COLD REGIONS SCIENCE AND ENGINEERING

Part I: Environment

Sect B: Regional

I-B2

Sept 1961

The Greenland Ice Sheet

U. S. ARMY

COLD REGIONS RESEARCH AND ENGINEERING LABORATORY

Corps of Engineers

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F. J. Sanger, Editor
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by Henri Bader

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PREFACE

This monograph summarizes existing knowledge of Greenland for the use of professional engineers engaged in design or construction in that region.

The paper has been reviewed and approved for publication by the Office of the Chief of Engineers, U. S. Army.

W. L. NUNGESSER
Colonel, Corps of Engineers
Director

Manuscript received 1 March 1961

Department of the Army Project 8S66-02-001

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EDITOR'S FOREWORD

This is the second section of "Cold Regions Science and Engineering" to appear. As part of the larger work, it is not a complete dissertation on Greenland but presents basic information, with a minimum of theorizing, about the particular ice sheet. General geology, physiography, environment and glaciology are dealt with in other sections; history and exploration are not included in a work written to give background material for engineering on the polar land ice sheets.

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

I. Environment
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  1. Micrometeorology
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- B. The Physics and Mechanics of Snow
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- B. Ice engineering
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- D. General engineering

IV. Miscellaneous

F. J. SANGER

29 September 1961

THE GREENLAND ICE SHEET

by

Henri Bader

The Greenland ice sheet, covering most of the island of Greenland, lies between the parallels of $60\frac{1}{2}$°N and 82°N, and meridians 20°W and 72°W (Fig. 1).

DIMENSIONS

Linear dimensions

The greatest length is almost 2400 km (1500 miles) along the 44°W meridian, and the greatest width is some 1100 km (700 miles) at latitude 77°N. A point located near $43\frac{1}{2}$°N $77\frac{1}{2}$°W has the maximum distance of 430 km (270 miles) from the nearest land. The mean east-west width between latitudes 63°N and 81°N is 920 km (460 miles), and the mean north-south length between 24°W and 60°W is 1370 km (860 miles).

Surface areas

Areas were measured[2] by planimetry of 15 sheets of the World Aeronautical Chart, scale 1:1,000,000.

	km²	Miles²
Area of Greenland, including its islands	2,186,000	843,800
Area of the ice sheet (inland ice, ice cap)	1,726,400	666,400
Area of isolated peripheral glaciers (examples, Flade Isblink, Sukkertoppen)	76,000	29,300
Area of ice-free land	383,600	148,100

Thus 5/6 of Greenland is covered by glaciers, 4/5 by the main ice sheet. In elevation, the area of the ice sheet is distributed as follows:[2]

Elevation above sea level		Area		
(ft)	(meters)	(km²)	(mile²)	(%)
0-1000	0-305	27,000	10,400	1.6
1000-2000	305-610	37,200	14,400	2.2
2000-3000	610-915	64,000	24,700	3.7
3000-4000	915-1220	93,400	36,000	5.4
4000-5000	1220-1525	124,000	47,900	7.2
5000-6000	1525-1830	174,000	67,300	10.1
6000-7000	1830-2135	237,300	91,600	13.7
7000-8000	2135-2440	290,000	112,000	16.8
8000-9000	2440-2745	310,500	119,900	18.0
9000-10,000	2745-3050	254,200	98,100	14.7
10,000-10,800	3050-3290	114,200	44,100	6.6
		1,726,400	666,400	100.0

The mean elevation of the ice sheet is 7000 ft (2135 m).

Thickness

There is a good seismic coverage of the area south of latitude 73°N by the French expeditions, and of a cross-section strip between 76°N and 79°N by American and British expeditions. The French maps[2] given here (Fig. 2, 3) make use of a single profile at 79°N, and would not be significantly changed by incorporation of the more recent thickness data. The bed may be deeper in the central area where there is no coverage.

