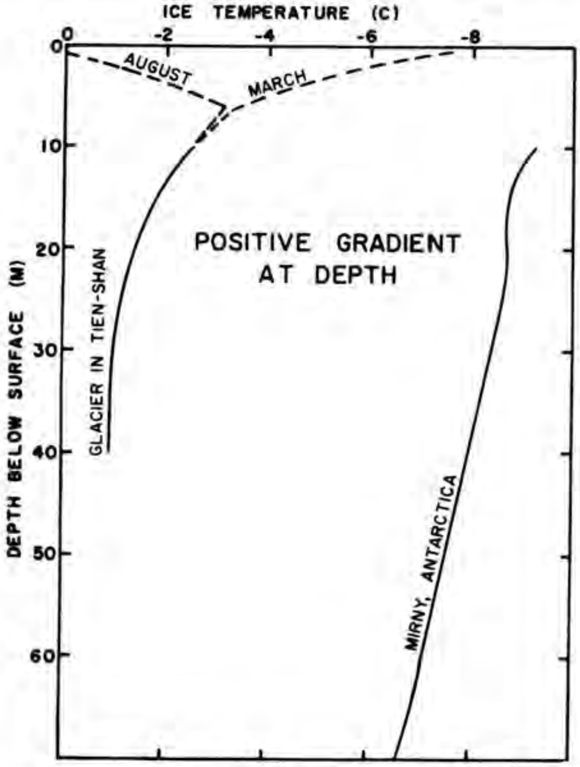

Figure VIII-12. Two temperature profiles in which the temperature increases with depth (positive temperature gradient). (After Bogoslovski, ref. 4, and Tsykin, ref. 25)

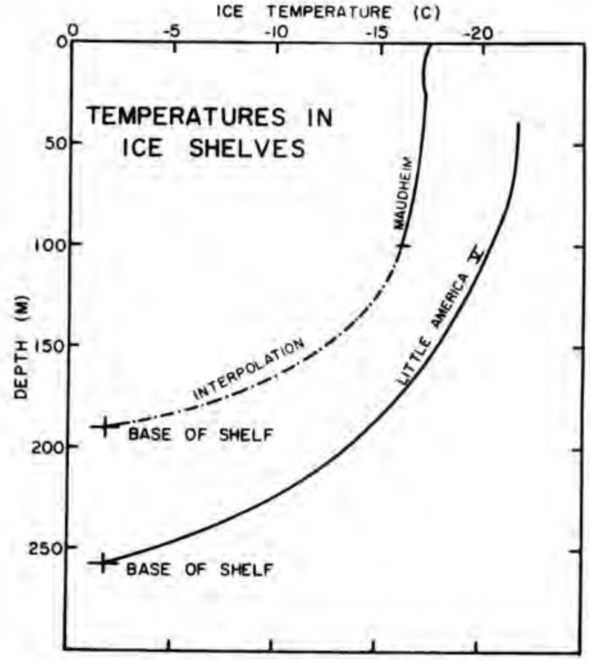

Figure VIII-13. Temperature profiles in Antarctic ice shelves. The temperature gradients here are positive. (After Bender and Gow, ref. 2, and Schytt, ref. 22)

conducted up through the ice. If the base is at the melting point, however, heat supplied to the base goes into melting ice. At the base of an ice shelf the temperature is at the melting point, but the amount of melting taking place still cannot be calculated with confidence.

The two types of temperature distribution mentioned so far, the isothermal "warm" ice and the "positive" gradient, are plainly stable situations, but a third type not yet described is more puzzling. It has been found from measurements in deep boreholes on the Greenland and Antarctic ice sheets (Fig. VIII-14) that temperature frequently decreases with increasing depth, i.e., there is a negative temperature gradient, and this seems somewhat curious at first sight, since a negative gradient cannot exist as a steady-state condition in a simple static situation. A negative gradient could be developed, however, if surface temperature increased with time.

Surface temperature on an ice sheet can increase with time in two ways: by general secular warming over a region as a result of climatic change, or by movement of a given vertical column of ice from a colder to a warmer local area. The second process should be explained more fully, since it is not immediately obvious.

As was seen above, the mean annual temperature on the surface of an ice sheet varies chiefly with elevation, so that as ice moves outward and downward, each succeeding snow layer deposited above a particular vertical column is laid down at a higher mean temperature than the previous one. The relative magnitudes of velocity, slope, lapse rate, accumulation rate and thermal diffusivity are such that small negative gradients are formed by this process over much of Greenland and Antarctica. The effect was first recognized and described quantitatively by Robin,[20] who derived a simple expression for the gradient by geometrical reasoning, assuming that conduction could be neglected:

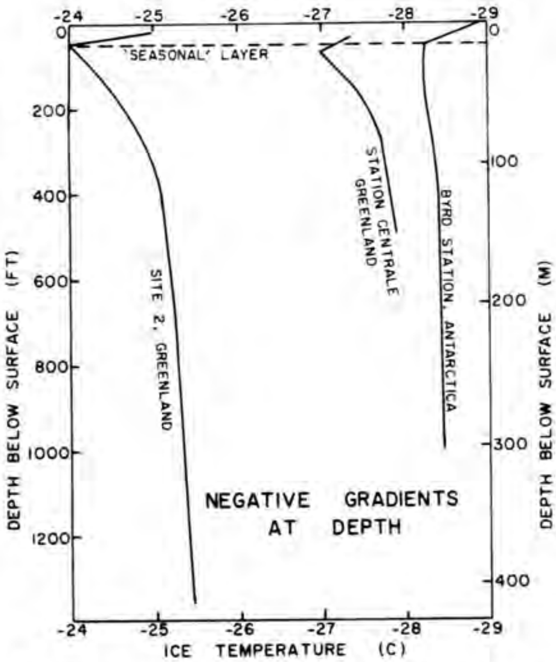

Figure VIII-14. Deep temperature profiles from Greenland and Antarctica in which the temperature gradients are negative. (After Hansen and Landauer, ref. 10; Heuberger, ref. 11; Bender and Gow, ref. 2)

Temperature gradient,

$$\frac{d\theta}{dx} = - \frac{\phi \lambda V}{v} \quad (4)$$

where ϕ = surface slope, λ = vertical lapse rate of mean annual temperature, $\underline{V}$ = horizontal ice surface velocity, $\underline{v}$ = downward velocity of a given layer relative to the surface (which corresponds to rate of snow accumulation near the surface).

Radok[19] gave the problem a more rigorous treatment by regarding the ice sheet as a semi-infinite solid and taking a system of co-ordinates which moves at the same horizontal speed as the ice, so that the equation for vertical linear conduction at depth $\underline{x}$ and time $\underline{t}$ becomes:

$$a \frac{\partial^2 \theta}{\partial x^2} - v\frac{\partial \theta}{\partial x} - \frac{\partial \theta}{\partial t} = 0 \quad (5)$$

where a = thermal diffusivity, θ = temperature, $\underline{v}$ = vertical velocity of a given ice layer relative to the surface. Horizontal temperature gradients in the ice are quite uniform, and therefore do not enter into consideration.

An appropriate solution was found by adapting a result obtained by Benfield while studying a similar geological problem. This solution, which has been called the Benfield-Radok equation, gives transient temperature gradients and is rather complex, but it reduces to a very simple asymptotic form as $t \to \infty$. This steady-state form is identical to Robin's relationship (eq 4). Figure VIII-15, from new work by Jenssen and Radok, shows graphically the lengths of time necessary for transient gradients to reach the limiting asymptotic gradient for various snow accumulation rates.

For real ice sheets and ice shelves of finite thickness the Benfield-Radok equation is not applicable when the complete temperature distribution through the ice is considered, and an exact solution of equation 5 for appropriate boundary conditions is hardly feasible in view of the complexity. Jenssen and Radok[12] have therefore put equation 5 into finite difference form and solved it numerically with the aid of a digital computer for particular models. Taking into account variation of vertical velocity and thermal diffusivity with depth, the finite difference form of equation 5 is:

$$a_x \left(\frac{\theta_{x+1,t} - 2\theta_{x,t} + \theta_{x-1,t}}{\Delta x^2} \right) + \left[\left(\frac{a_{x+1} - a_{x-1}}{\Delta x} \right) - V_x \right] \left(\frac{\theta_{x+1,t} - \theta_{x-1,t}}{2\Delta x} \right) = \left(\frac{\theta_{x,t+1} - \theta_{x,t}}{\Delta t} \right). \quad (6)$$

The surface temperature change can take any form, i.e., both climatic warming and the downhill flow of ice can be considered. Results obtained so far suggest that surface warming by downhill flowismore important than climatic change in establishing the negative gradients observed in Greenland and Antarctica.

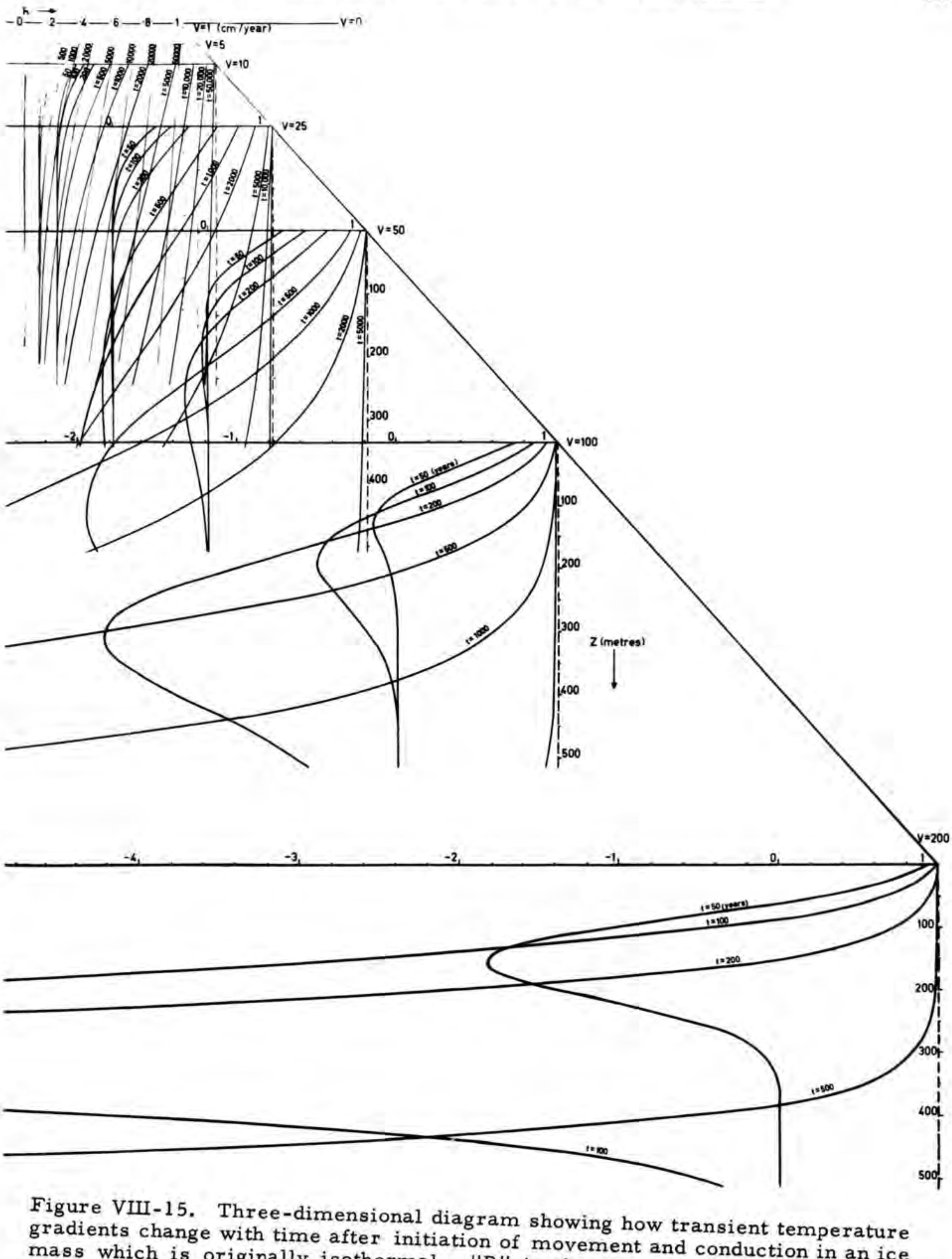

Figure VIII-15. Three-dimensional diagram showing how transient temperature gradients change with time after initiation of movement and conduction in an ice mass which is originally isothermal. "R" is the ratio of actual temperature gradient (for given depth and accumulation rate) to the asymptotic gradient. "Z" is depth below surface in meters. "V" is accumulation rate, the the V-axis is intended to be perpendicular to the plane of the paper. (Jenssen and Radok, to be published)

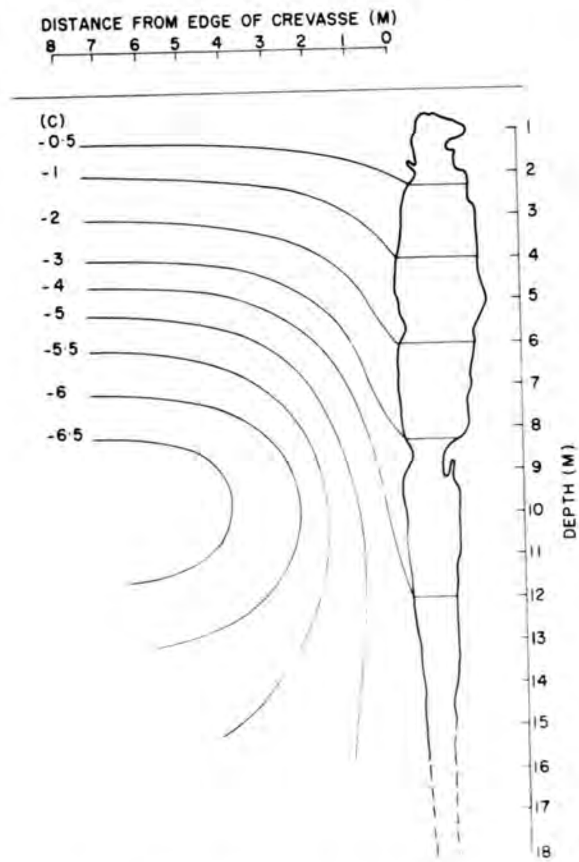

Figure VIII-16. Disturbance of the temperature field where a crevasse breaks through the ice, permitting convection to take place. (After Meier, et al., ref. 16)

The problems of temperature distribution deep within the ice should become more tractable when data from the new super-deep boreholes become available.

Temperatures in the vicinity of a crevasse

Figure VIII-16 shows how the temperature field is distorted near a crevasse. The temperature difference between a snow bridge over a crevasse and the surrounding snow is the basis of a technique for mapping crevasses by infrared photography from an airplane.

BIBLIOGRAPHY

1. Ahlmann, H. W. (1948) Glaciological research on the North Atlantic coasts, Royal Geographical Society (London), Research Series, no. 1.

2. Bender, J. A. and Gow, A. J. (1960) "Deep drilling in Antarctica," in Symposium on Antarctic Glaciology, General Assembly of Helsinki, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55.

BIBLIOGRAPHY (Cont'd)

3. Benson, C. S. (1960) Stratigraphic studies of the snow and firn of the Greenland ice sheet, California Institute of Technology, Ph. D. thesis.

4. Bogoslovski, V. N. (1958) "The temperature conditions and movement of the Antarctic glacial shield," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

5. Cameron, R. L. et al., (1959) Wilkes Station glaciological data 1957-58, Ohio State University Research Foundation, Report 825-1, Part III.

6. Carslaw, H. S. and Jaeger, J. C. (1959) Conduction of heat in solids. Oxford: Clarendon Press.

7. Chizhov, O. P. (1961) "Glaciological investigations in Novaya Zemlya in 1957-59," in Study of glaciers and glacier regions (ed. G. A. Avsyuk). Moscow: USSR Academy of Sciences (text in Russian).

8. Crary, A. P. (1958) Arctic ice island and ice shelf studies, Arctic, vol. 11, no. 1.

9. Giovinetto, M. J. (1960) Glaciology report for 1958, South Pole Station, Ohio State University Research Foundation, Report 825-2, Part IV.

10. Hansen, B. L. and Landauer, J. K. (1958) "Some results of ice cap drill hole measurements," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

11. Heuberger, J. C. (1954) Forages sur l'inlandsis (Bore-hole studies on the ice cap). Glaciologie, Groenland, vol. I. Paris: Expéditions Polaires Francaises.

12. Jenssen, D. and Radok, U. (1960)"Transient temperature distributions in ice caps and ice shelves," in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55.

13. Langway, C. C. (1961) Accumulation and temperature on the inland ice of north Greenland, 1959, Journal of Glaciology, vol. 3, no. 30.

14. Lister, H. (1959) in Geophysical investigations of the Commonwealth Trans-Antarctic Expedition, Geographical Journal, vol. 125, parts 3-4.

15. Lorius, C. (1962) Contribution to the knowledge of the Antarctic ice sheet, Journal of Glaciology, vol. 4, no. 31.

16. Meier, M. A., et al. (1957) Preliminary study of crevasse formation, U. S. Army Snow, Ice and Permafrost Research Establishment, Technical Report 38.

17. Mellor, M. (1960) Temperature gradients in the Antarctic ice sheet, Journal of Glaciology, vol. 3, no. 28.

18. ________ (1962) "Polar snow — a summary of engineering properties," in Ice and snow — processes, properties and applications (W. D. Kingery, Editor). Cambridge, Mass.: MIT Press.

19. Radok, U. (1959) Temperatures in polar ice caps, Nature, vol. 184, 3 October.

20. Robin, G. de Q. (1955) Ice movement and temperature distribution in glaciers and ice sheets, Journal of Glaciology, vol. 2, no. 18.

21. Rubin, M. J. (1962) The Antarctic and the weather, Scientific American, vol. 207, no. 3.

22. Schytt, V. (1960) Glaciology II, part D, Scientific results of the Norwegian-British-Swedish Antarctic Expedition, vol. IV, Oslo.

BIBLIOGRAPHY (Cont'd)

23. Sorge, E. (1935) "Glaziologische Untersuchungen in Eismitte, (Glaciological Investigations at Eismitte)," in Wissenschaftliche Ergebnisse der Deutschen Gronland Expedition Alfred Wegener 1929 und 1930-31, Band 3. Leipzig: F. A. Brockhaus.

24. Tien, C. (1960) Temperature distribution of idealized ice cap, USA SIPRE Research Report 64.

25. Tsykin, Y. N. (1962) The mass income in the firn zones of glaciers, IGY glaciology results, no. 8. Moscow: USSR Academy of Sciences (text in Russian).

26. Wexler, H. (1960) Heating and melting of ice shelves, Journal of Glaciology, vol. 3, no. 27.

27. Bader, H. (1961) "The Greenland ice sheet," in Cold Regions Science and Engineering (F. J. Sanger, Editor), U. S. Army Cold Regions Research and Engineering Laboratory Monographs, Part I, Sect. B2.

CHAPTER IX. PAST GLACIATIONS.

Evidence of past glaciations

By observing the behavior of existing glaciers it can be seen that they leave their marks upon the landscape in unmistakable fashion, and, if present-day glaciers were suddenly to melt away and disappear, clear evidence of their previous existence would remain. It is from characteristic erosion forms and from characteristic deposits that the extent of former glaciers has been deduced, revealing that more than one-quarter of the earth's surface bears marks of glaciation. Some typical indicators of past glaciation are listed below.

Valley profiles. A valley formerly occupied by a glacier tends to have a U-shaped cross-section, in contrast to the various V-shapes more typical of river valleys cut solely by water action.* The longitudinal profiles of glacial valleys are sometimes stepped, and lakes may form in basins left on the "steps".

Hanging valleys. Some tributary glaciers leave valleys which have their floors considerably higher than the floor of the main valley at the place where the two join. After deglaciation, mouths of the tributary glaciers are high up the side of the main valley, and waterfalls form.*

Truncated spurs (faceted spurs). Glaciers tend to follow pre-existing valley systems, but modify the original forms by ice erosion. One of these modifications is a squaring-off of spurs separating adjacent tributary valleys as a result of the main glacier broadening its valley.