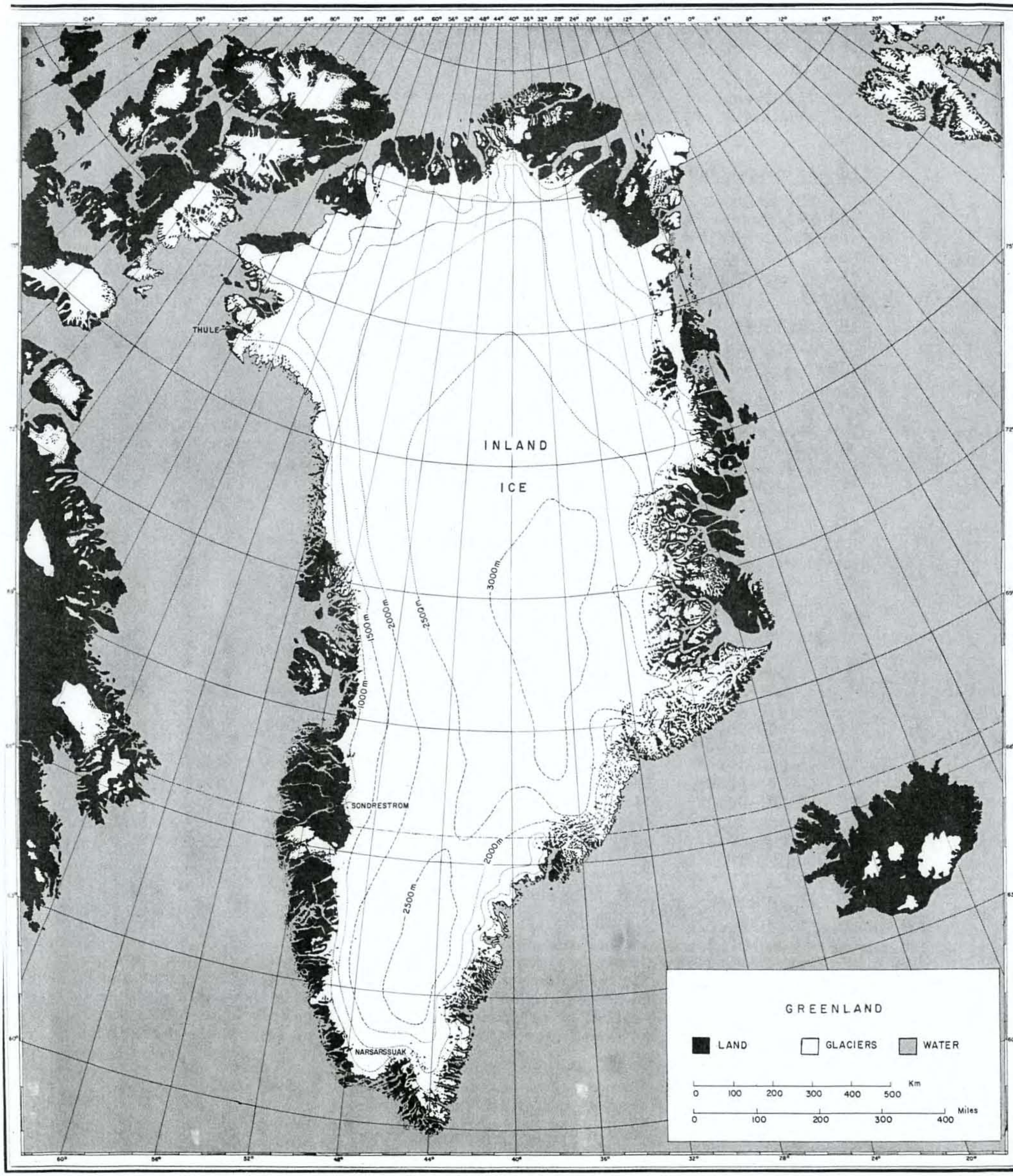

Figure 1. The Greenland ice sheet (CRREL edition of map by P. E. Victor).

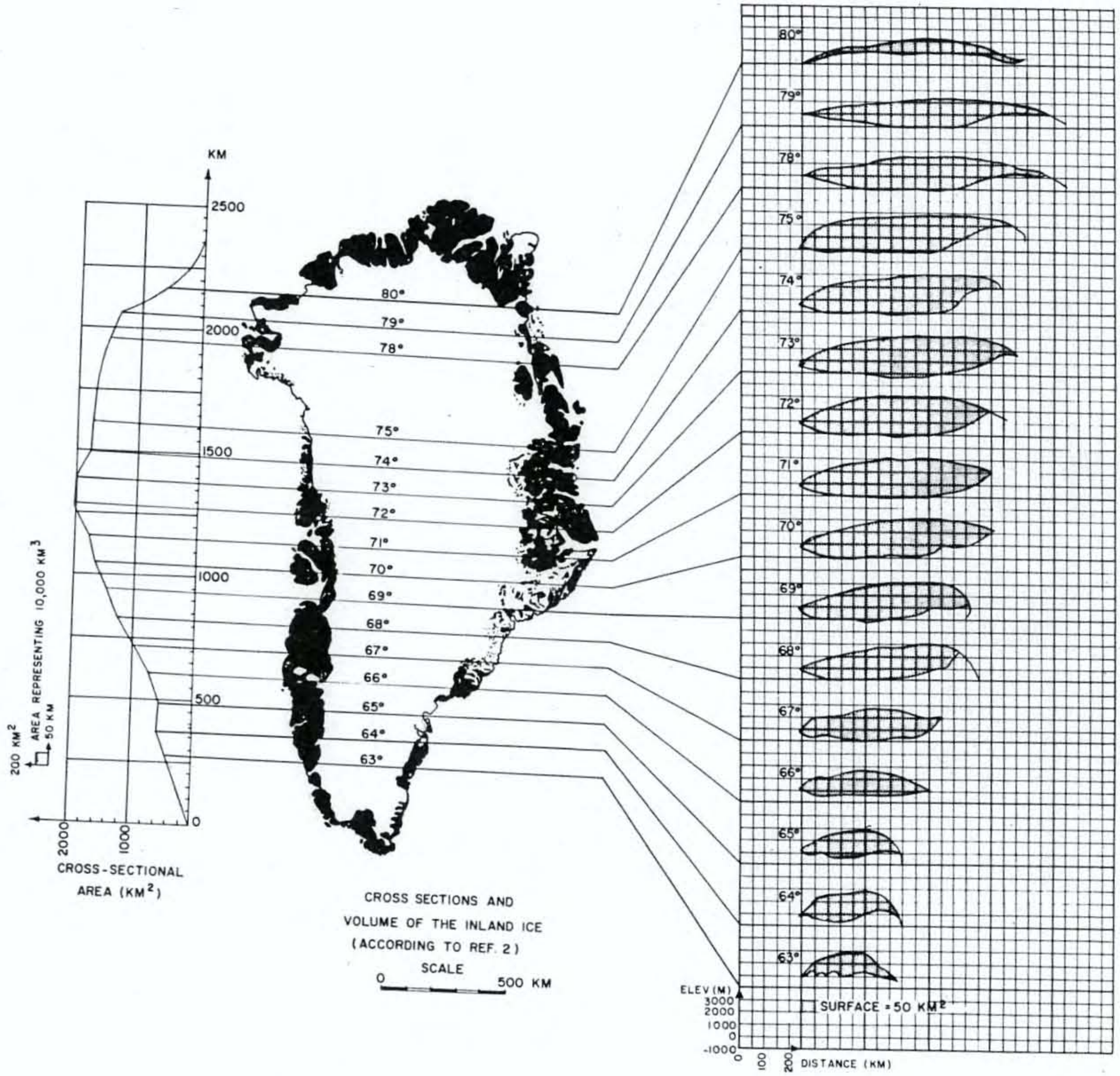

Figure 2. Cross sections and volume of the inland ice (From Holtzscherer, 1954, Fig. 19).