Fjords. Recession of coastal valley glaciers has left long narrow tidewater inlets in some regions (e.g., the west coast of Norway, the coasts of Greenland, the southwestern coast of South Island, New Zealand). Fjords are frequently steep-sided and the water is relatively deep.

Cirques (corries, cwms). Mountain glaciers frequently erode the mountainside at their heads, producing steep-sided, bowl-shaped depressions. Deep hollows scooped out of the mountainside by ice in this form are called cirques (French), corries (Scottish), or cwms (Welsh). Small lakes often form in the basin at the bottom of the cirque; they are usually known as tarns.

Alpine sculpture. Glacier erosion and frost action produce the characteristic sharply cut topography of an alpine region. As the cirques of adjacent glaciers develop, the ridge separating them becomes narrow and steep-sided, eventually reaching the sharp-edged form known as an arete. Cirques eating back from opposite sides of a ridge break through to leave a jagged gap referred to as a col. The effect of erosion at the headwalls of cirques all around a mountain is to steepen and sharpen the peak into the characteristic "horn" shape.

Striations (striae) and roches moutonées. The exposed rock of glaciated terrain often shows evidence of scraping and smoothing by rock embedded in moving ice. Rock surfaces may bear long scratches, or striations, where they have been scraped by ice-borne rock fragments. These striations indicate the direction of motion of the former glacier. Hummocks of outcropping rock may be smoothed into a rounded shape, with the former upstream side having gentle slopes and the downstream side remaining abrupt. These are known as roches moutonées.

Crag-and-tail. Where hard and soft rocks occur together, bosses of the harder material are left protruding after the glacier has eroded preferentially. A wedge of moraine material is deposited in the lee of the boss, or crag, and the resulting feature is called a crag-and-tail.

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* Both U-shaped valleys and hanging valleys can be formed in other ways, but glacially sculptured ones are almost unmistakable when all the evidence is considered.

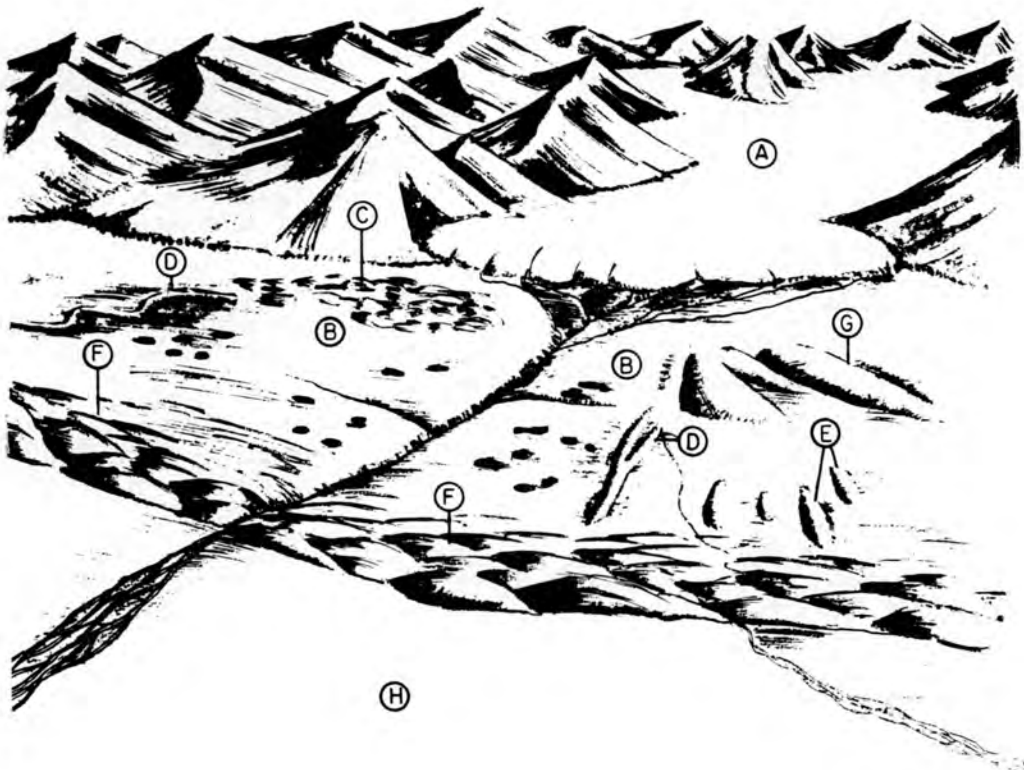

Figure IX-1. Sketch illustrating some glacial land forms: A - valley glacier; B - ground moraine, or till plain; C - kettles and kames; D - eskers; E- drumlins; F - terminal moraine, or end moraine; G - glaciated rock outcrops; H - outwash plain. (From EM 1110-1-377, CRREL)

Figure IX-2. Glaciated mountain landscape. The main valley has the characteristic U-shaped cross-section, and hanging valleys are discernible to the sides.

Till. Areas once covered by glaciers bear deposits of material in which the rocks and soil particles are not well sorted, and a wide range of grain sizes, from fine clay particles to large boulders, can occur. This is till. Boulder clay is one form of till; other kinds may be predominantly stony, with a low fines content.

Moraines. When glaciers recede, their moraines remain to become part of the new topography, lateral moraines forming stony ridges along the edges of valleys, and terminal moraines becoming transverse ridges (which sometimes dam the valley to form a lake). Most of an area previously covered by an ice sheet is covered by ground moraine of till.

Eskers and kames. Glacial streams deposit coarse sediments of sand and gravel, producing gravelly ridges known as eskers (often resembling sinuous railroad embankments about 50 ft high). These mark the courses of former subglacial streams. Similar deposits along the edges of a glacial valley are called kame terraces. Gravelly mounds formed by deposition from moraine pockets, or where melt streams debouched from the ice, are called kames. Eskers are valuable sources of non-frost-susceptible soils for construction.

Drumlins. Drumlins are streamlined whaleback mounds, usually of hard till, deposited by moving ice and nearly always present in large numbers, giving a "basket-of-eggs" topography. The long axis is parallel to the direction of ice flow, and the upstream end has steeper slopes than the downstream end. They are commonly about a mile long and a quarter of a mile wide, with heights from 50 to 100 ft. Small swamps commonly occur among them. Some drumlins may be rock-cored, ('rock drumlins') and some are loess-covered.

Kettle holes. Stagnant ice masses or detached ice blocks can become covered by moraine and other detritus. When the ice masses melt, bowl-shaped depressions called kettles, or kettle holes, are formed. They are commonly up to half a mile across and 20 to 30 ft deep. Kettles are commonly associated with kames in a very irregular topography on a small scale. They are often found as scenic lakes and some have become swamps from aquatic plant growth.

Erratic boulders. Erratic boulders, or erratics, are large rock fragments lying free on the surface at a considerable distance from the bedrock in which they originated. Their presence can be explained only by invoking transport by glacier ice.

Boulder trains (indicator fans). Erratics of a given rock type can sometimes be traced back to a definite source of origin along a sinuous trail of deposition. These trails of identifiable rocks mark out a flow path of the former glacier, and they are known as boulder trains or indicator fans.

Varved clays. Varved deposits are laminated sediments which show regular periodic variations of particle size, usually an annual alternation. The coarser and thicker (silty) varves are laid down in lakes from vigorously flowing water in the melt season, and the fine (clay), thinner layers are deposited during the winter period, when runoff is greatly reduced. Each cycle gives a thickness of about $\frac{1}{2}$ in. Varves can be counted like tree rings to give rates and datings. Varved clays are best developed in glaciated areas, although rhythmic alternation of sedimentation does occur in recent alluvium in any area. However, glacial varved deposits are much more consolidated and extensive in area and there should be no mistake in identification.

Outwash sediments. Outwash sediments are stratified deposits of water-borne material laid down by melt streams near the margins of glaciers. They range in grain size from small boulders down through gravels to sand, the grain size generally decreasing with distance downstream in any given layer. They provide excellent sources for construction materials (fill and concrete aggregates).

Loess. Loess is a uniform fine-grained soil (clayey silt with fine sand) laid down by the wind after transport from an arid area. Dried-up lake beds, vegetation-free outwash, and till exposed by retreating glaciers provide sources of loose fine-grained material, while glacier winds are capable of transporting dusty sand. Extensive loess

Figure IX-3. Valley glaciers flowing into the sea on the coast of Greenland. Note the long narrow fjords, formed at a time when the glaciers were more extensive.

deposits exist in North America and Eurasia in regions south of the great Pleistocene ice sheets. The material blankets the topography.

Lakes and swamps. Continental glaciation leaves unmistakable evidence in the disruption of the normal drainage system, shown by innumerable lakes and swamps from the Great Lakes to thousands of small ones in North America and Northern Europe.

Supplementing the direct geological indicators of glaciation are data relating to climatic change from biological sources. In addition, there is evidence which can be used to establish a chronology. Among the dating techniques available are botanical methods, such as tree ring, lichen and pollen analyses, archaeological correlations (for fairly recent events), physical methods based on the decay of radioactive isotopes, and geologic deduction from fossils, stratigraphic succession, erosion, and sedimentation (e.g., varve chronology).

Figure IX-4. Mt. Tasman and Mt. Cook, New Zealand, exhibit typical alpine sculpture, with sharp peaks, knife-edged ridges, and bowl-shaped cirques from which glaciers flow.

Figure IX-5. Ice-scraped rock of the Canadian Shield in the region east of Yellowknife. (From ACFEL Tech. Rpt. 41)

Sequence and extent of past glaciations

Three major glacial epochs are recognized by geologists: the Pleistocene, occupying the past million years of the earth's history, the Permian/Carboniferous, which occurred roughly 230 million years ago, and the late Precambrian, which was perhaps 500 million years ago. Of these, the Pleistocene has been most intensively studied.

During the Pleistocene epoch there occurred four well-defined glacial maxima, at which times large ice sheets occupied the north and south polar regions, and extensive mountain glacier systems existed in the temperate zones and even in tropical regions. These glacial periods were separated by interglacial stages, in which climatic conditions were less conducive to glacier growth.

In North America the four main glacial stages are termed the Wisconsin, the Illinoian, the Kansan, and the Nebraskan. Corresponding stages in Europe are known as the Würm, Riss, Mindel, and Günz glacials.

Many attempts have been made to derive a Pleistocene chronology, but the various estimates are not in good agreement. Recent studies, such as analyses of cores from marine sediments, tend to shorten earlier age estimates, which had spread the Pleistocene glaciations over a total period of about one million years.

The last period of extensive glaciation probably culminated some 15,000-20,000 years ago, eventually declining about 10,000 years ago with a rapid retreat of the ice. Following the general recession of glaciers came a period during which temperatures are believed to have been higher than those prevailing today. This warm period, the Climatic Optimum, ran from perhaps 5000-2000 B.C. There followed a moderate re-advance of glaciers (the "Little Ice Age"), maximum extensions being reached in the eighteenth and nineteenth centuries. Since the middle of the nineteenth century, most glaciers have been receding, although there have been a few advances from time to time. The past one hundred years of general glacier retreat is too short a time to decide whether the present situation represents the onset of another major recession; direct and indirect evidence of glacier fluctuations in recent centuries show clearly that minor, short-period variations are superimposed on the long-term trends.

The ice front advanced and retreated so that it is often impossible to determine how many glacial and interglacial periods actually occurred in a given region.

It is estimated that at the last glacial maximum over 15,000,000 square miles of the earth's surface were covered by glacier ice, an area equivalent to about 28% of the present land surface. The ice sheets of Antarctica and Greenland were more extensive than now, and major ice sheets, up to 10,000 ft thick, lay on North America and northern Europe. Smaller ice sheets overlay Iceland, Spitsbergen and Franz Josef Land, and mountain glaciation was extensive in Siberia, the Himalayas, the Caucasus, the Alps and Pyrenees, the Andes, and in New Zealand. Mountain glaciers existed in the high mountains of Japan, Hawaii, East Africa, New Guinea, and Tasmania.

Canada and the northern U. S. were covered by the Laurentide Ice Sheet, which extended from the eastern seaboard across the continent to unite with the glacier complex lying over the western mountains. From the Arctic Ocean in the north, it ran down as far south as New York City, Cincinnati, and St. Louis (leaving behind the valuable sand and gravel deposits lacking in the south). Large parts of Alaska apparently remained unglaciated.

Causes of ice ages

To the question "What causes ice ages?" there is a direct answer: "Nobody knows". There is no shortage of hypotheses, however, for since the mid-nineteenth century the problem has intrigued reputable scientists from many disciplines, in addition to a variety of unconventional theorists.

The numerous suggestions advanced can be grouped basically under the following headings, although individual theories tend to combine elements from several groups.

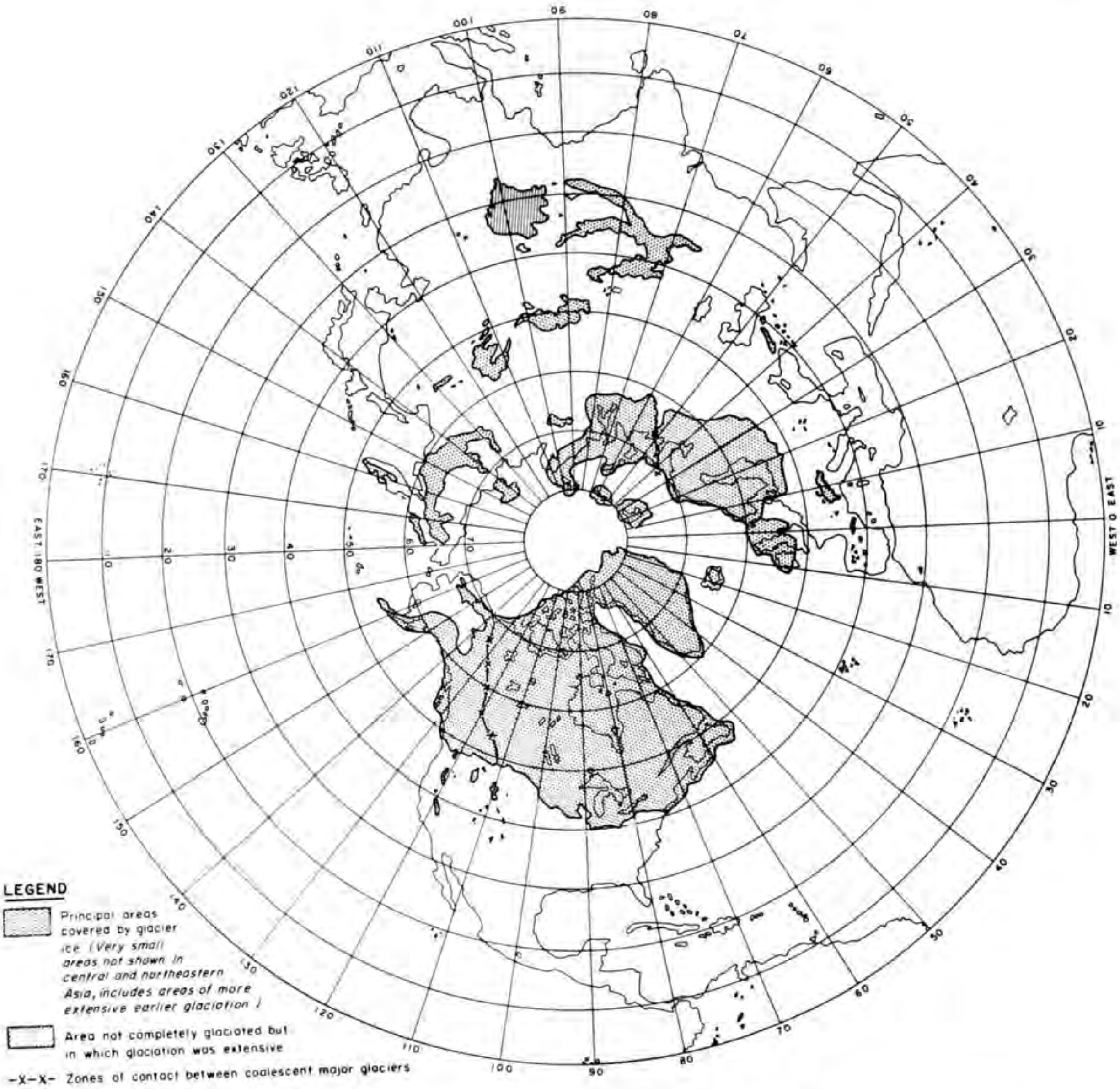

Figure IX-6. Extent of northern hemisphere glaciers at the last maximum. (From various sources)

Solar luminosity. Since the sun is the prime energy source for terrestrial processes, fluctuations in intensity and type of solar radiation may be expected to influence climate. Changes due to variation of the rate of internal energy release, and to interception of interstellar hydrogen clouds have been considered. Fluctuations of intensity of the ultraviolet band of the solar spectrum have also been cited.

Orbital disturbance. The earth's orbit is subject to periodic variations, which affect the distribution of solar radiation received at the earth's surface. This should influence climate.

Atmospheric absorption and reflection. The radiation balance at the earth's surface will be affected by changes in the constituents of the atmosphere (where radiation is selectively absorbed and reflected). Variations in the amount of water vapor, carbon dioxide, ozone, volcanic dust, and cosmic dust have been put forward to explain climatic change.

Polar displacement and continental drift. Climatic disturbance would result from displacement of land masses relative to the geographic poles. Movement of the entire earth's crust relative to the axis of spin, and rearrangement of the continents have been put forward as causes of climatic change.

Continental uplift. High land stimulates the conditions of low temperature and high precipitation which favor glacier growth. The waxing-and-waning of ice sheets has been attributed to uplift and depression of continental land masses, and the attendant climatic disturbances.

Oceanic factors. The oceans exert a strong influence on glacier regime by supplying the water for snowfall and by their thermal effects. Redirection of warm and cold currents, rise and fall of sea level, and covering of seas by ice have been adduced in the discussion of major glacier fluctuations.

Any hypothesis brought forward must be tested against the known facts of geology and biology pertinent to climatic change and ice inundation. None of the above suggestions, nor the ingenious combinations of them, fully satisfy the tests. Some are quantitatively inadequate, some fail to satisfy established periodicity and chronology, while others do not allow for synchronous glaciation in both hemispheres. Of currently popular hypotheses, one of the more convincingly presented is Flint's solar-topographic concept, which envisages a combination of major uplift of highlands and solar variation as the cause of glacial fluctuation.[4] Another idea which has attracted considerable interest in recent years is that of Ewing and Donn.[3] They postulate a combination of polar displacement and ocean change to explain Arctic events, but do not treat the Antarctic problem.

Final solution of the problem must await new facts and fresh insight; in the meantime it remains an entertaining topic for the exponents of unconventional science.

BIBLIOGRAPHY

1. Ahlmann, H. W. (1953) Glacier variations and climatic fluctuations, American Geographical Society, Bowman Memorial Lecture, no. 3.

2. Charlesworth, J. K. (1957) The Quaternary Era. London: Arnold (2 vols.).

3. Ewing, M. and Donn, W. L. (1956; 1958) Theory of ice ages, Science, vol. 123; vol. 127.

4. Flint, R. F. (1957) Glacial and Pleistocene geology. New York: Wiley.

5. Hollin, J. T. (1962) On the glacial history of Antarctica, Journal of Glaciology, vol. 4, no. 32.

6. Lewis, W. V. (1960) Norwegian cirque glaciers, Royal Geographical Society (London), Research Series, no. 4.

7. Mercer, J. H. (1962) Glacier variations in the Antarctic, American Geographical Society, Glaciological Notes, no. 11.

8. U. S. Army Corps of Engineers (1963) Engineering and design-terrain evaluation in arctic and subarctic regions, EM 1110-1-377.

CHAPTER X. SEA ICE.*

Ever since "congealed seas" astonished Pytheas and his fellow voyagers from the Mediterranean in the 4th century B. C., sea ice has exerted a powerful influence on life and trade in the North, closing important ports in the winter, blocking transpolar navigation, and restricting access to many Arctic coasts. In spite of this, surprisingly little organized research has been devoted to details of the behavior and properties of sea ice.

General behavior

Each winter the sea freezes along the coasts of eastern and northern Canada, Alaska, northern USSR, Greenland, and in the Baltic, and even the Caspian and Black Seas. Freezing occurs around all the shores of the Antarctic continent, and along the coasts of the Kamchatka Peninsula and parts of Hokkaido, Japan. Detailed records of ice distribution are now available in a number of sea ice atlases.