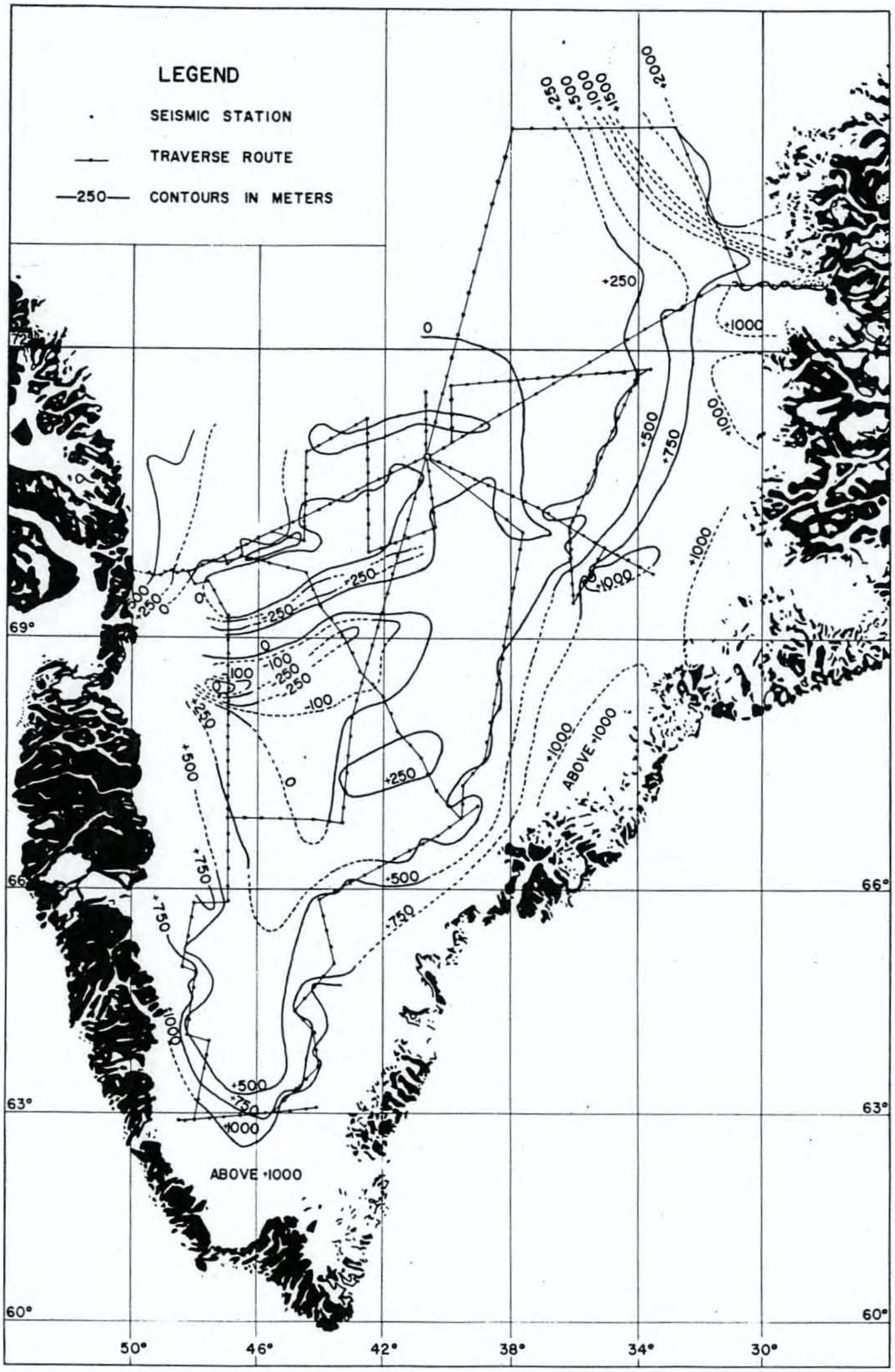

Figure 3. Map of bed of southern part of the inland ice (From Holtzscherer, 1954, Fig. 18).

The areas of the base of the ice sheet show the following distribution in elevation.[2]

Elevation above sea level		Area		
(ft)	(meters)	(km^2)	($mile^2$)	(%)
-1300 to 0	-400 to 0	529,000	204,000	31
0-3300	0-1000	632,000	244,000	37
3300-6600	1000-2000	419,000	162,000	24
6600-9900	2000-3000	123,000	47,000	7
9900-12,000	3000-3700	17,000	7,000	1
		1,720,000	664,000	100

The values should not be considered precise, being based on very sparse seismic coverage over large areas.

The mean elevation of the bed of the ice sheet is 620 m (2060 ft), and almost $\frac{1}{3}$ of its area lies below sea level. The mean thickness of the ice sheet (equal to the difference between the mean elevations of surface and bed) is 1515 m (4970 ft). The ice volume is 2,600,000 km^3 = 620,000 $miles^3$. The water equivalent is 2,350,000 km^3 (560,000 $miles^3$). If distributed over the world's oceans, it would raise the sea level by 6.5 m (21 ft). The maximum thickness of the ice sheet is about 3000 m (10,000 ft).

REGIMEN

Precipitation (Accumulation)

For practical purposes, all the precipitation on the ice sheet is in the form of snow. Only summer rain very near to the edge can run off immediately. Annual accumulation and precipitation can be equated when considering large areas. The magnitude of inequality of precipitation and accumulation caused by snow drifting is not known but is unlikely to be very large; the effect is to remove snow from the interior downslope towards the edge.

The contour map of accumulation (Fig. 4) was drawn from data gathered largely by SIPRE* Expeditions[3,7,9]. The reader is warned that the isohyets (lines connecting points of equal precipitation) are likely to be inaccurate in many places, particularly to the east of 40° meridian.

Table I is the result of measuring individual areas and calculating the accumulations. The mean precipitation is 36.7 cm water equivalent over the whole area of the ice sheet, corresponding to a volume of 636 km^3 of water equivalent. Figure 4 and Table I are interesting. It appears (the values are probably not very accurate) that the southern part (south of 66° lat which cuts through the saddle between the two domes), with $\frac{1}{8}$ of the area, receives almost $\frac{1}{4}$ of the snowfall. The region north of 78°, 1/5 of the whole area, receives only 1/10 of the snowfall. The mean temperature at 10 m depth is about -14C in the southern area and about -24C in the northern. In the northern area the slopes are less steep than in the south. This all points to a highly dynamic situation in the south, against a relatively static situation in the north.

Latitude $73\frac{1}{4}$°N bisects the area of the ice sheet. The colder northern half receives $\frac{1}{3}$ of the precipitation; the warmer southern half receives $\frac{2}{3}$.

The existence of the saddle between the south and north domes could be partly due to low accumulation in the area.

Long-term accumulation rates have been determined by stratigraphic studies at only one point (Site 2: 77°N 56°W). Because of the difficulty of locating layers exactly one year apart, the annual values given in Figure 5a can have large errors, but the 5-year means are fairly accurate.

Figure 5b is more informative. The residual mass curve rises when the annual accumulation is greater than the mean (40.7 cm water equivalent between 1885 and 1954), and falls when accumulation is less than the mean. The curve shows that the rate of accumulation was predominantly low from 1885 to 1890 (6-year mean 38.2 cm),

* U. S. Army Snow Ice and Permafrost Research Establishment, Corps of Engineers, now the Cold Regions Research and Engineering Laboratory (CRREL).

Figure 4. Mean annual accumulation.

Table I. Distribution of annual accumulation on the Greenland ice sheet.

Annual accumulation in cm of water equivalent	North of 78°N latitude		Between 78°N and 75°N latitude		Between 75°N and 72°N latitude		Between 72°N and 69°N latitude		Between 69°N and 66°N latitude		Between 66°N and 63°N latitude		South of 63°N latitude		Total area		Total annual accum.	
	Area km²	Accum. km³	Area km²	Accum. km³	Area km²	Accum. km³	Area km²	Accum. km³	Area km²	Accum. km³	Area km²	Accum. km³	Area km²	Accum. km³	km²	%	km³	%
10-15	93,800	11.7	95,600	12.0	-	-	-	-	-	-	-	-	-	-	189,400	11	23.7	3½
15-20	64,700	11.3	73,300	12.8	44,300	7.8	2,400	0.4	-	-	-	-	-	-	184,700	11	32.3	5
20-30	140,500	35.1	101,900	25.5	102,300	25.6	63,300	15.8	20,500	5.1	-	-	-	-	428,500	25	107.1	17
30-40	13,800	4.8	22,800	8.0	78,100	27.3	80,500	28.2	103,800	36.3	35,200	12.3	-	-	334,200	19	116.9	18½
40-50	3,800	1.7	30,500	13.7	35,700	16.1	67,100	30.2	53,800	24.2	32,400	14.6	-	-	223,300	13	100.5	16
50-60	3,300	1.8	24,300	13.4	30,500	16.8	50,900	28.0	15,200	8.4	20,900	11.5	1,400	0.8	146,500	8½	80.7	12½
60-70	-	-	11,900	7.7	7,100	4.6	10,000	6.5	10,400	6.8	17,600	11.4	5,700	3.7	62,700	3½	40.7	6½
70-80	-	-	-	-	-	-	6,200	4.7	13,300	10.0	11,000	8.3	12,800	9.6	43,300	2½	32.6	5
80-90	-	-	-	-	-	-	5,200	4.4	20,000	17.0	12,400	10.5	24,800	21.1	62,400	3½	53.0	8½
90-100	-	-	-	-	-	-	3,300	3.1	10,500	10.0	15,700	14.9	21,900	20.8	51,400	3	48.8	7½
Total area %	319,900 18½		360,300 21		298,000 17		288,900 17		247,500 14		145,200 8½		66,600 4		1,726,400	100		
Total accum. %		66.4 10½%		93.1 14½%		98.2 15½%		121.3 19%		117.8 18½%		83.5 13%		56.0 9%			636.3	100%
Mean accum.		21cm		26cm		33cm		42cm		48cm		58cm		84cm			36.7cm	

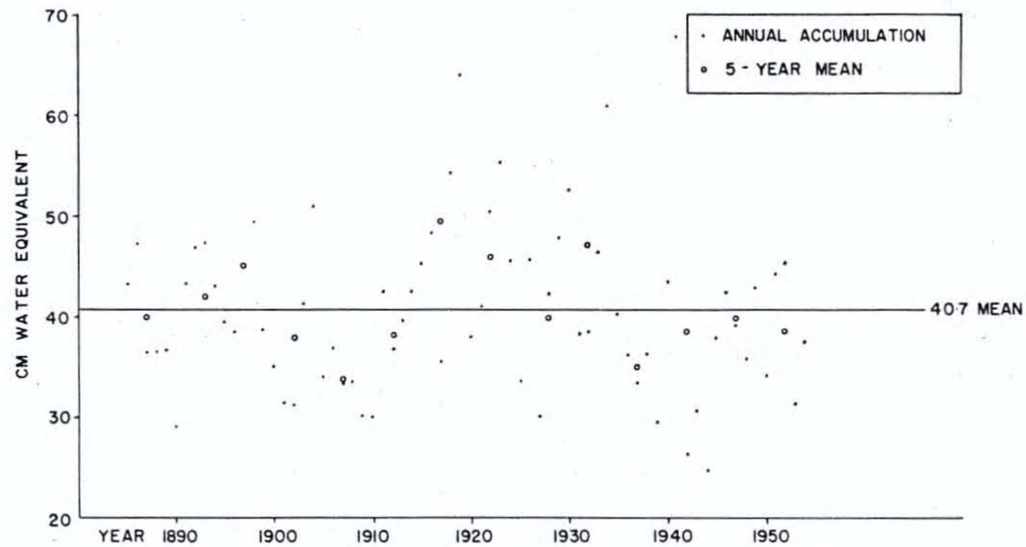

Figure 5a. Annual accumulation at 77°N 56°W.