Coastal ice formations are relatively stable, since bays and headlands provide shelter and anchorage, but farther out to sea the ice tends to break up and disperse repeatedly under the action of wind and waves. Sheltered and indented coasts favor stable ice growth, whereas steady growth is inhibited in places subject to severe storms, persistent winds, strong or warm currents, or outflow from rivers.

Freeze-up occurs in autumn or early winter, and unbroken ice thickens gradually through the winter and early spring. With the return of strong, prolonged sunshine in spring, heat loss from the ice surface ceases and growth ends. With higher air temperatures and stronger solar radiation, heat flow through the ice is reversed and the decay process begins, with ablation on the surface and melting at the base of the ice. Finally the ice weakened by melting breaks up under the action of wind or swell, and discrete slabs (floes) drift out as pack ice.

Pack ice drifts with wind and current until dispersed in the open ocean or trapped in some constricted area. The ice which moved freely out to warmer seas soon breaks up and melts away, but floes which are unable to escape are eventually frozen together again in a matrix of new ice. In simple situations the old floes form flat mosaics cemented by new ice, but quite commonly horizontal pressure builds up in certain zones, and the floes are thrust over each other to form rafted ice (sometimes described by the apt term "tombstone country"). In other cases the floes buckle and break under horizontal thrust, forming pressure ridges.

A thin sheet of newly formed ice is quite flexible, and can sometimes be seen rippling in response to ocean waves and swells beneath. Under the action of horizontal force, e.g., the effect of wind shear, thin new ice may overthrust, broad fingers of ice driving alternately over and under the ice of the resisting section, displaying a characteristic "square-wave" pattern when viewed from above.[9,16] All sea ice which is not grounded rises and falls with the tides. Where sea ice abuts the land, belts of undisturbed ice are formed as a result of repeated shearing and thrusting of the floating ice. Simple cracking may occur out on the flat expanses of the fast ice as a result of thermal expansion and contraction. Compared with fresh ice such cracks are rare. Even thick fast ice will rise and fall over short time periods when ocean swells penetrate from the outer limit of the ice; an engineer's level set up on shore and sighted onto a rod held on the ice sometimes shows these relatively high frequency oscillations superimposed on the tidal motion.

*A comprehensive and detailed treatment of sea ice will be published separately in this series.

Drift of pack ice in the Arctic basin is governed by surface currents, which follow a general clockwise circulation in the area between Siberia and the north coasts of Alaska, Canada, and Greenland.[14] On the fringes of the main clockwise circulation are a number of smaller counterclockwise eddies (Fig. X-2). Inflow of Atlantic water is primarily from the eastern Atlantic in the vicinity of Norway. Outflow to the Atlantic is mainly on the western side, largely from the East Greenland current. Because the Arctic basin is largely landlocked, only a small proportion of the pack ice escapes. Old ice which has been repeatedly refrozen and rafted is typical of the region.

At the end of winter, say in the months of March and April, the entire Arctic basin is ice-covered with the exception of leads caused by wind stresses. Average ice margins extend across the Bering Strait, from Labrador to West Greenland, and from the southern tip of Greenland over to West Spitsbergen. Part of the coast of southwest Greenland and all of the coasts of Iceland are substantially ice-free. From West Spitsbergen the edge of the ice runs southeast to meet the Russian coast in the White Sea area, northern Norway and Finland being outside the ice-bound region.[14]

In the Antarctic the situation is quite different, since the polar continent is surrounded by open ocean. Surface currents near the coast move in a general counterclockwise direction along the coastlines, i.e., from east to west.[14] The pack ice is not confined, and so it tends to spiral northwards, away from the continent and into warmer seas. There are, of course, local departures from this general trend, one notable example being the concentration of ice in the Weddell Sea, where circumpolar drift is blocked by the Antarctic Peninsula.

At the end of winter, say in the month of September, the belt of sea ice surrounding Antarctica is very wide. In some sectors ice extends as far as 1000 miles out from the coast. During the summer much of this ice drifts away and disappears. By March the belt of pack ice is no wider than about 100 miles along much of the coast of East Antarctica. Ice is more extensive around West Antarctica, however, the great embayments of the Ross and Weddell Seas accounting for much of the total.

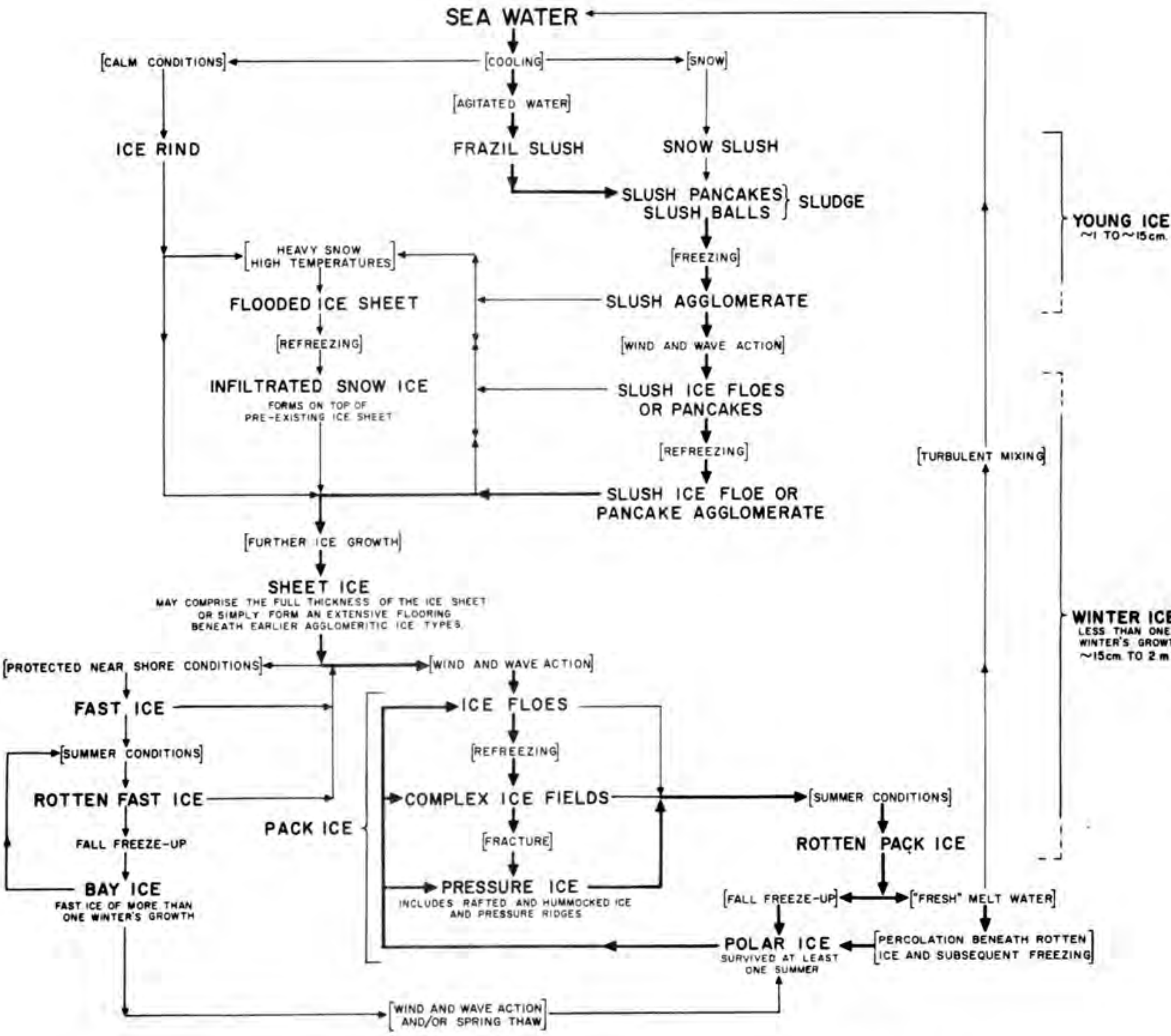

Figure X-1. Sea ice classification scheme drawn up by W. F. Weeks of CRREL. The diagram gives sea ice terminology and shows the interrelationship of various ice forms.

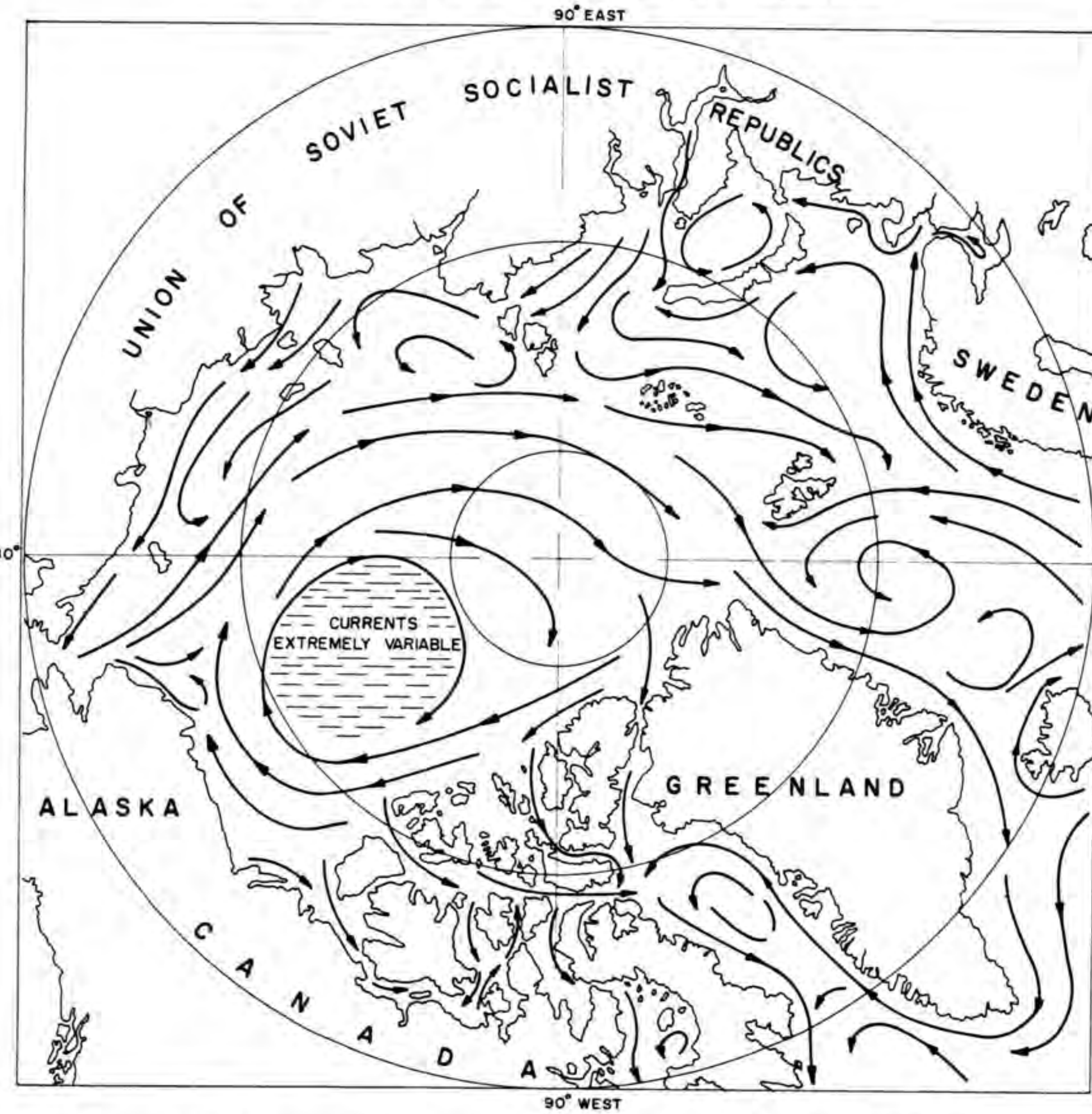

Figure X-2. General circulation of surface waters in the Arctic. (From ref. 14)

Sea-ice terminology

Many sea-ice terms were coined at a time when descriptions of different ice forms were largely a matter of personal preference, but with the growth of international communication a need for universally acceptable terminology has arisen. There are now available several widely circulated ice glossaries,[21,22,23] but complete international agreement has not yet been reached. Many of the quaint terms traditionally listed can be omitted here, but a few definitions are helpful in reading the literature on sea ice.

Bay ice is the name given to sea ice which has remained intact at its place of origin for more than one year. Having more than one winter's growth, it is thicker than annual ice, and it may have received additional nourishment from accumulation of snow on the surface. An arbitrary upper limit of thickness is placed at about 2 m above sea level; ice thicker than this is called an ice shelf (see Ch. II, p. 23).

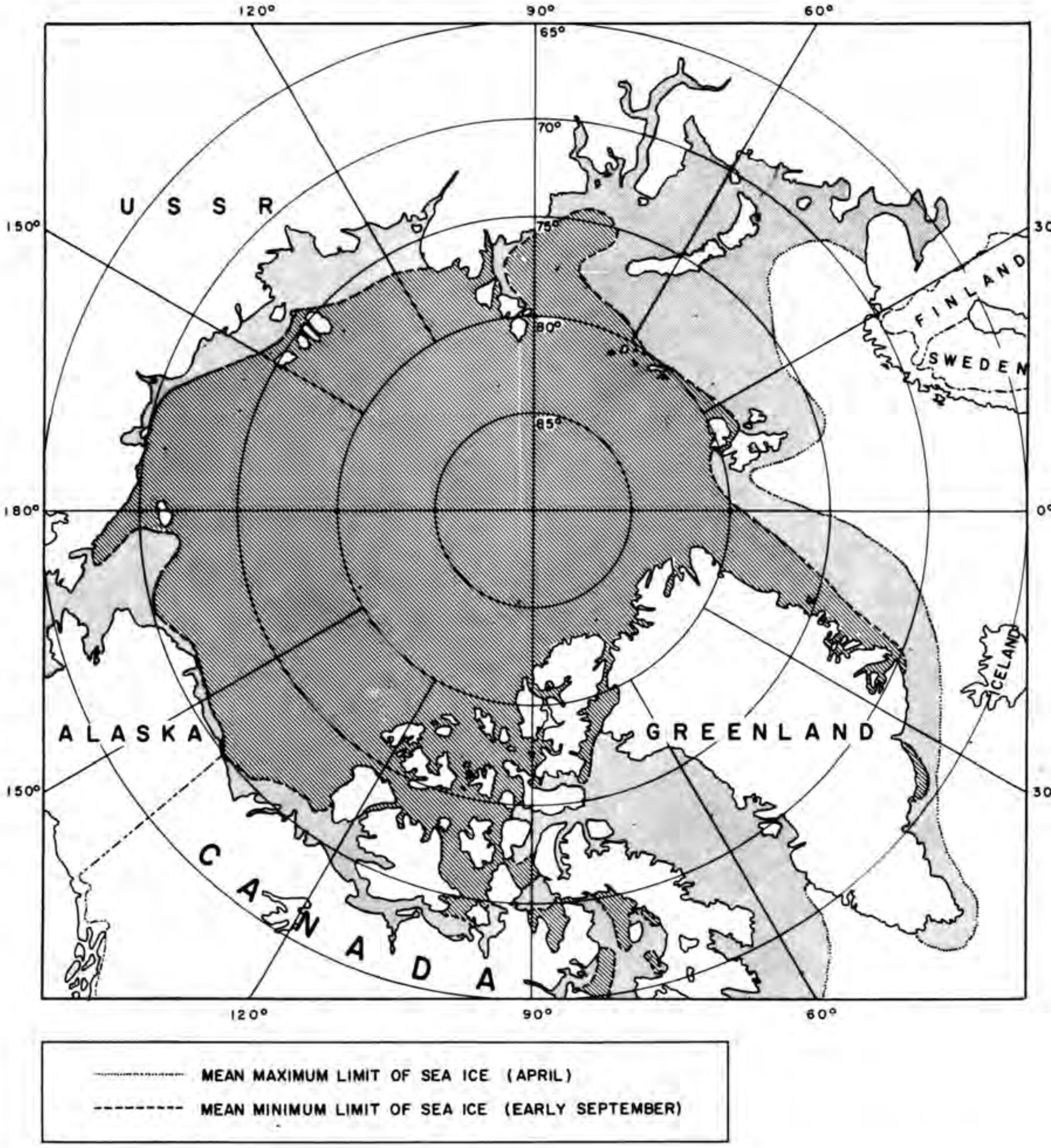

Figure X-3. Mean maximum and minimum limits of sea ice in the Arctic. (From ref. 14)

<u>Brash ice</u> is an accumulation of small fragments of floating ice. The fragments may be remnants of true sea ice or pieces of glacier ice. Two meters is set as the upper limit of size for fragments in this category.

<u>Fast ice</u> is an intact expanse of sea ice which remains anchored to land or to glacier ice. Fast ice usually breaks up in the summer; if it does not, the term <u>bay ice</u> is applicable.

<u>Floe</u>. A floe, or ice floe, is a substantial piece of sea ice which is subject to drifting under the action of wind and current. It may be flat or <u>hummocked</u>.

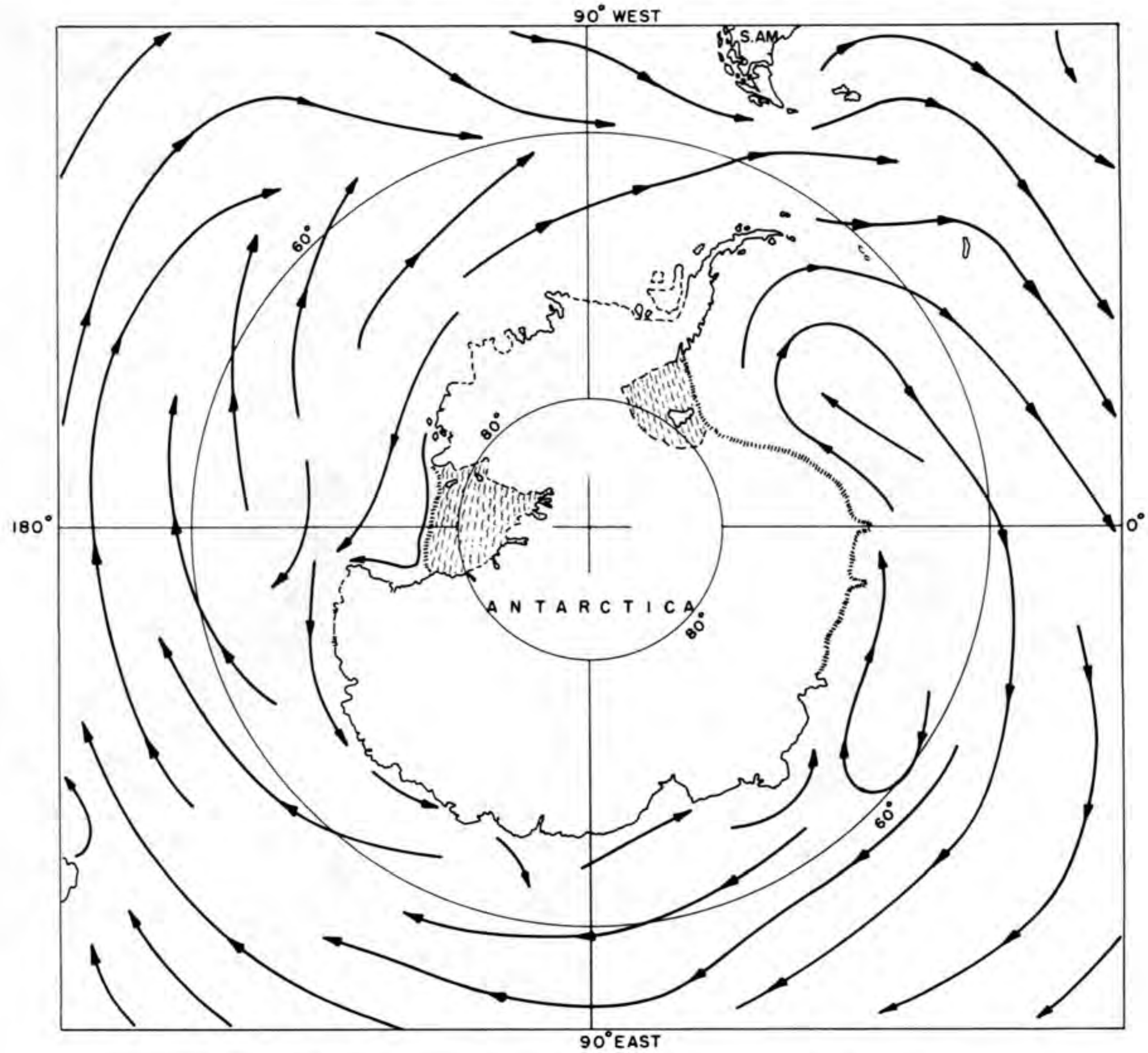

Figure X-4. General circulation of surface water in the Antarctic. (From ref. 14)

Frazil is an aggregation of fine spicular or platelike crystals floating on the surface of the sea. It marks the first stage of the freezing process.