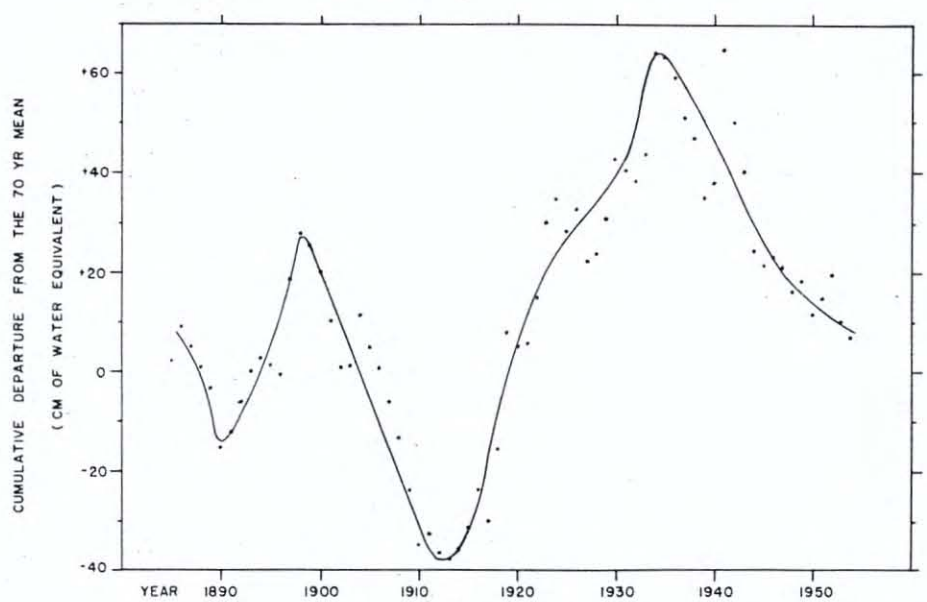

Figure 5b. Residual mass curve for accumulation at 77°N 56°W.

from 1899 to 1913 (15-year mean 36.3 cm) and from 1935 onward (20-year mean 37.8 cm). Rate of accumulation was high from 1891 to 1898 (8-year mean 46.0 cm), and 1914 to 1934 (21-year mean 45.5 cm). The general decrease in accumulation since the middle thirties has been emphasized previously.[4] Accumulation trends over hundreds of years should soon become available from analysis of cores from deep drill holes made by USA SIPRE (now USA CRREL).

Ablation

The loss of mass by runoff of melt water and by calving off of icebergs is very difficult to estimate. Recent estimates[2,3] are 315 km^3 and 272 km^3 loss by water runoff, based on loss of 110 cm^8 and 107 cm^3 over ablation areas of 286,600 km^2

and 254, 600 km^2 respectively. Loss by calving is estimated[2] at 215 km^3. A new appraisal can be made using Figures 4 and 6. Figure 6 is based on a glacier surface area of 254, 600 km^2 lying below the firn line. This was reduced somewhat by lowering the firn line in the south.

Areas below the firn line and annual precipitation thereon

	Area (km^2)	Precipitation (km^3)
North of 78°	51, 700	13.9
Between 78° and 75°	28, 600	8.6
Between 75° and 72°	25, 600	8.0
Between 72° and 69°	36, 600	19.2
Between 69° and 66°	54, 200	34.2
Between 66° and 63°	26, 000	16.9
South of 63°	24, 100	21.6
	246, 800	122.4

The firn line, which in a temperate glacier separates the areas of net accumulation and net ablation, does not do so in a cold polar glacier. A portion of the melt water here refreezes at the base of the snow cover, with the result that much of the bare ice exposed at the end of the summer below the firn line is superimposed ice, which represents net accumulation. Thus a part of the 122 km^3, which is the water equivalent of the annual snowfall on the area below the firn line, is retained in the form of superimposed ice. But, in the areas near the edge, the superimposed ice plus older ice is melted after the snow has gone. There is also significant runoff from areas above the firn line. It has not been estimated.