Hummocked ice consists of floes, and fragments of floes, piled up haphazardly by horizontal pressure. It may have been modified by melting and snow accumulation.

Ice cover is the relative amount of sea ice present in a given area. It is estimated in eighths (formerly tenths) of the surface covered by ice.

Ice edge. This is the boundary between sea ice and the open sea at any given time and place.

Ice field is the name given to an expanse of pack ice which extends beyond the limits of visibility.

Lead (or lane). A lead is a navigable passage through pack ice. It is considerably wider than a crack.

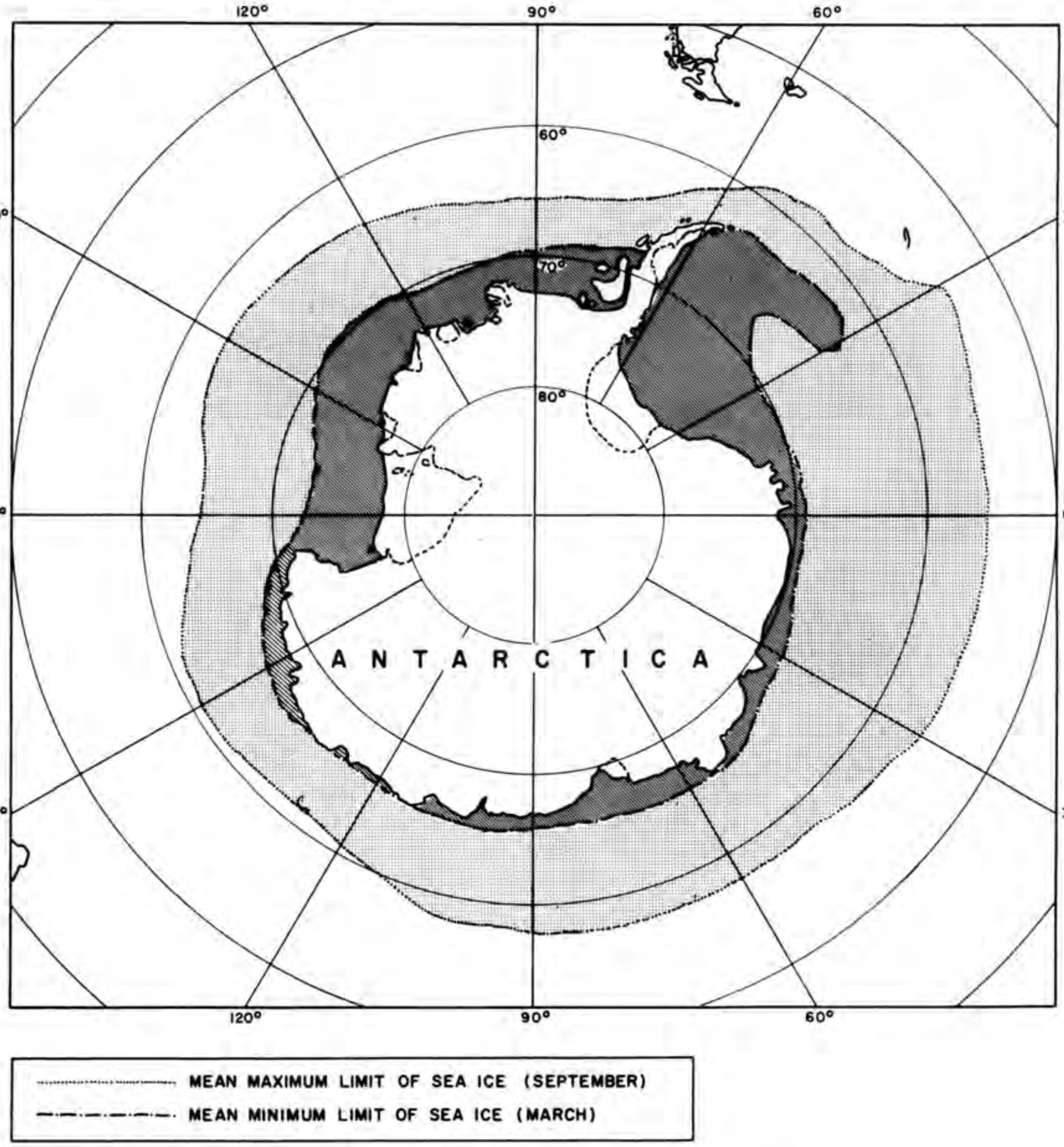

Figure X-5. Mean maximum and minimum limits of sea ice in the Antarctic. (From ref. 14)

Pack ice consists of numerous floes in appreciable concentration (say covering more than one-eighth of the visible sea surface). The floes may be free-floating or jammed tightly together by wind or current.

Pancake ice is an accumulation of soft, newly formed ice in which discrete pieces take on a circular shape and develop raised rims as a result of repeated collisions with their neighbors.

Polynya. A polynya is a large area of open water surrounded by sea ice, which persists in the same position year after year. Polynyas may be maintained by outflow from a river, by persistent local winds, or by unusual currents.

Pool. A pool is any area of open water, surrounded by sea ice, of obviously limited extent (cf. "puddle").

Pressure ridges are strips of ice which have been forced into ridges and hummocks by horizontal thrusting. Shattered floes are up-ended and stacked one upon another along the ridges.

Puddles are water-filled depressions on the surfaces of floes. The water comes from melted snow or surface ice and is often salt-free and potable.

Rafted ice is ice which has been subjected to moderate horizontal thrusting, resulting in floes overriding each other and stacking, while remaining more or less horizontal. The process is called rafting.

Rotten ice is ice in an advanced state of decay by melting. It is soft and honeycombed as a result of preferential melting at the walls of brine pockets.

Shore lead. This is a strip of open water separating the pack ice from the shore. It is often caused by offshore winds.

Sludge is a surface concentration of frazil. It appears as a thick soupy layer of ice crystals and water.

Tide cracks are the fractured shear zones at the junction between floating sea ice and immovable land or grounded ice. Repeated rise and fall of the sea ice, together with horizontal thrust, leaves ridges of disturbed ice along the tide cracks.

Ice formation

When fresh water freezes, ice can form at the surface while the water beneath it is still at temperatures above the freezing point. This is because pure water reaches its maximum density at +4C, and expands on cooling below that temperature; therefore a stable thermal stratification is possible in fresh water cooled from the surface after general water temperatures have dropped to +4C.

Sea water behaves differently: for salinities in excess of 24.7‰* (normal sea water is about 35‰), density increases as the temperature is lowered to the freezing point, and therefore water cooled at the surface tends to sink. This means that, unless surface cooling is extremely rapid, ice cannot form until the entire depth of water (or at least a great depth) is isothermal at the freezing point.† This is the main reason why sea water freezes later in the season than nearby freshwater lakes.

When salinities are less than 24.7‰, density currents are not set up, as this water behaves in a similar manner to fresh water. This critical salinity of 24.7‰, which can be regarded as separating brackish water from true sea water, is the salinity at which the temperature of maximum density (the inversion temperature) coincides with the freezing temperature. Both inversion temperature and freezing temperature are approximately linearly related to salinity, and when plotted these functions cross at 24.7‰ salinity and -1.3C (Fig. X-6).

Surface waters of the polar seas commonly have salinities in the range 30-35‰, and freezing begins when the water temperature drops to -1.8 or -1.9C. The first ice to appear is a surface layer of thin, delicate crystals which float freely, being unattached to their neighbors. They usually have a platelike shape, up to 2 cm or so across and less than a millimeter thick, but spicular and dendritic forms also occur. These initial crystals consist of pure ice, i.e., they contain virtually no impurities. The salts rejected during their freezing serve to concentrate the brine surrounding them.

Agitation of the initial crystals (often called frazil) by wind and waves prevents a coherent ice cover from forming until an appreciable concentration of the crystals has been produced. With the onset of calm weather conditions, the frazil crystals, which

*‰ = parts per thousand, by total weight.

† This statement is simplified for clarity. In nature the situation is complicated by effects of currents, turbulent exchange, and compressibility of the water.

are already loosely matted together, freeze into a continuous ice cover. If swells continue to disturb the ice it may break into roughly circular pieces, which bump against each other and develop raised rims; this is known as pancake ice.

As the frazil crystals freeze together, brine is trapped in the ice mass. The amount and concentration of brine caught this way, and hence the overall salinity of the ice, is dependent on the freezing rate: the more rapid the freeze, the greater the resulting salinity.[8] Ice frozen very rapidly may have up to 20‰ salinity, while ice frozen very slowly may have only 4‰.

When the platelike frazil crystals are first formed they tend to float flat upon the surface, and since the principal optic axis, or c-axis, is perpendicular to the plane of the disk, that axis is generally near the vertical.[15] Subsequent jostling and matting forces some of the crystals together, so that their c-axes tilt toward the horizontal. Selective growth of crystals having near-horizontal c-axes results in eventual development of a preferred horizontal orientation. This has been attributed to anisotropic thermal conductivity and/or to favored growth planes in the crystal lattice.

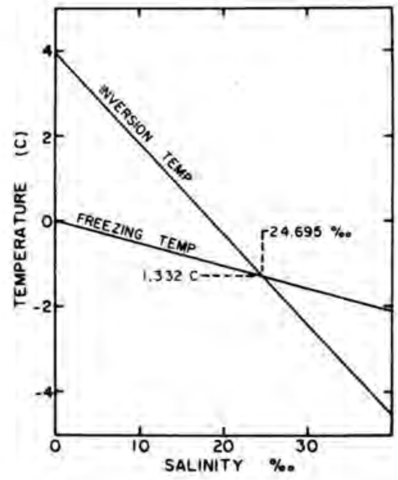

Figure X-6. Relationships between freezing temperature and salinity, and inversion temperature and salinity. (After Defant, ref. 5)

The inclusions of brine and air in sea ice form discrete cells along selected planes within the crystals and at the crystal boundaries. The cells are usually elongated in a vertical direction. The brine inside a cell tends to preserve an equilibrium as temperature changes occur. If the temperature falls, some of the brine freezes at the edge of the pocket, until the salts rejected by freezing raise the salinity of the residual brine to a point where it is in equilibrium with ice at that temperature. If the temperature rises, the converse process takes place. This effect permits brine to diffuse through ice subjected to a temperature gradient, since the brine cells are usually not in equilibrium. During winter the underside of sea ice, which remains at the freezing point, is warmer than the top surface, and brine migrates downwards, even though the ice is much colder than the freezing temperature of sea water.

In addition to this freshening of the ice by brine diffusion along a temperature gradient, gravity drainage of the brine occurs in summer when higher temperatures permit the brine cells to enlarge and interconnect. Ice which survives a summer thus may become almost completely fresh, particularly if melt water from surface snow drains through and flushes out the pockets.

As the temperature of new sea ice falls, concentration of entrapped brine increases, and abrupt changes occur at certain temperatures owing to precipitation of salts. The main dissolved salts in sea water are sodium chloride, sodium sulphate, and magnesium chloride; when they are mixed in the ratios usual for sea water, freezing and cooling cause sodium sulphate decahydrate ($Na_2SO_4 . 10H_2O$) to precipitate at about -8C, and sodium chloride dihydrate ($NaCl . 2H_2O$) at about -23C. This leads to sharp changes in physical properties of the ice at these temperatures.[1] Tables of composition and brine volumes are available.[19]

From the foregoing it will readily be understood that the physical properties of sea ice are profoundly influenced by temperature and salinity (Fig. X-7-11). The strength of sea ice of a given thickness therefore depends very much on temperature and salinity, unlike lake ice, where the salinity factor is absent and temperature is of less importance.[19,20] Thus, in considering such practical problems as bearing strength (for landing aircraft or running vehicles on ice) and ice breaking (for navigation), temperature and salinity are of paramount importance. For example, a sheet of sea ice 2 ft thick will be very much stronger during the cold winter period than a sheet of the same thickness in the height of summer.

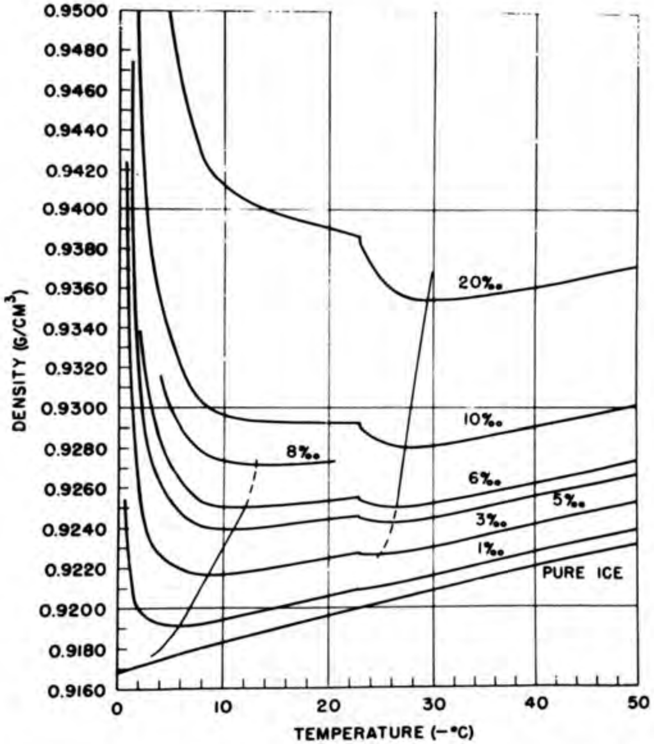

Figure X-7. Density of sea ice as a function of temperature, with salinity as parameter. (After Anderson, ref. 1)

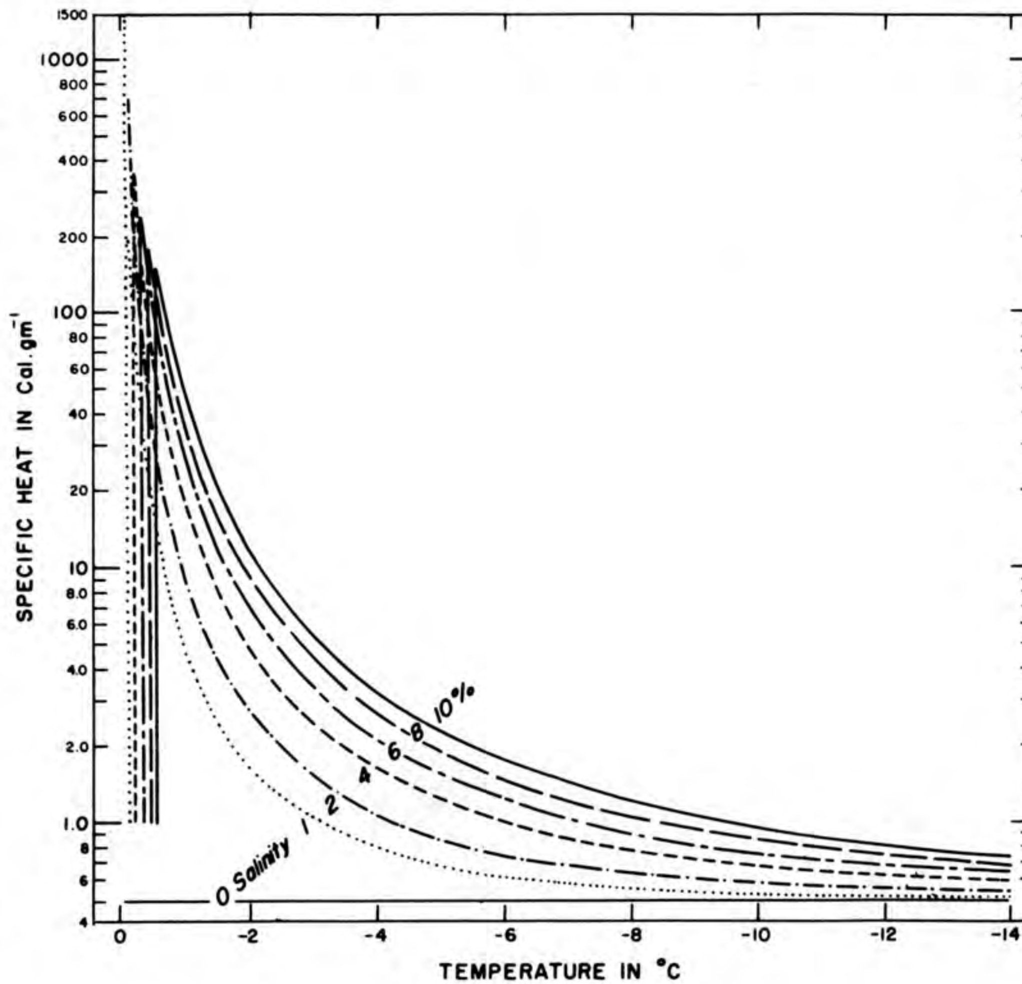

Figure X-8. Apparent specific heat of sea ice as a function of temperature, with salinity as parameter. (After Schwerdtfeger, ref. 18)

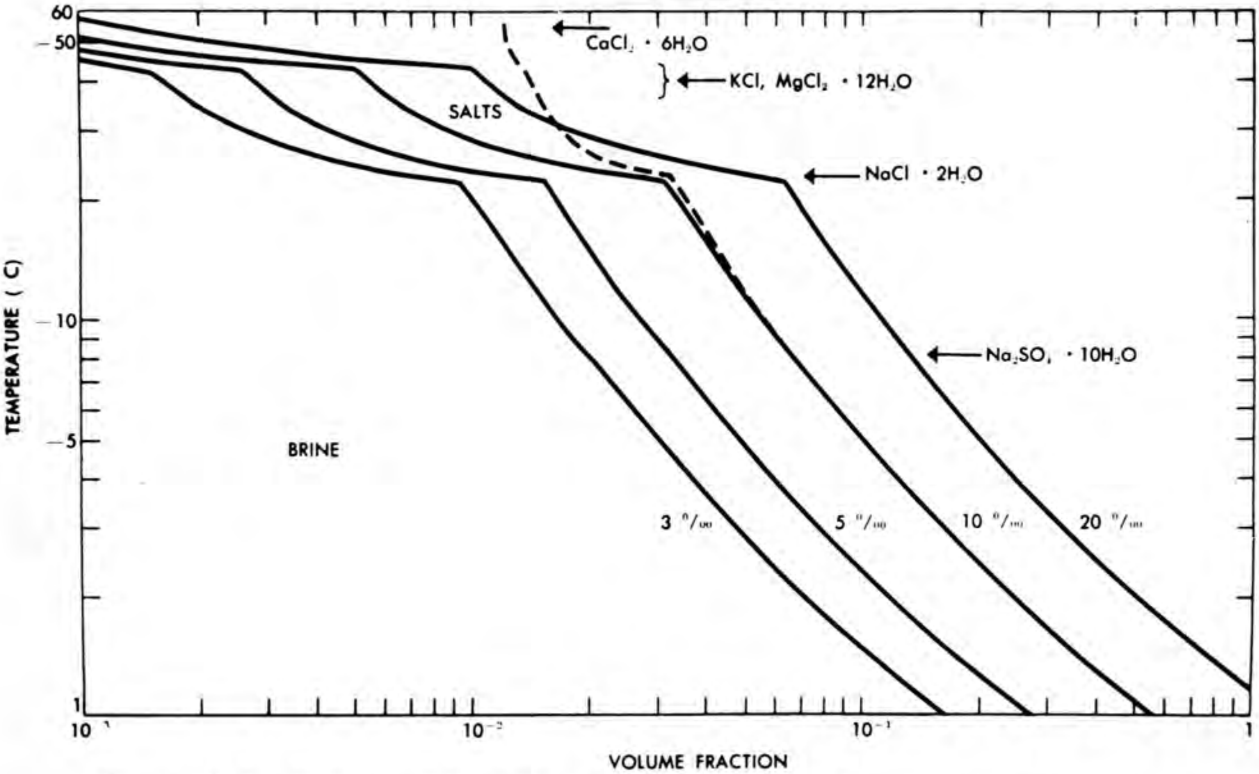

Figure X-9. Volume of liquid brine in sea ice as a function of temperature, with salinity as parameter. (After Anderson, ref. 1)

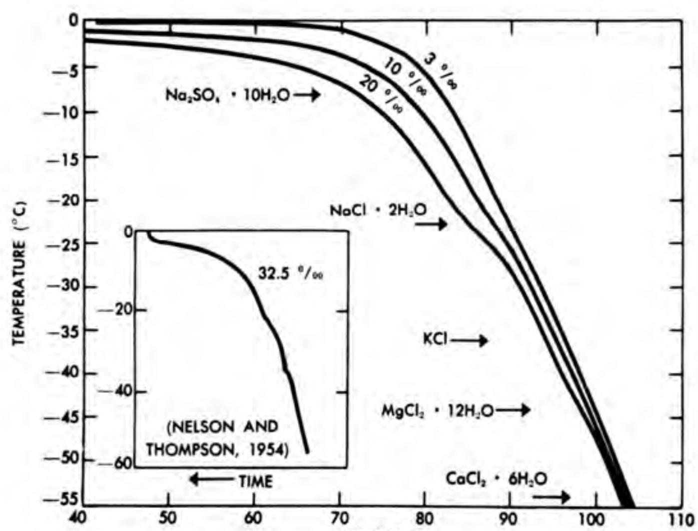

Figure X-10. Heat of melting as a function of temperature. Insert shows warming curve versus time. (After Anderson, ref. 1)

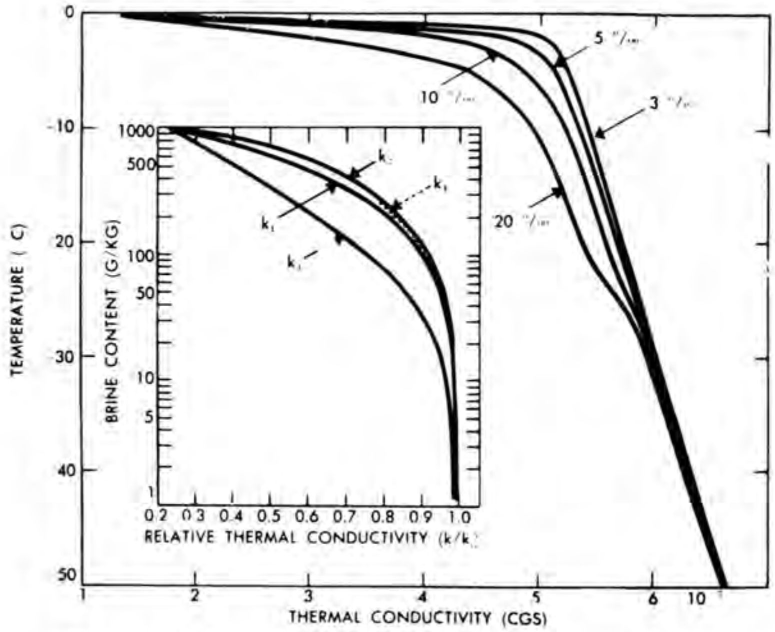

Figure X-11. Thermal conductivity as a function of temperature. (After Anderson, ref. 1)

Predicting time of initial freezing

The practical problem of predicting the date of first ice formation at a given locality is dealt with by estimating heat loss from the water surface, using air temperatures forecast from climatological records. In the ice potential method pioneered by Zubov,[17] the vertical distributions of temperature and salinity are obtained from ocean soundings, and heat transfer is analyzed using an appropriate model. From a knowledge of the total heat of a water column, the rate of heat transfer to the surface, and the rate of energy exchange between the water surface and the air, the date of first ice formation can be calculated. A simpler, more empirical technique is described by Rodhe[11]. In this method, rate of heat loss from the sea is assumed proportional to the difference between air temperature and sea surface temperature, and freeze-up date is predicted by forecasting air temperatures and making day-by-day calculations of heat loss until the water temperature is reduced to the freezing point. Records of previous years are used in trial solutions in order to determine the constant of the heat loss equation for a given locality.

Sea-ice growth

Once a coherent ice cover has formed, the rate of growth is governed by heat flow through the ice. This, in turn, depends upon the temperature gradient and the thermal conductivity of the ice.

The temperature gradient (roughly speaking, the temperature difference between the upper and lower surfaces divided by the thickness) is influenced by several factors. The temperature of the upper surface is controlled by heat conduction through the ice, convective heat transfer from the air above, the balance between incoming and outgoing radiation, and condensation or evaporation. The temperature of the lower surface is a constant "ice-bath" temperature of about -1.9C; heat may be supplied by turbulent transfer from the water, however, reducing the demands on latent heat of fusion and hence retarding ice growth. As the ice thickens, of course, a given temperature difference across the ice will give a diminishing temperature gradient, leading to reduction of growth rate with time.

To further complicate the matter, thermal conductivity of sea ice is sensitive to both the temperature and salinity, since the response of the included brine cells to temperature variations may cause release or absorption of latent heat throughout the thickness of the ice.

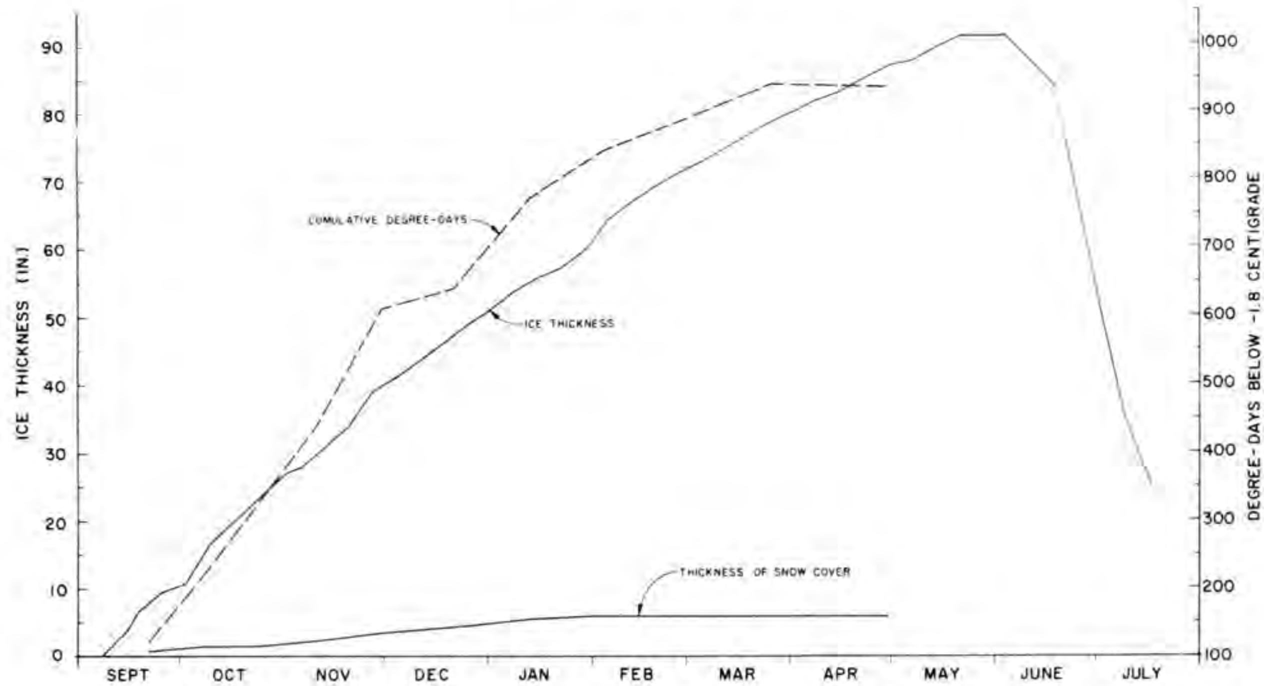

Figure X-12. Growth and decay of sea ice at a station in the Canadian Arctic. (Eureka, N. W. T.). (From records compiled by M. Bilello, ref. 3)

Finally, snow cover upon the ice adds another complication: it is a layer with quite different thermal properties and with characteristics which alter the surface heat balance (higher albedo, affecting radiation; surface roughness, affecting convection; and low thermal conductivity).

At the end of the nineteenth century the growth rate problem was analyzed mathematically with simplifying assumptions by Stefan, whose solution subsequently yielded the expression:

$$h_t = A\sqrt{\Sigma\,\theta}$$

where h_t is ice thickness at time t after freeze-over, A is a constant dependent on the physical properties of the ice and $\Sigma\,\theta$ is a freezing index found by summing the daily average temperatures below freezing point from freeze-over to time t (temperatures being those of the ice surface).

While this is a satisfactory solution for an ideal model, application to the prediction of sea-ice growth is restricted by the variability of thermal properties mentioned above, the simplifying assumptions, and the difficulty of determining true surface temperatures of the ice. Many attempts at refinement have been made, but the results have been of little practical value because of their cumbersome form, and because of the lack of detailed observational data. A preferred approach has been to relate ice thickness to accumulated sub-freezing air temperatures ("degree-days of frost"), deriving a simple power law ($y = ax^n$) in which the coefficient and the exponent are found empirically. Some examples of published results are given below.[2,3]

Assur (given for fresh-water ice, but held approximately true for sea ice)

$$H = 1.06\,K\,\sqrt{S}$$

Bilello (Canadian Arctic — no snow cover)

$$h = 3.55\,\sqrt{\Sigma\theta} \quad \text{(Fig. X-14, curve A)}$$

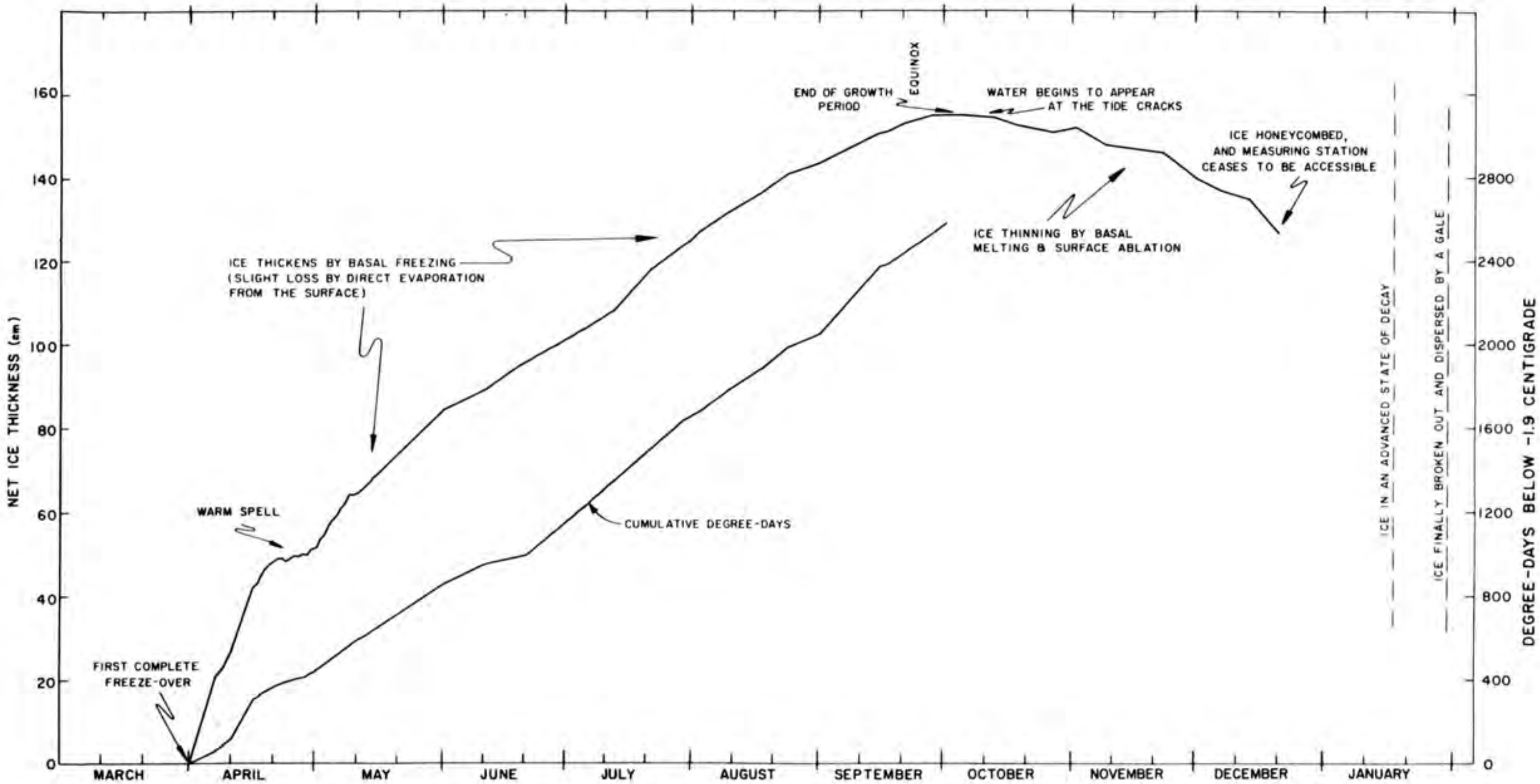

Figure X-13. Growth and decay of snow-free sea ice at a station on the coast of East Antarctica. (After M. Mellor, ref. 9)

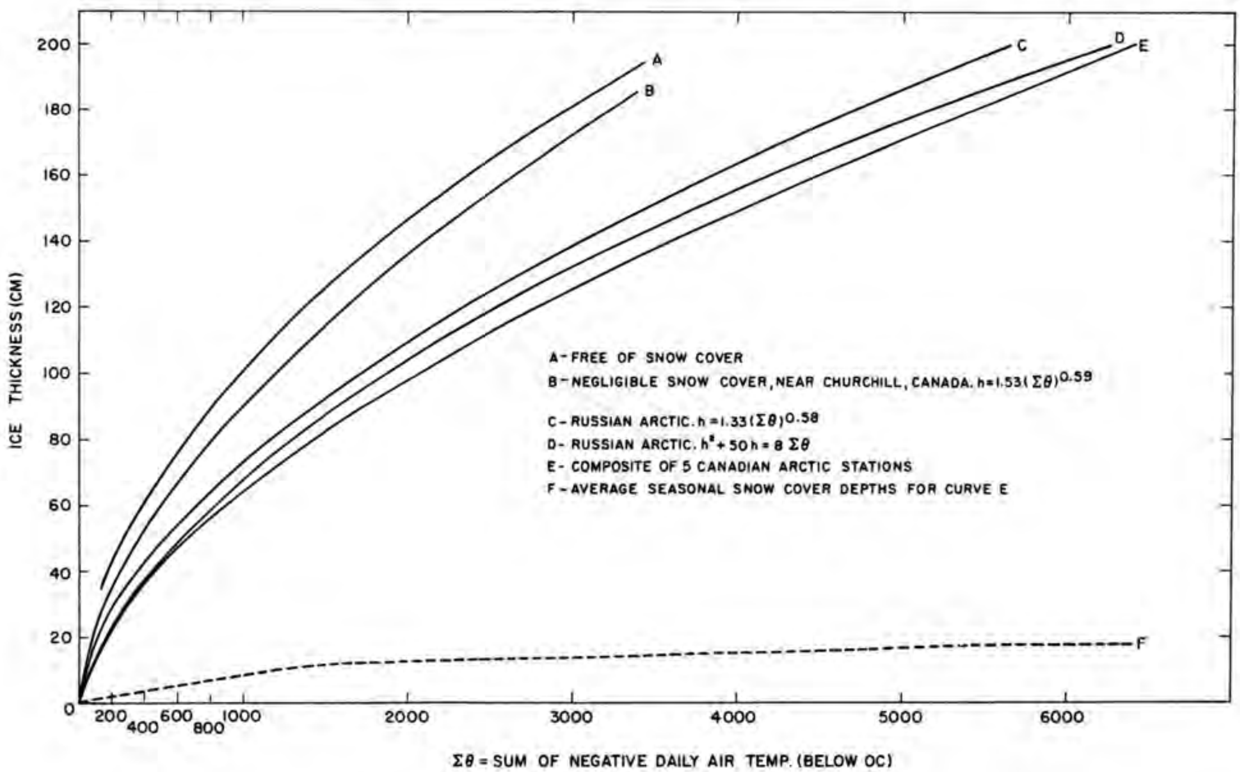

Figure X-14. Summary of empirical relationships between ice thickness and accumulated degree-days of frost. (After Bilello, ref. 3)

<u>Graystone</u> (Churchill, Canada — negligible snow cover)

$h = 1.53\,(\Sigma\theta)^{0.59}$ (Fig. X-14, curve B)

<u>Lebedev</u> (Russian Arctic — average snow cover)

$h = 1.33\,(\Sigma\theta)^{0.58}$ (Fig. X-14, curve C)

<u>Zubov</u> (Russian Arctic)

$h^2 + 50h = 8\Sigma\theta$ (attempting to allow for heat capacity) (Fig. X-14, curve D)

where:

H = ice thickness (in.)
h = ice thickness (cm)
S = summation of average daily freezing temperatures (Fahrenheit degree-days) — below 32F.
$\Sigma\theta$ = summation of average daily freezing temperatures (degree-days) — below 0 C for Graystone, Lebedev, and Zubov; below -1.8C (approximate freezing point of sea water) for Bilello.
K = coefficient accounting for snow cover and local conditions, about 0.70-0.75 for Arctic sea ice.[2]

These expressions are only empirical relationships between air temperature and growth for definite areas; they are not universally applicable, and generally best represent the growth of fairly thick sea ice. Initial growth rates usually do not conform to the equations. Bilello[3] gives equations for the effect of snow cover.

Coastlines in the Arctic and Antarctic accumulate from 3000 to 7000 Celsius degree-days of freezing exposure during the course of a winter, resulting in maximum ice thicknesses of 1.5 to 2.5 m (varying mainly with snow cover) for a single winter's growth.

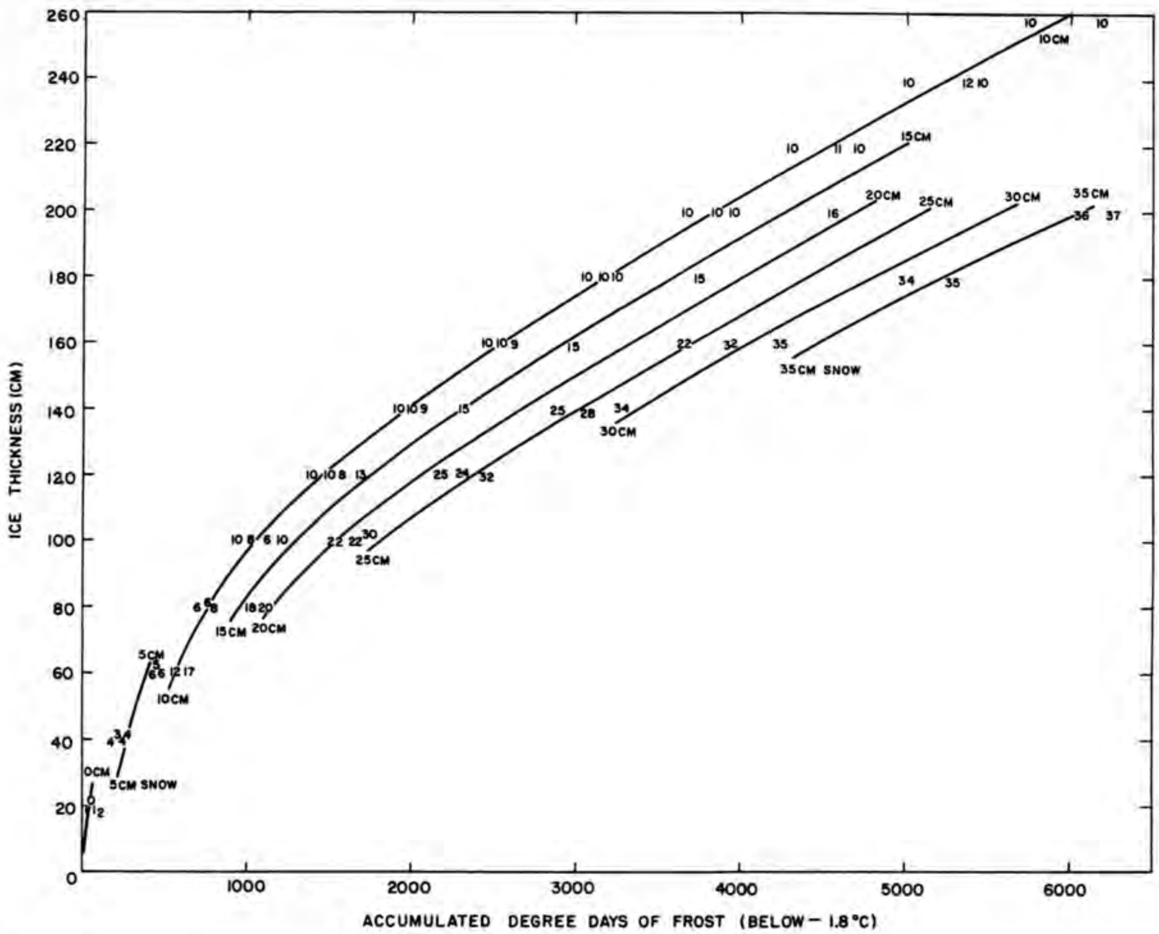

Figure X-15. Relationship between ice thickness and accumulated degree-days of frost, showing the influence of snow cover. The numbers give snow cover in cm. (After Bilello, ref. 3)

Melting and refreezing

In most areas of the Arctic and Antarctic, growth of sea ice ceases in spring when solar radiation becomes strong. At this time, air temperatures may still be well below the melting point, but the ice surface is warmed by radiation and thus heat flow through the ice is greatly reduced.