Measurements[11] indicate that north of 76° the melt may not remove all of the year's accumulation below the firn line, but there is little doubt that it more than does so in the south, so that 120 km^3 is a fair minimum value for ablation by melt, while the previously mentioned estimate of 315 km^3 is probably too high, being based on an ablation area of 286, 600 km^2, which is now known to be on the high side. Adding the 240 km^3 of estimated iceberg discharge, we obtain a small or a large over-all positive mass budget depending on the chosen values. There remains no doubt that the northern half of the ice sheet is gaining mass, while the southern half may be gaining or losing. It is easy to demonstrate that, in the north, ice flow does not remove more than a small fraction of the volume added by accumulation, so that the volume of the ice sheet is increasing although the area is decreasing. Consider, for instance, a vertical cross section from the edge of the ice sheet at Thule to the crest 700 km to the east. The mean accumulation along the line is at least 25 cm of ice, a total of 175, 000 m^2, which must be removed by ice flow if the glacier is to be stationary (not change its shape). This means that, for instance, an ice cliff 175 m thick should be advancing over the land at a rate of 1000 meters per year all along the edge of the ice sheet in this region. The 7-km wide Moltke Glacier removes the accumulation from a strip only two or three times its own width. Such large fast-moving calving glaciers are few and far apart along the ice edge.

In summary, the mass budget of the Greenland ice sheet is estimated as follows:

Accumulation:	630 km^3
Ablation:	120 to 270 km^3
Loss by iceberg calvings:	240 km^3
Annual gain in mass:	120 to 270 km^3

Previous estimates[2,3] indicated a net annual loss of mass. This is unlikely to be the case, unless iceberg calving has been greatly underestimated. It must be admitted, however, that the errors in estimating the mass budget parameters can be large.

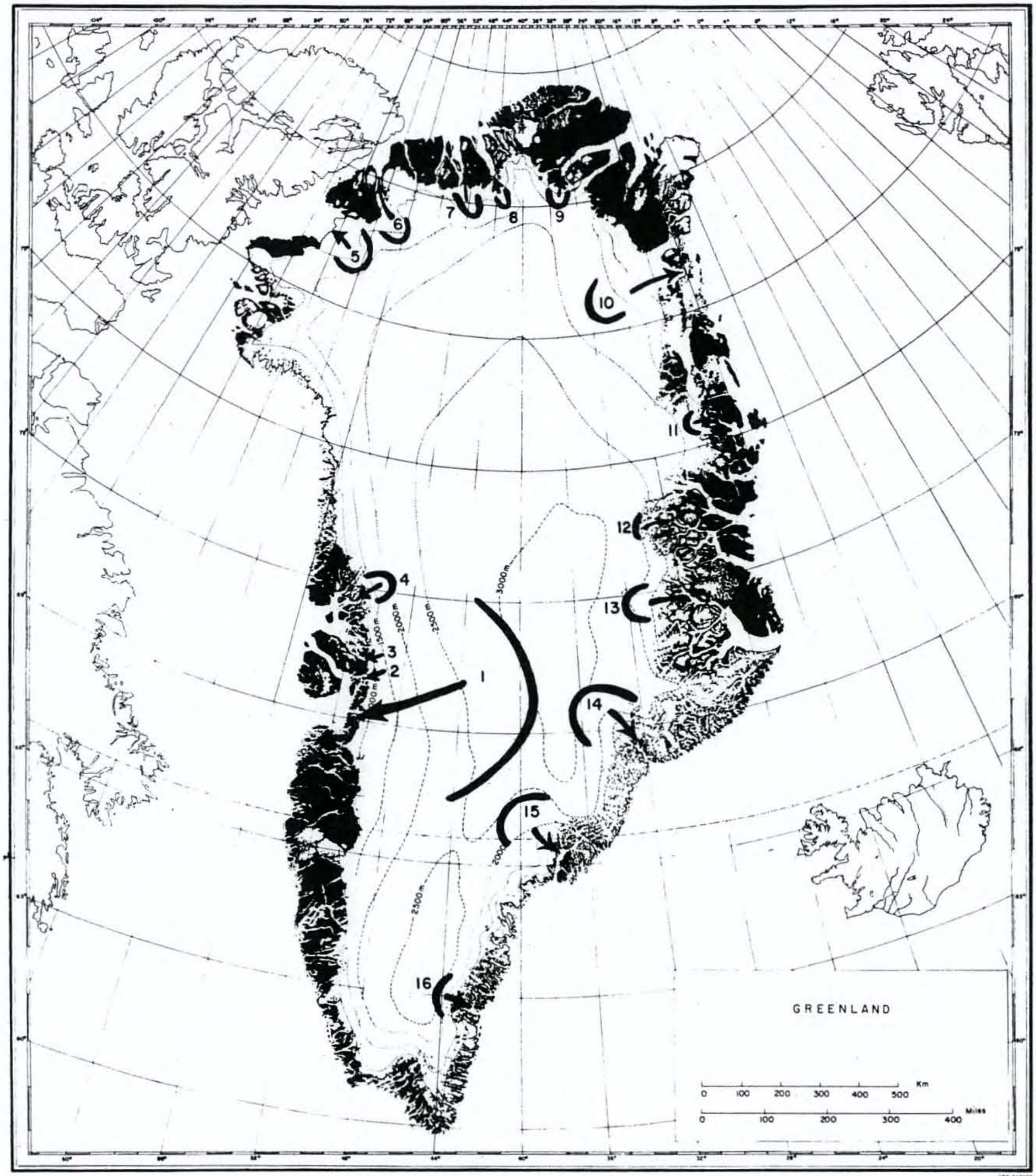

Figure 6. Drainage basins of the ice streams (From Bauer 1954, Fig. 40).