The effectiveness of solar radiation in warming the ice is strongly dependent on the albedo, or reflectivity, of the surface. When covered by clean snow the albedo may be as high as 0.9, and thus only 10% of the incident radiation is absorbed. Snow-free ice, however, commonly has an albedo of around 0.5, so that 50% of the incoming radiation can be absorbed.

As the season advances, convective heating from the air becomes significant, and heat may also be supplied to the base of the ice by water circulation.

Warming of the ice, even while temperatures generally are well below the normal freezing point of sea water, permits internal melting to take place, so that the temperature-salinity equilibrium of the included brine cells may be restored. The volume of liquid brine increases as the temperature rises, until eventually adjacent cells unite and the ice becomes permeable. This allows the brine to drain out by gravity, leaving the ice purer and more resistant to melting.

The rate of thinning has been related empirically with air temperature; Bilello and Karelin both suggest linear relationships between decrease of ice thickness and accumulated degree-days above a chosen datum temperature.[3]

When ice survives for more than one year it continues to grow, but at an ever-decreasing rate until equilibrium between winter freezing and summer melting is achieved.[17] Maximum thicknesses for floes of old ice are in the range 3-4 m. During summer the floe may thicken by accretion of fresh melt water at its base; relatively pure melt water from the surface percolates down through leads and holes, freezing at the underside. In winter saline ice freezes to the base of the ice. Eventually the ice becomes too thick to develop much temperature gradient; the annual temperature wave is greatly attenuated in the lower layers, and there is no further growth.

Masses of sea ice which are much thicker than 4 m exist, but they are formed by floes being piled on top of each other by horizontal pressure.

BIBLIOGRAPHY

1. Anderson, D. L. (1960) The physical constants of sea ice, Research, vol. 13.
2. Assur, A. (1956) Airfields on floating ice sheets, U. S. Army Snow, Ice and Permafrost Research Establishment (USA SIPRE), Technical Report 36.
3. Bilello, M. A. (1960) Formation, growth and decay of sea ice in the Canadian Arctic Archipelago, USA SIPRE Research Report 65.
4. Callaway, E. B. (1954) An analysis of environmental factors affecting ice growth, Washington: U. S. Navy Hydrographic Office.
5. Defant, A. (1961) Physical oceanography, vol. 1. Pergamon Press.
6. Kolesnikov, A. G. (1958) "On the growth rate of sea ice," in Arctic sea ice. U. S. National Academy of Sciences, National Research Council, pub. 598.
7. Lee, O. S. and Simpson, L. S. (1954) A practical method of predicting sea ice formation and growth, Hydrographic Office, Technical Report 4.
8. Malmgren, F. (1927) On the properties of sea ice, Scientific Results, Norwegian North Polar Expedition, vol. 1, no. 5.
9. Mellor, M. (1960) Sea ice measurements at Mawson and Davis, 1954-58, Interim report 19, Australian National Antarctic Research Expeditions, Melbourne.
10. Pounder, E. R. (1961) "The physics of sea ice," in An analysis of environmental factors affecting army operations in the polar basin. Arctic Institute of North America for(U. S. Army Cold Regions Research and Engineering Laboratory) (USA CRREL).
11. Rodhe, B. (1952) On the relation between air temperature and ice formation in the Baltic, Geografiska Annaler, vol. 34, no. 3-4.
12. Sverdrup, H. U. (1956) "Arctic sea ice," in The dynamic north. U. S. Navy.
13. __________, Johnson, M. W., and Fleming, R. H. (1942) The oceans. New York: Prentice-Hall.
14. U. S. Navy Hydrographic Office (1957-58) Oceanographic atlas of the polar seas, Part I: Antarctic, Part II: Arctic. Washington: H. O. publication no. 705.
15. Weeks, W. F. (1958) "The structure of sea ice: a progress report," in Arctic sea ice. U. S. National Academy of Sciences, National Research Council, Pub. 598.
16. __________ and Anderson, D. L. (1958) Sea ice thrust structures, Journal of Glaciology, vol. 3, no. 23.
17. Zubov, N. N. (1945) L'dy Arktiki (Arctic ice). Moscow: Izdatel'stvo Glavsevmorputi. English translation published by U. S. Navy, Electronics Laboratory, 1964.

BIBLIOGRAPHY (Cont'd)

18. Schwerdtfeger, P. (1963) The thermal properties of sea ice, Journal of Glaciology, in press.

19. Assur, A. (1958) "Composition of sea ice and its tensile strength" in Arctic sea ice, U. S. National Academy of Sciences, National Research Council, Pub. 598. Also USA CRREL Research Report 44, 1960.

20. Weeks, W. F. (1961) Studies of salt ice I: The tensile strength of NaCl ice, USA CRREL Research Report 80. Also Journal of Glaciology, vol. 4, no. 31, 1962.

21. Armstrong, T. and Roberts, B. (1956) Illustrated ice glossary, Polar Record, vol. 8, p. 4-12.

22. U. S. Navy Hydrographic Office (1952) A functional glossary of ice terminology. Washington.

23. World Meteorological Organization (1956) Abridged international ice nomenclature, W.M.O. Executive Committee Report, vol. 8, p. 107-116.

CHAPTER XI. GLACIOLOGICAL TECHNIQUES.

The methods and equipment reviewed here are those commonly employed in what might be called "traditional glaciology", where field study of naturally occurring snow and ice masses is of most interest. The many varied and sophisticated techniques used in detailed laboratory studies of snow and ice as materials are not discussed.

Observing snow crystals.[26,29,38,46] Simple field observations on the size and form of particles of falling or blowing snow can be made by catching snow on a card covered with millimeter-graduation graph paper and examining through a hand lens. More detailed study can be made with the aid of a microscope, the particles being caught and examined on glass slides. Photomicrographs can be made when a suitable camera and adapter are available. The forms of crystals can be preserved by making replicas. Formvar resin (polyvinyl formal resin dissolved in ethylene dichloride) is delicately spread over the snow crystals; it hardens into a sheet, which can then be stripped and stored.

Measuring snow accumulation. Snow accumulation can be found by firmly planting long stakes (usually of bamboo) and measuring the exposed length from time to time. Since the density of accumulating snow is variable, and lying snow settles with time, density measurements must be made so that accumulation can be reported in terms of mass or water equivalent.[25,44,45] To obtain representative values for an area, a network of stakes is required; on an open snowfield a large cross of stakes can be used, one line running at right angles to the wind direction and another along the wind direction. The spacing of stakes varies with the size of the area to be studied and the resources available.

On permanent snowfields the glaciologist often measures only the annual surplus of accumulation, i.e., the amount of snow permanently added to the surface each year. If annual layering can be identified, the thickness of an annual layer multiplied by its density gives the required result in terms of mass or water equivalent.[5,6,41,49] In some areas annual layering is obvious when a pit is dug to reveal a vertical profile, icy melt bands or dust-scattered laminas marking previous summer surfaces. In other places, particularly the cold interior parts of Greenland and Antarctica, interpretation is more difficult, and careful studies of grain structure and texture, density, and crust occurrence are demanded. Stratigraphy in a pit wall may be rendered more easily visible by brushing with a whisk broom, by applying dye sprays, by etching with smoky flames, or by viewing in light transmitted from a borehole behind the pit wall. When a site is to be occupied or revisited, identifiable reference horizons may be laid down with dyes or string lines.

In some areas, layer thicknesses can be measured rapidly by correlating stratigraphy with penetrability. The instrument used is a Rammsonde, a spear-pointed rod which is driven vertically into the snow by repeated blows from a weight falling through a fixed height. Penetration resistance gives a measure of hardness at various depths.[7,47]

Core drilling yields snow and ice samples which can be taken to the laboratory for study by more complex methods. Variations of mechanical properties with depth can be obtained, thin section photographs taken to show structural characteristics, and isotope analyses made.

One interesting stable isotope technique is measurement of the ratio of oxygen isotopes in snow from different layers.[11,42] The ratio of the oxygen isotope with an atomic weight of 18 (O^{18}) to the common form having an atomic weight of 16 (O^{16}) varies in water precipitated in the atmosphere, and since the ratio is temperature-dependent (the concentration of O^{18} decreases as temperature falls) it gives a measure of relative temperatures for the periods of initial precipitation. Winter and summer snowfalls give different values for the O^{18}/O^{16} ratio, and alternations of the ratio with depth allow annual layers to be recognized, although they are indistinguishable by eye. The ratio is measured in a mass spectrometer. The method indicates long-term temperature trends when applied to deep cores. Hydrogen also has a heavy stable isotope, deuterium, which is present in ice. Again the concentration of the heavy isotope decreases with precipitation temperature, so that deuterium analysis can be applied in the same way as the oxygen isotope analysis.

The decay rates of such natural radioactive isotopes as tritium can be used to estimate the age of deeply buried snow layers (for snow which fell prior to hydrogen bomb

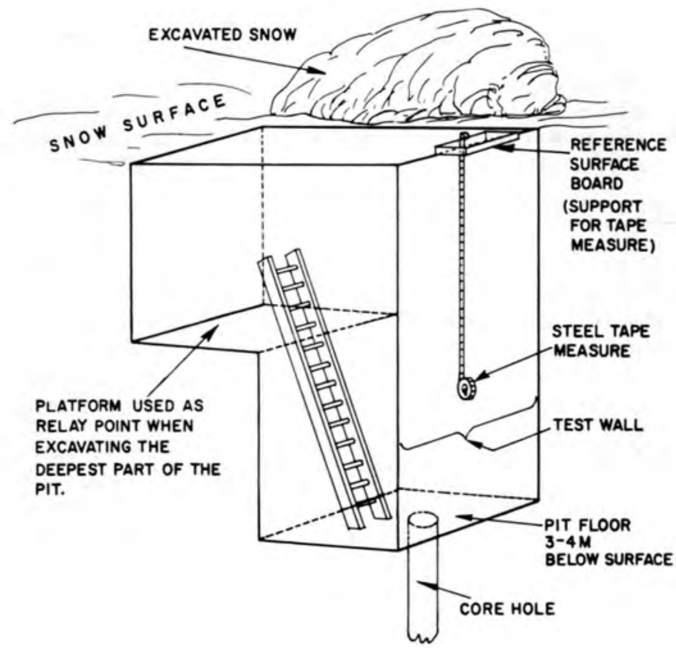

Figure XI-1. Typical shallow test pit in snow. The pit is dug in undisturbed snow and oriented so that the test wall is shaded from sunshine; spoil from the excavation is not piled above the test wall. (After Benson, refs. 5 and 6)

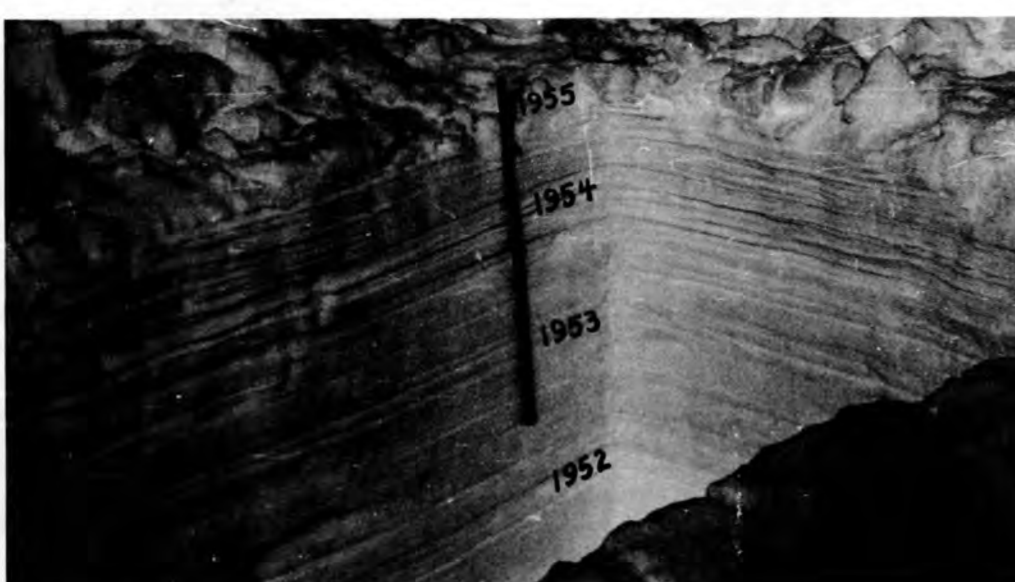

Figure XI-2. Stratification in dry snow revealed by brushing the walls of a test pit. Annual layers have been identified. (After Benson, ref. 6)

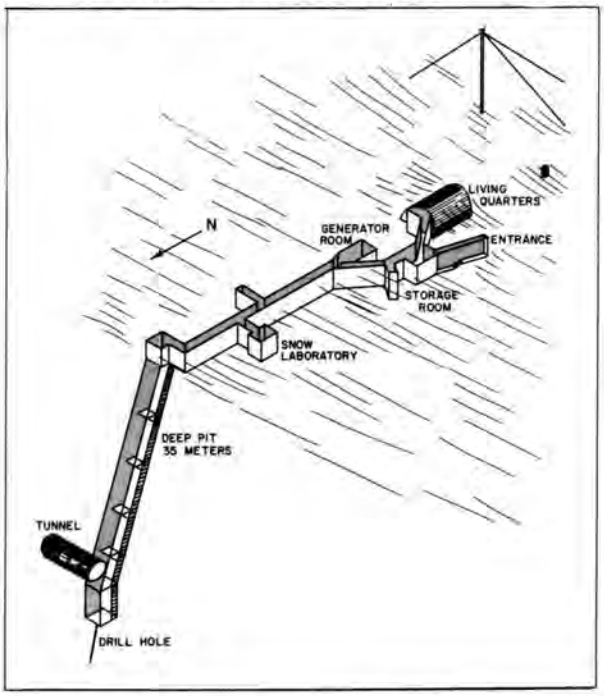

Figure XI-3. Layout of an ice cap test site for deep pit studies. The pit is inclined at about 15° to the vertical, and staging is provided about every 15 ft on the ladder. A bucket-hauling track is left clear alongside the ladder. Working and storage space is provided by excavations in the surface snow. (After Cameron et al., ref. 53)

testing in 1954).[1,30] The carbon-dating method (decay of the radioactive C^{14} isotope in CO_2) on entrapped air requires large volumes (amounting to tons in weight) of snow or ice, and therefore cannot be applied to cores, but it has been done with icebergs.[40]

For measuring snow accumulation in remote places a gamma radiation device has been developed.[13] A radioactive source is placed at the surface and a detector is suspended a fixed distance above it. As snow accumulates over the source the intensity of radiation at the detector decreases. The information is telemetered to a control center, where it is interpreted in terms of mass of snow.

Ablation measurements. Lowering of a snow or ice surface by melting, evaporation, or wind deflation can be measured on stakes.[25] To avoid having stakes displaced by melting during periods of high air temperatures or strong solar radiation, they are planted firmly in deep augered holes, so that only the tips project early in the season. Measurements on impermeable ice are straightforward, changes in exposed length giving the ice loss (which is multiplied by ice density to give water equivalent). On a snow surface, careful checks on the snow density must be made to account for melt-water infiltration and snow settlement.

Measurement of melt-water runoff involves stream gaging with suitable flow meters. This is laborious if many individual streams have to be gaged, and it is preferable to gage the major channels near the ice margin, using current meters, not pitometers, owing to extreme turbulence.

Where only light winds prevail, detailed records of ablation rate can be obtained from ablatographs, which are similar to simple water level recorders. As the surface is lowered by ablation, a specially designed "float" actuates a stylus, which in turn marks a trace on a revolving (clockwork) drum. The recorder itself is fixed firmly on immovable supports sunk deep into the ice.

Temperature measurements. When snow and ice temperatures are measured in places giving direct access, e.g., in pits, trenches, and tunnels, either liquid-in-glass or bimetal dial thermometers can be used.[6,46] Accurate measurements can be made with mercury-in-glass down to about -35C, and spirit-in-glass thermometers can be used to the lowest natural temperatures. Glass thermometers are fragile, and require metal casings for mechanical strength when pushed into snow. Bimetal dial thermometers are very convenient and easy to read, but are generally less accurate than liquid thermometers. Instruments with stems from 8 to 18 in. long are readily obtainable. Bimetal thermographs can be used for continuous recording of air temperature in ice cavities; to obviate ink freeze-up problems, a trace can be scratched on a smoked drum by a dry stylus.

Measurement of temperatures within a mass of snow or ice calls for remote-indicating instruments. Three main types are used.

a. Thermocouples: these bimetallic junctions are made by connecting two dissimilar wires, commonly copper and constantan. Readings are made on a suitably calibrated potentiometer, using a junction immersed in an ice-water bath as reference. They are cheap and expendable, and therefore can be permanently frozen-in after installation.

b. Thermohms: these are temperature sensitive wire-wound resistances. Readings are made on a Wheatstone bridge calibrated to give temperatures, a standard resistance in the bridge circuit serving as reference.

c. Thermistors: these are temperature sensitive semi-conductors, which are used in a bridge circuit in the same way as thermohms. They are capable of very high precision when used with suitable bridges and cables, but require careful individual calibration.

All electrical thermometers can be used with recording potentiometers when continuous records are required.

Vertical temperature profiles are often measured in open boreholes, the temperature sensitive element hanging freely in air or in the drilling fluid. This is usually satisfactory, but the possibility of temperature disturbance should be anticipated where metal borehole casing is placed.

When no electrical thermometers are available, borehole temperatures can be checked with liquid thermometers; the bulb is thermally insulated, and the thermometer read quickly on being raised to the surface after a lengthy "soaking" period.

When readings are made near the surface of snow or ice, care is needed to ensure that the sensing element does not absorb radiation and thus give spurious readings. If readings are made from vertical strings of thermometers near the surface, the problem of a moving datum arises: thermometers close beneath an ablation surface ablate out, those beneath an accumulation surface become buried more deeply.

Drilling in ice. Hand drilling in ice can be very laborious if special tools are not used. Good equipment for hand-drilling 1 or $1\frac{1}{2}$ in. diameter holes, and for core-drilling to extract a 3 in. diameter core, has been developed by CRREL, and this equipment can be obtained from commercial sources.

The hand auger for drilling $1\frac{1}{2}$ in. holes is part of the sea ice thickness kit (Fig. XI-7). It consists of a stainless steel auger 1 m long, with extension rods giving a total length of 3.5 m. A wood brace is used for turning the auger, and fast coupling is effected by spring-loaded button connectors.

The hand coring auger consists of a 1 m long core barrel fitted with a special cutting head and having external auger flights. It is turned by means of either a wood brace or a T-bar. Extension rods are each 1 m long, and are coupled rapidly by special pins. The equipment can be used purely by hand to a depth of about 10 m (33 ft), and can core to about 30 m (100 ft) when a hoisting tripod is used. Small motor attachments, such as those used on post-hole diggers, have been used for turning the drill, although there is a danger of jamming at high rotational speeds.

Figure XI-4. CRREL (formerly SIPRE) test kits. The items are, from left to right: Rammsonde kit (see Fig. XI-5). 3-in. ice coring auger kit (comprising 1 m long core barrel, coring head with removable bits, core barrel connector, brace and T-bar, extension rod adapters and connecting pins, spare cutting bits and small tools for removing and sharpening bits. The box containing 1 m long extension rods is not shown). Snow density kit (see Fig. XI-6). Ice thickness kit (see Fig. XI-7). Salinity kit (comprising four hydrometers calibrated in salinity units at 15C, and covering the range 0-40‰; flotation cylinder with thermometer; temperature correction chart).

Figure XI-5. Rammsonde kit, comprising: 1-m long spear-pointed sonde, with shaft graduated in centimeters; four 1-m long graduated extension rods (3 cm diam); two pierced cylindrical hammers (1 and 3 kg weight); graduated guide rod for hammers (for maximum drop of 50 cm).

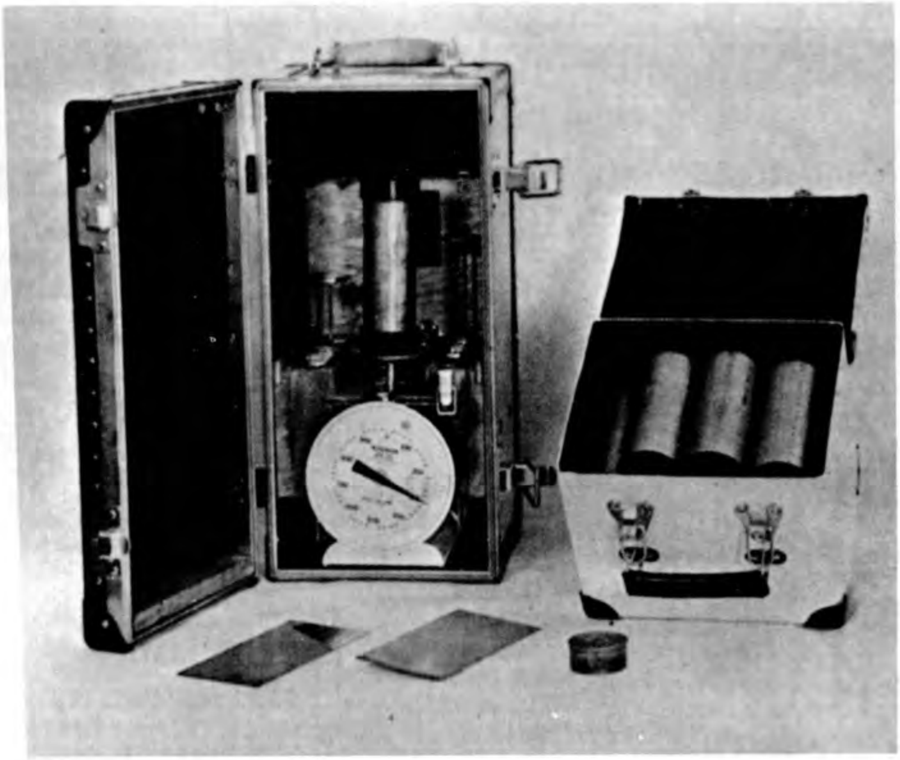

Figure XI-6. Snow density kit. Contains: 1000-g capacity spring-weighing scale with rotatable dial; stainless-steel snow-sampling tubes, 500 cm^3 volume, with rubber end caps; metal plate for trimming samples; cards with 1-mm square graduation, laminated in plastic; 8-in. stem dial thermometers, with piercing rod. Some kits contain a Canadian hardness tester.

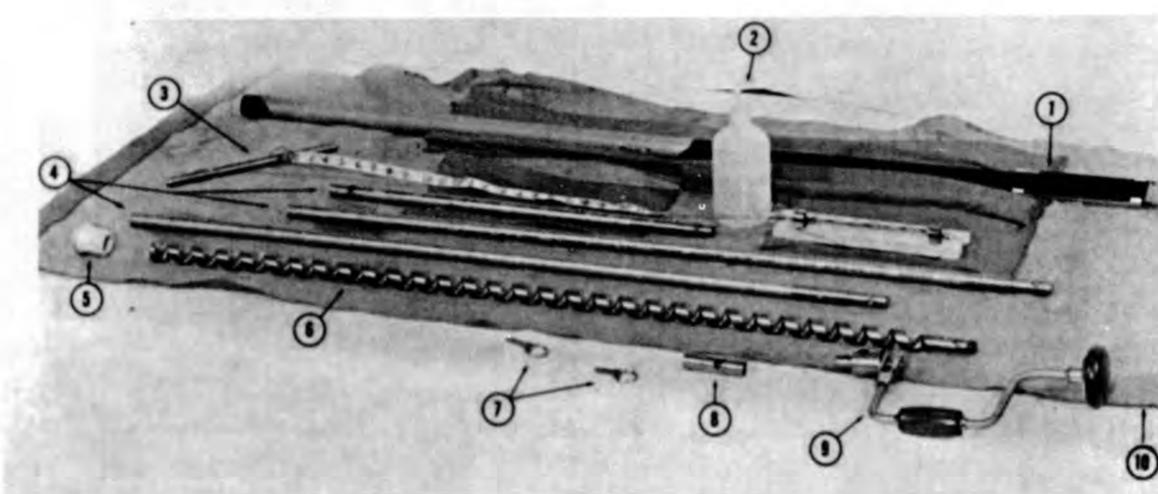

Figure XI-7. Ice thickness kit. Items shown are: 1. Ice chisel; 2. 1-pint polyethylene wash bottle; 3. Measuring tape with end rod; 4. Extension rods, 1 m and $\frac{1}{2}$ m long; 5. Rubber protective cap for auger; 6. Auger, 1 m long; 7. Retaining pins; 8. Adapter; 9. Brace; 10. Canvas cover with carrying strap.

Figure XI-8. 3-in. ice coring auger being used with a lifting tripod.

Other types of augers have been developed for drilling to shallow depths by hand or by motor attachment. The augers used by ice fishermen might be mentioned, and the electric-powered augers used in construction have been employed.

For medium-depth drilling, say in the range 20-50 m, power augers have been used. Shot-holes for seismic studies, and open holes for temperature measurements, have been bored with portable flight augers on which feed, lift and rotation were hydraulic-powered. Sled-mounted drills with independent power units, and vehicle-mounted drills run from power take-offs have been employed on ice caps.

Deep core drilling is specialized work requiring heavy equipment and experienced drillers.[22, 31, 33] The deepest holes drilled in Greenland and Antarctica so far (about

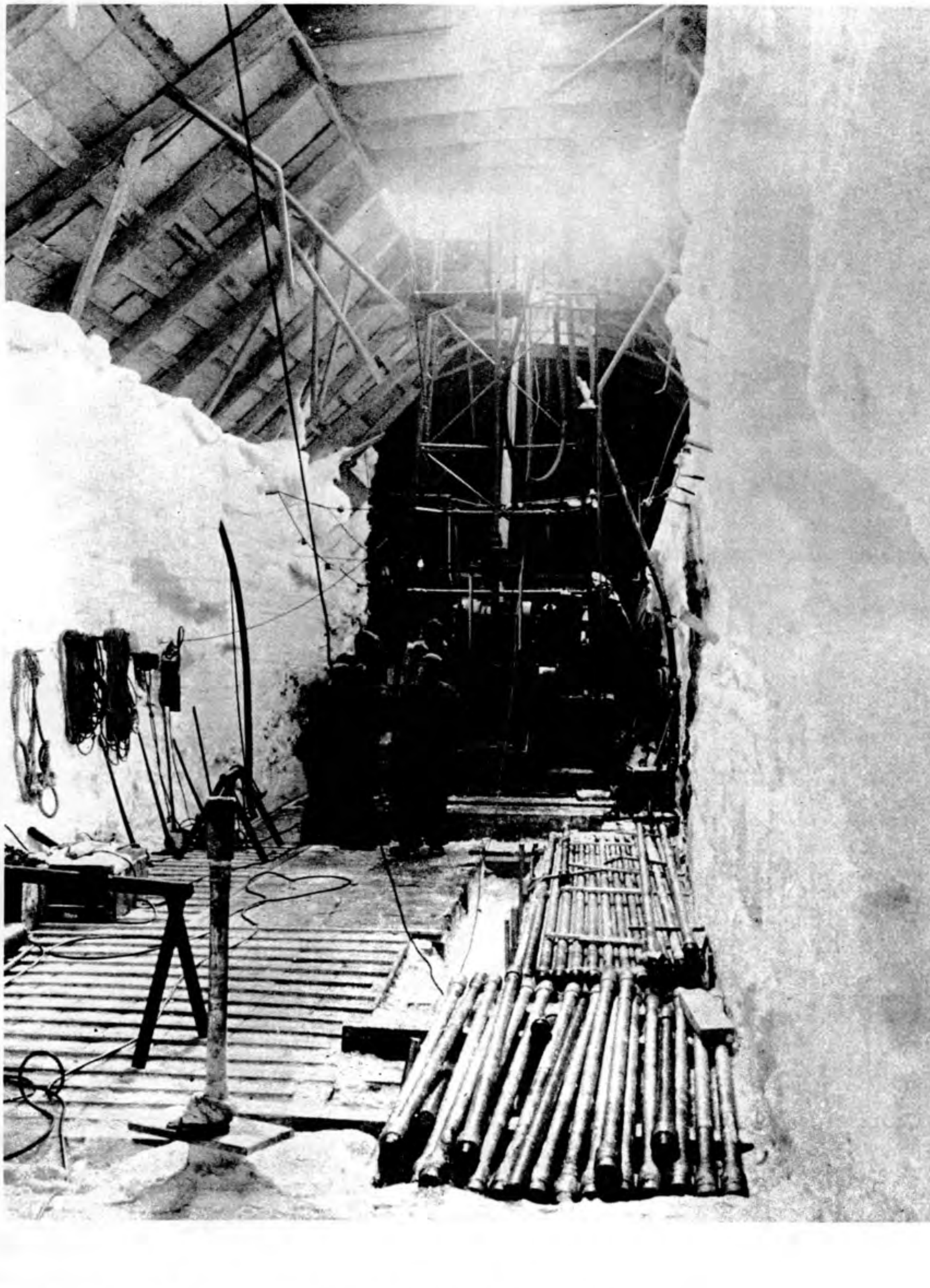

Figure XI-9. Failing Model 1500 well-drilling rig coring in the Greenland Ice Cap. The rig is housed in a trench covered by a timber roof.

1400 ft deep) were made by core drilling with conventional rotary equipment*, using refrigerated compressed air as drilling fluid, and a controlled uniform rate of feed with the drill string in tension. Good core was recovered, but it is believed that core quality in deep ice could be improved in future operations by using cooled diesel fuel instead of compressed air. Borehole casing must be placed when penetrating the permeable snow mantle on an ice cap in order to retain the drilling fluid. Russian borings in Antarctica have apparently been made by large augers, but few details are available.

Experiments with thermal drilling have been made from time to time, and a number of small diameter holes have been sunk in glacier ice with hot-point devices, usually electrically heated. A new rig for very deep core drilling has been developed by CRREL using the melting principle, and it is hoped that it will be capable of penetrating the full thickness of a polar ice sheet. An annulus of ice is melted by an electrically heated ring, and melt water is drawn into a heated container while the core feeds into the core barrel. The drill is lowered down the hole on a load-bearing electrical cable, power being supplied at the surface. The hole is filled with a cold fluid to prevent closure. Advantages include good core quality, high rate of hoisting and lowering, cable replacing heavy drill steel, and transmission of power electrically rather than mechanically.

One of the limitations of the thermal drill is its inability to penetrate rock-bearing ice layers, such as exist near the bed of a glacier. With this in mind, new equipment consisting of a self-contained electric-powered rotary drill is being modified at CRREL for use in ice. The entire coring unit, complete with electric motor and reaction device, is to be lowered down the hole on a wire line, power being supplied from the surface.

Measuring the density of snow and ice. Density is a useful index to a wide range of physical properties, particularly for snow. It is normally measured by weighing a known volume of snow or ice,[46] but occasionally the weight is obtained indirectly by measuring the volume of melt water from a sample.

In soft snow, open-ended sampling tubes can be driven into pit walls, the samples thus cut being trimmed to size with a spatula or by special cut-off plates in the tube. Samples are weighed on spring scales or a beam balance. A SIPRE (CRREL) snow kit facilitates this work; it contains many 500-cm^3 sampling tubes (of uniform weight), with rubber sealing caps and a spring scale which can be zero-set to allow for the weight of an empty tube. Multiplication of the scale reading by a factor of 2 gives density.

In very hard snow, or in ice, tubes cannot be driven. Instead, cylindrical cores are cut with the auger described above, or rectangular blocks are sawn out and dressed with the aid of a miter box. The core or block is carefully measured to give volume, and is then weighed.

The volume of a small irregular piece of impermeable ice can be measured by displacement of a low-density cold liquid, such as cold kerosine, in a graduated vessel.

Attempts have been made to measure density of snow and ice in-place from boreholes. Back-scattering of neutrons and gamma radiation has been employed, but so far the accuracy and discrimination have been inadequate.

Measuring surface movement on glaciers. Measurements of surface movement on glaciers are based on established surveying techniques, and for best results the work should be planned and executed by trained surveyors.

One simple method for measuring flow is the displacement of stakes planted along a line transverse to the general direction of flow. A transverse line is defined optically with a transit or a theodolite, set on exposed rock at the ice margin, either by sighting it on a fixed reference mark on the opposite side of the glacier, or by turning an angle from some other fixed object. Stakes are set on the line initially, and their displacement at subsequent dates is found by measuring from the stakes to the optical line with a tape. There are limits to sighting and signalling distance, so that for very long transverse cross-sections the line must be ranged forward by repeated set-ups of the theodolite.

*A Failing Model 1500 well-drilling rig (military model 314)

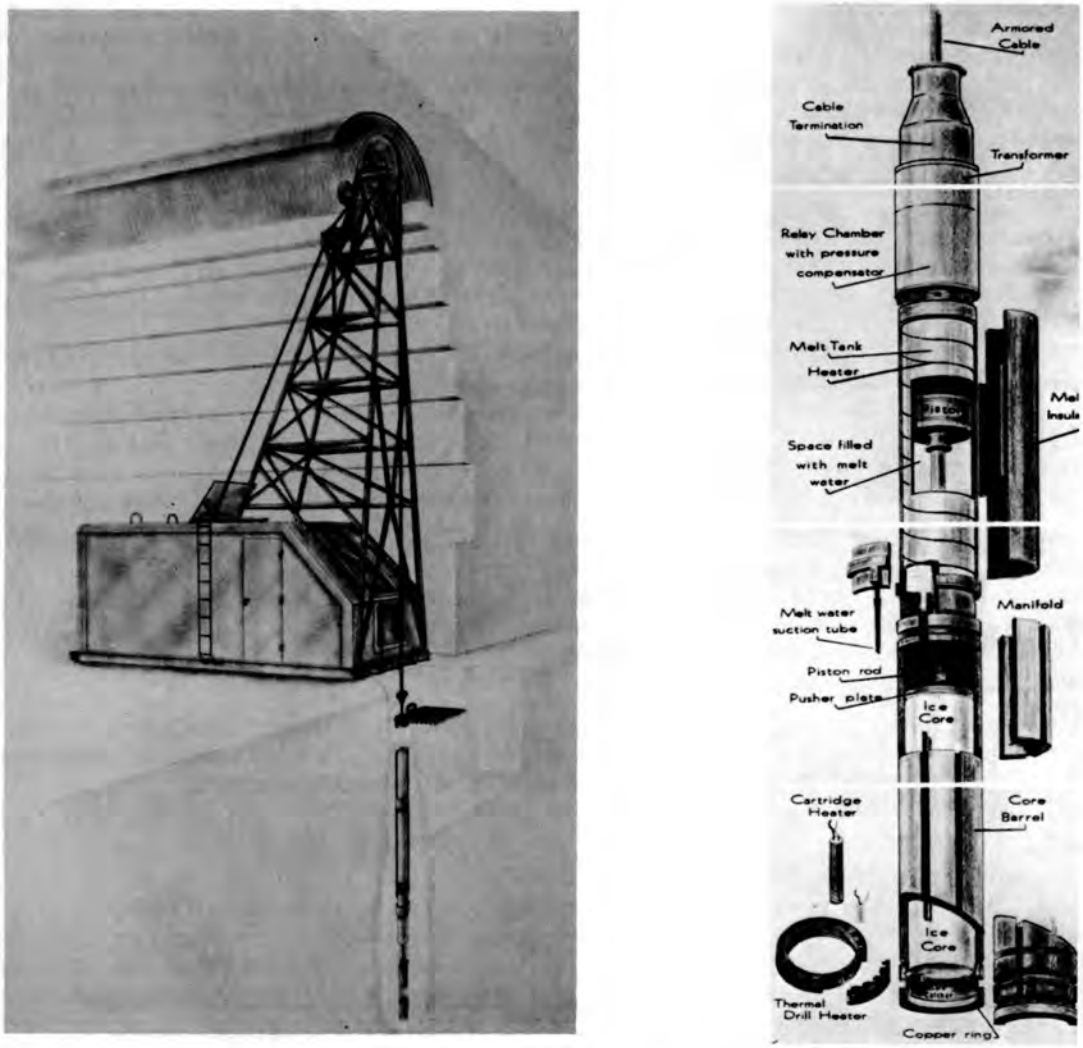

Figure XI-10. CRREL thermal drill for deep coring in ice. An annular heating element melts through the ice, and melt water is drawn into a storage chamber while core feeds into the core barrel.

The position of stakes, not necessarily along a single line, and of natural marks such as boulders, can be checked from time to time by triangulation. This calls for sighting with a theodolite from two or more fixed stations whose coordinates are known. If the coordinates of the instrument stations are not already known, or cannot be determined from existing triangulation systems, a base line must be measured. This method is more demanding in observation and data reduction than the simple method described above.

Theodolite measurements give movement data for relatively few points on the ice; more detailed information on the distribution of flow velocities over the whole glacier surface can be obtained by photogrammetry.[12, 17, 18, 52] Ground phototheodolites have been used effectively to study flow features on small glaciers, but repeated air photogrammetry is more potent for large and relatively inaccessible glaciers. When photographic runs are flown over snow and ice, great care must be exercised in selecting suitable lighting conditions and exposure, so that surface texture appears on the prints. Movements of both natural surface marks and artificial targets can be checked from photographs; absolute and relative movements can be found if the survey is tied to fixed rock control points, but only relative motion can be measured if there is no fixed reference. Some plotting can be done by "hand" methods,[28] but more detailed work can be done with the aid of plotting machines. Air photogrammetry cannot be undertaken without the aid of trained specialists.

On the unbroken expanses of an ice cap or ice shelf, where no rock is visible, the only method available for checking absolute movement is repeated astrofixing. This is not a very good method for slowly moving ice, as many years must elapse before movement exceeds the probable error in determination of position (unless expensive and time-consuming repetitive determinations with special instruments are made). Measurements of relative motion can be made by laying out a network of markers and checking their positions from time to time by triangulation, working from measured baselines.

Linear distances can now be measured rapidly and accurately with electronic devices, and trilateration is therefore becoming feasible for measuring both absolute and relative movements over large areas. The tellurometer, a device which transmits a micro-wave signal and receives it after return from a remote unit, measures distances of 50 miles or more with high precision. Other devices, the geodimeter and the Electrotape, serve a similar purpose, the geodimeter being particularly suited for short lines of a few miles in length.

Measuring subsurface movements and strains. To find the vertical distribution of velocity in a glacier, observations on the deflection of a borehole are made. A deep borehole is repeatedly surveyed for magnitude and direction of inclination at various depths, using a directional inclinometer. Inclinometers of the type used for geophysical exploration have a pendulum hanging over a compass card. The card is photographed by a small built-in camera, recording direction and inclination. Another type used in glacier ice runs in a grooved plastic borehole casing, employing a compass and camera for azimuth determination and a pendulum-actuated potentiometer for inclination readings. A special high-precision inclinometer, embodying pendulums and strain gages, is being made by CRREL for measurement of small angular changes.

Subsurface deformation has also been studied through the propagation of ultrasonic signals through glacier ice.[21] Piezo-electric ceramics were placed in a set of boreholes, and changes in the geometry of the original prism were measured by transmitting signals from one ceramic to the other.

Borehole closure rates at various depths are measured by means of electrical calipers.[15] The instrument currently used by CRREL is based on a variable resistance actuated by a set of arms which press against the side of the borehole; a new one will depend on strain gages mounted on arched leaf springs.

The simplest way of observing deformation in tunnels, pits and trenches is direct measurement between fixed points, using survey tapes. Wood dowels frozen firmly into the ice make convenient points. Visual grids may be marked on the ice for subsequent remeasuring and for photographic records of deformation. Dial micrometers provide a convenient means of measuring relative movement: struts are so set that the dial gage is actuated as the anchor points of the struts move. Continuous records of deformation have been obtained at unmanned sites by directing a collimated beam of gamma radiation at a sensitive plate: traces on the plate record relative motion of the points where gamma source and plate are anchored. Remote-indicating deformation gages have been built from helical potentiometers.[50] Two points are connected by a weighted cord, which turns a pulley driving the potentiometer as relative movement of the points takes place. Resistance of the helipot, as this device is called, is read on a bridge, and calibration data give a conversion to linear movement in the cord.

Measuring ice thickness over land and water. The standard method for measuring thick ice (more than 200 ft thick) over land is the seismic reflection method.[34,35,51] An explosive charge is fired at the surface, preferably in a borehole, and the travel time for elastic waves to reach the base of the ice and return to an array of surface geophones after reflection is measured. Wave propagation velocities for various depths in the ice are measured by refraction shooting, and thus the ice depth can be computed. With floating ice there is only weak reflection at the ice-water interface, and more complicated interpretation of seismic records is called for.[9] One method for determining the thickness of floating ice is based on multiple reflections between the sea floor and the undersurface of the ice. Seismic work requires specialized equipment and trained personnel.

Measurements with a sensitive gravimeter can be used for depth measurements over a short distance between control points. Control is provided by seismic soundings, or by readings on rock exposures. Hence, gravimeters can be used to fill in detail between seismic shotpoints and to measure thickness across small glaciers. For precise gravity work, surface elevation and latitude must be accurately known.

Seismic soundings have been checked in a number of places where the ice depth is known from deep drilling.

Continuous depth sounding through ice would be highly desirable, and development work is in progress on electronic devices designed to transmit a signal of suitable frequency through the ice and receive it after reflection from the bed. This concept grew from the failure of early aircraft radio-altimeters, the signals of which penetrated the ice cap instead of reflecting from the surface.

The thickness of a free-floating ice shelf in a given area can be estimated by finding the elevation of the surface above sea level. Gross densities of the ice and the sea water are fairly well known, so that ice thickness can be calculated from simple buoyancy considerations.

The thickness of sea ice is measured by boring through it with an auger (hand or power), and lowering a tape fitted with a tilting crossbar at the lower end (included in CRREL sea ice thickness kit).

Studying the internal structure of snow and ice. To study the size, shape, orientation, and bonding of grains and crystals in snow and ice, thin sections are prepared for examination and photographing in transmitted polarized light.

Preparation of ice thin sections begins by freezing the smoothed surface of a sample to a gently warmed glass slide, working in a cold environment where the temperature is below -10C. The bulk of the sample is then cut off by sawing parallel to the glass slide with a hand saw or band saw, leaving a layer about 1 mm thick. Final reduction to a thickness of about 0.4 mm or less is achieved by gentle abrasion, gentle warming or, most desirable, by shaving with a microtome.[23]

Most glacier ice has quite large grains — of the order of 1 cm diameter — and no magnification is needed for thin section studies. Thin sections of ice are therefore examined on a special universal-stage capable of carrying thin sections up to about 5 in. across.[23] The Rigsby universal stage used by American glaciologists has four axes of rotation, which permit a thin section to be pivoted any way until the orientation of the optic axis of any given crystal is found by extinction of light transmitted through crossed polaroids and the crystal lying between them. Use of the stage, and plotting of crystal orientations on fabric diagrams has been described in detail by Langway[23] and others; the techniques are familiar to geologists and crystallographers.

Reflex cameras can be fitted to the universal stage to record details of each thin section studied. A glass plate with an etched graticule (a 1 cm grid) is laid over the thin section during photographing to provide scale on the prints.

When thin sections of snow are required, simple shaving or grinding will not serve, as the grains crumble apart. The sample is therefore impregnated with water-saturated aniline (other liquids have been used, but aniline is convenient) at a temperature between -5 and -10C, and frozen into a solid block at -20C, with a glass plate frozen on one side. It is then shaved to a thickness of about 0.1 mm with a microtome or a carpenter's plane. The aniline solution can be melted out again at -5C, leaving a thin section for study of grains and intergranular bonds.[19]

Figure XI-11. Four-axis universal stage with camera. (Photo from Langway, ref. 23)

Snow grains are commonly smaller than 1 mm in diameter, and snow thin sections are studied with the aid of a low-power polarizing petrographic microscope. Photomicrographs are made by attaching a camera to the eyepiece end.

Bubble studies. Studies of the bubbles of entrapped air in glacier ice throw light on metamorphic processes, relations between bubble pressure and stresses in the ice, and on the history and source of origin of the ice.

Dimensions and locations of bubbles are found by visual examination. The average bubble pressure in a sample is determined by measuring the initial porosity (density) of the sample and then finding the volume of air released by melting at atmospheric pressure.[24, 43] The sample is melted in a liquid-filled system so that air from the bubbles collects by displacement in an inverted burette, where its volume at atmospheric pressure is measured. Application of the gas law allows the pressure corresponding to the volume and temperature of the original bubbles to be computed. The pressure in individual bubbles has been measured by slowly melting an ice sample inside a liquid-filled pressure vessel fitted with a pressure regulator and observation window.[39] As a bubble is opened by melting, the air inside it is held back by regulating the pressure inside the vessel, which then equals the bubble pressure.

Detailed gas analyses have been made on air from bubbles in glacier ice.[40] Gas from the bubbles is extracted by boiling the sample under vacuum in a sealed vessel, the gas being collected in a mercury extractor while steam is returned by a reflux condenser. The composition of air in old ice which has always been free from melting and contamination should yield data on ancient atmospheres, and carbon-dating of air from very large samples should determine the age of the ice. Samples collected from icebergs and glacier snouts on the Greenland coast, however, have evidently been subjected to melting (air dissolved in melt water differs in quantitative composition from free air) and contamination by organic dust.

Measurements on blowing snow. The rate at which blowing snow is transported horizontally at any given level is found by extracting snow from a narrow airstream during a timed interval.[14; 26, 27] Gages extend a slim nozzle upstream into the wind, and orient in the wind direction by means of a pivot and vane. Snow grains can be extracted from the air inside the gage by expanding the stream and reducing the velocity so that grains settle out. Baffles and filter gauzes are used in some gages, but since these introduce appreciable resistance to air flow, a controllable extractor at the exhaust end is needed to ensure that there is no back-pressure at the inlet. Gages are mounted on a mast at various heights above the surface, vertical spacing usually increasing with increasing height.

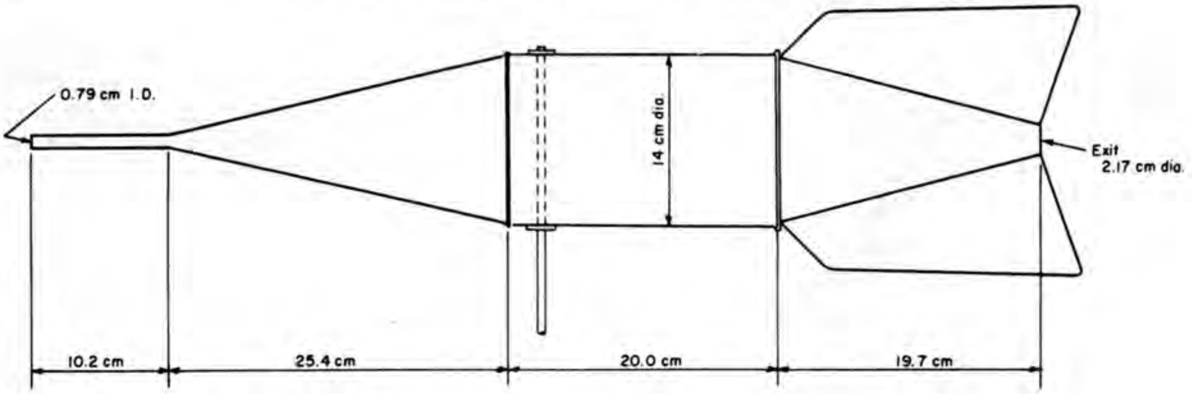

Figure XI-12. Rocket type gage for measuring blowing snow. Gages are pivot-mounted on a mast at various heights above the surface. Snow-bearing air enters the small orifice, which always points into wind, and the snow particles settle out as velocity decreases in the expanded center section. (After Mellor, ref. 27)

To find the drift density, or mass of snow present per cubic meter of air, wind velocity measurements must be made simultaneously at the gage heights.

Drift density can also be measured by means of light beams shining on photoelectric cells. The amount of drifting snow between the light source and the cell governs the intensity of the light received by the cell. Experiments have also been made using gamma radiation instead of visible light.

The electrical effects of blowing snow have been studied, usually by exposing antennas and discharging them through recorders. Readings of snow transport and wind speed, together with visual observations of grain sizes and shapes, should be made simultaneously.

Studying solid inclusions. Solid matter present in snow and ice to any appreciable concentration can be removed by sieving, filtering, centrifuging, or sedimentation. Microscope examination and chemical analysis are used to determine the composition of solid inclusions.

When extremely small particles at very low concentrations (e.g., the particulates of polar ice sheets) are to be studied, special techniques are required. Stringent measures are taken to avoid contamination during collection and handling of samples and removal of water substance. Electronic particle counters are utilized for detection and measuring. Electron microscopy is a recent technique used in the micro-chemistry of these particulates.

Determining the free water content of snow. The amount of liquid water contained in wet snow is usually found by melting a sample in warm water inside a vacuum flask. Repeated weighings give the amount of water initially in the flask, and the weight of wet snow added. Temperature measurements give the temperature drop after the sample has been melted. The percentage of free water in the snow can then be calculated if the water equivalent of the calorimeter (flask) is known.[3,48]

Wet snow can also be melted while totally enclosed in a calorimeter by passing a metered electric current through it.[16]

A variant of the calorimeter method, which is less demanding in precision of temperature measurement, is based on freezing the liquid water instead of melting the snow.[32]

Other methods depend upon water removal by centrifuging,[20] chemical reactions, and the compressibility characteristics of wet snow.

Measuring the salinity of sea water and sea ice. The salinity of sea water, and of melt water from sea ice, is usually measured by sensitive hydrometers, which can be calibrated directly in salinity units. Comparison with titration shows this to be sufficiently accurate for most purposes. A set of three hydrometers is required to cover the range from low-salinity sea ice to normal sea water with adequate precision. Water temperature is taken at the time the hydrometer is read, and salinity is then obtained from a correction chart.

Meteorological and micrometeorological observations. The weather observations needed for glaciological studies are usually made with standard instruments. Some care is called for in siting the instruments so that anomalous wind and radiation effects are avoided.

Micrometeorological observations should be matched to the study in progress. Indiscriminate data collection can lead to the project becoming overloaded with a mass of unusable records.

General physical properties of snow and ice. The mechanical, thermal and electrical properties of snow and ice are measured by a wide variety of laboratory methods, many of which can be used in field laboratories. These techniques are outside the scope of the present work, but are described elsewhere (e.g., SIPRE/CRREL Reports).

BIBLIOGRAPHY

1. Bader, H. (1958) "U. S. polar ice and snow studies," in Geophysics and the I.G.Y. American Geophysical Union, publication no. 590.

2. ________ (1951) Introduction to ice petrofabrics, Journal of Geology, vol. 59, p. 519.

3. ________ (1950) Note on the liquid water content of wet snow, Journal of Glaciology, vol. 1, p. 466.

4. ________ (1950) The significance of air bubbles in glacier ice, Journal of Glaciology, vol. 1, p. 443.

5. Benson, C. S. (1959) Physical investigations on the snow and firn of N.W. Greenland during 1952, 1953 and 1954, U. S. Army Snow, Ice and Permafrost Research Establishment, Research Report 26.

6. ________ (1960) Stratigraphic studies in the snow and firn of the Greenland ice sheet, California Institute of Technology, Ph.D. thesis.

7. Bull, C. (1956) Use of the Rammsonde as an instrument for determining the density of firn, Journal of Glaciology, vol. 2, p. 714.

8. Coachman, L. K., Hemmingsen, E., and Scholander, (1956) Gas enclosures in a temperate glacier, Tellus, vol. 8, p. 415.

9. Crary, A. P., et al. (1962) Glaciological studies of the Ross Ice Shelf, Antarctica, American Geographical Society, I.G.Y. glaciological report no. 6.

10. Dingle, W. R. J. and Radok, U. (1960) "Antarctic snow drift and mass transport," in Symposium on Antarctic Glaciology, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 55.

11. Epstein, S. and Sharp, R. P. (1959) Oxygen isotope studies, Transactions, American Geophysical Union, vol. 40, p. 81.

12. Finsterwalder, R. (1958) "Measurement of ice velocity by air photogrammetry," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

☆ U. S. GOVERNMENT PRINTING OFFICE: 1981--701-665--106

BIBLIOGRAPHY (Cont'd)

13. Gerdel, R. W. and Mansfield, C. W. (1950) Automatic snow gage for measuring and telemetering water equivalent of snow, Proceedings of the Western Snow Conference.

14. Govorukha, L. S. and Kirpichev, E. F. (1961) First results of snow drift observations with the Cyclone snow trap, Bulletin of the Soviet Antarctic Expedition, no. 26.

15. Hansen, B. L. and Landauer, J. K. (1958) "Some results of ice cap drill hole measurements," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

16. ________ and Jellinek, H. H. G. (1957) A portable adiabatic calorimeter, USA SIPRE Technical Report 49.

17. Haumann, D. (1960) Photogrammetric and glaciological studies of Salmon Glacier, Arctic, vol. 13, no. 2.

18. Hofmann, W. (1958) Bestimmung von Gletschergeschwindigkeiten aus Luftbildern (Determination of glacier velocities from aerial photographs), Bildmessung und Luftbildwesen, no. 3, p. 71.

19. Kinosita, S. and Wakahama, G. (1959) Anirin koteiho ni yoru sekisetsu no hakuhen no kansatsu, (Preparation of thin sections of deposited snow by the use of aniline), Teion-kagaku, Series A, vol. 18, p. 77 (text in Japanese).

20. La Chapelle, E. R. (1956) Centrifugal separation of free water from melting snow, Journal of Glaciology, vol. 2, p. 769.

21. Lafargue, M. and Millecamps, R. (1958) Sur l'application d'une méthode électro-acoustique originale à des mesures physiques et mecaniques de glaciologie (On the application of a new electro-acoustic method for physical and mechanical measurements in glaciology), Acad. Sci. (Paris), Comptes Rendus, vol. 246, p. 970.

22. Lange, G. R. (in preparation) Deep rotary core drilling in ice, U. S. Army Cold Regions Research and Engineering Laboratory, Technical Report 94.

23. Langway, C. C. (1958) Ice fabrics and the universal stage, USA SIPRE, Technical Report 62.

24. ________ (1958) "Bubble pressures in Greenland glacier ice," in Physics of the movement of the ice, Symposium of Chamonix, International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 47.

25. Lister, H. (1958) "Glaciology," in Venture to the Arctic (R. A. Hamilton, Editor). Penguin Books.

26. ________ (1960) Solid precipitation and drift snow, Trans-Antarctic Expedition, Scientific Reports, no. 5.

27. Mellor, M. (1959) "Gauging Antarctic drift snow," in Antarctic meteorology. Pergamon Press.

28. ________ (1958) Photogrammetric flow measurements on Antarctic glaciers, Transactions, American Geophysical Union, vol. 39, no. 6.

29. Nakaya, U. (1954) Snow crystals, natural and artificial. Cambridge, Mass: Harvard University Press.

30. Oeschger, H., Renaud, A., and Schumacher, E. (1962) Essai de datation par le tritium des couches de névé du Jungfraufirn et determination de l'accumulation annuelle (Attempt at tritium dating of névé layers of the Jungfraufirn and determination of the annual accumulation), Bulletin de la Société vaudoise des Sciences naturelles, vol. 68, no. 306. Lausanne.

BIBLIOGRAPHY (Cont'd)

31. Patenaude, R. W., Marshall, E. W., and Gow, A. J. (1959) Deep core drilling in ice at Byrd Station, Antarctica, USA SIPRE Technical Report 60.

32. Radok, U., Stephens, S. K., and Sutherland, K. L. (1961) "On the calorimetric determination of snow quality," in Snow and Ice Commission, General Assembly of Helsinki (1960), International Union of Geodesy and Geophysics. International Association of Scientific Hydrology, Pub. no. 54.

33. Ragle, R. H., Hansen, B. L., Gow, A. J., and Patenaude, R. W. (1960) Deep core drilling in the Ross Ice Shelf, Antarctica, USA SIPRE Technical Report 70.

34. Robin, G. de Q. (1958) Seismic shooting and related investigations, Scientific Results of the Norwegian-British-Swedish Antarctic Expedition, vol. 5, Oslo.

35. ______ (1962) The ice of the Antarctic, Scientific American, vol. 207, no. 3.

36. Schaefer, V. J. (1947) Properties of particles of snow and the electrical effects they produce in storms, Transactions, American Geophysical Union, vol. 28, p. 587.

37. ______ (1950) A new method for studying the structure of glacier ice, Journal of Glaciology, vol. 1, p. 441.

38. ______ (1957) The preparation of snow crystal replicas, Proceedings of the Eastern Snow Conference.

39. Scholander, P. F. and Nutt, D. C. (1960) Bubble pressures in Greenland icebergs, Journal of Glaciology, vol. 3, no. 28.

40. ______, et al. (1961) Composition of gas bubbles in Greenland icebergs, Journal of Glaciology, vol. 3, no. 29.

41. Schytt, V. (1958) Glaciology II, Scientific Results of the Norwegian-British-Swedish Antarctic Expedition, vol. IV. Oslo.

42. Sharp, R. P. (1960) Glaciers. University of Oregon Press.

43. Shumskii, P. A. (1955) Osnovy strukturnogo ledovedniia, (Fundamentals of ice structure). Moscow: Izdatel'stvo Akademii Nauk.

44. Swithinbank, C. W. M. (1954) Use of stakes in measuring the accumulation of snow on glaciers, Rome assembly, International Union of Geodesy and Geophysics. Association of Scientific Hydrology.

45. ______ (1957) The regime of the ice shelf at Maudheim as shown by stake measurements, Scientific Results of the Norwegian-British-Swedish Antarctic Expedition, vol. III. Oslo.

46. U. S. Army CRREL (1962) Instructions for making and recording snow observations, Instruction manual 1.

47. U. S. Army SIPRE (CRREL), Instructions for use of Rammsonde (Kegetonde), USA SIPRE instruction sheet.

48. U. S. Army SIPRE (1952) Instructions for the measurement of the free water content in snow, USA SIPRE Operation Manual no. 2.

49. Waterhouse, R. W. (1962) Analysis of data from a snow profile, USA CRREL Research Report 90.

50. ______ (1962) Camp Century movement reference, USA CRREL Technical Report 121.

51. Woollard, G. P. (1962) The land of the Antarctic, Scientific American, vol. 207, no. 3.

52. Baussart, M., Carbonnell, M., and Bauer, A. (1962) Photogrammétrie et glaciologie, (Photogrammetry and glaciology), Expéditions Polaires Francaises, Pub. no. 226. Paris

53. Cameron, R. L., et al. (1959) Wilkes Station glaciological data 1957-58, Ohio State University Research Foundation, Report 825-1, Part III.