Glaciers

1. Jakobshavn
2. Torssukatak
3. Grand Karajak
4. Rink and Umiamako
5. Humboldt
6. Petermann
7. Ryder
8. C. H. Ostenfeld
9. Academy
10. Zachariaes and Nioghalvfjerd
11. Heinkel
12. Hamberg
13. Daugaard Jensen and West
14. Kangerdlugssuaq
15. Sermilik (Helheim)
16. Tingmiarmiut

Ice streams

An "ice stream" is something akin to a mountain glacier consisting of a broad accumulation basin and a narrower outlet valley glacier: but a mountain glacier is laterally hemmed in by rock slopes, while the ice stream is contained by slower moving surrounding ice. The edges of the ice stream are often crevassed, and the surface tends to be concave as the ice is "funnelled" down. Many of the large outlet glaciers in Greenland, particularly in the south are the narrow outlets of large ice streams which reach back many scores of miles into the ice sheet. The largest ones flowing into Disko Bay have corresponding depressions in the glacier floor. An interesting question is whether the ice stream makes the depression, or vice versa; and an interesting hypothesis is that the ice stream, once started, is self-perpetuating because its ice mass, warmed up by heat of internal friction, has a lower viscosity than the surrounding ice.

Figure 6 shows the main ice streams.[2] It is estimated that eight ice streams south of latitude 72°N annually discharge 120 km^3 of icebergs, which is one-third of the total accumulation of this southern half of the ice sheet. It has been pointed out[4] that the high north-south crestline of the ice sheet lies far to the east of the area of maximum accumulation. Between latitudes 69°N and 75°N, the area lying to the west of the crest (363,000 km^2) receives an annual accumulation of some 140 km^3 (mean 39 cm water). If this part of the ice sheet is to remain stationary, the major portion of this large volume of ice must drain out through the ice streams of Jakobshavn, Torssukatak, Grand Karajak, Rink and Umiamako The ice streams of the great northern glaciers, for instance the Humboldt, move more slowly and remove less ice. The estimate[2] is for a total of only 10 km^3 north of 78°N.

Variability

Table I shows that the mean rate of accumulation increases from north to south. The rate of increase is 2.5 cm water equivalent for each 1° of latitude from 80°N to 63°N. Then there is a sharp increase at the south tip of the ice sheet. It is evident that the flow picture of the ice sheet must be quite complicated owing to the great regional variations in thickness, temperature, slope and rate of accumulation, and it is further disturbed by the existence of the ice streams. Since all these factors are also temporal variables, one must conclude that the Greenland ice sheet is a mass of ice in a constantly mobile state of local readjustment to ever-changing local conditions. At present the whole ice sheet seems to be gaining mass, while the total area is shrinking. In comparison to Greenland, the much colder and much larger Antarctic ice sheet should be changing much more slowly.

TEMPERATURES

Near the surface

Figure 7 is a map of mean temperature of the upper snow layers. Where melting is insignificant, the mean snow temperature is within 1C or 2C of the mean air temperature. In the peripheral zone of net ablation by melting, the mean ice temperature must be lower than the mean air temperature, because the ice surface temperature cannot exceed 0C. In the percolation zone, where melt water refreezes to form superimposed ice or ice layers in the firn, the relation between air temperature and firn temperature is not known, because there are no long-term measurements and no theory. If the calculated[3,4] air temperatures on the southern dome are correct, then the snow temperatures measured during the SIPRE expedition of 1959 show that here the snow temperatures are also lower than the mean annual air temperature. A difference of 1C-4C is indicated.

The warming up of the Arctic (perhaps mainly higher summer temperatures) during the last few decades is clearly evident in the ice sheet. The stratigraphic record from the deep pit and drillhole at 77°N 56°W shows much evidence of summer melting back to about 1920, but almost no ice-layer formation during the preceding several hundred years. The temperature measured[5] in the drillhole shows a decrease with depth, which is also indicative of a warming up of the surface. The temperature is 1C lower at 300 m depth than near the surface. At Eismitte Station Centrale (71°N 40°W) the warm-up between 1930 and 1950 appears to have been 1C. Coastal stations

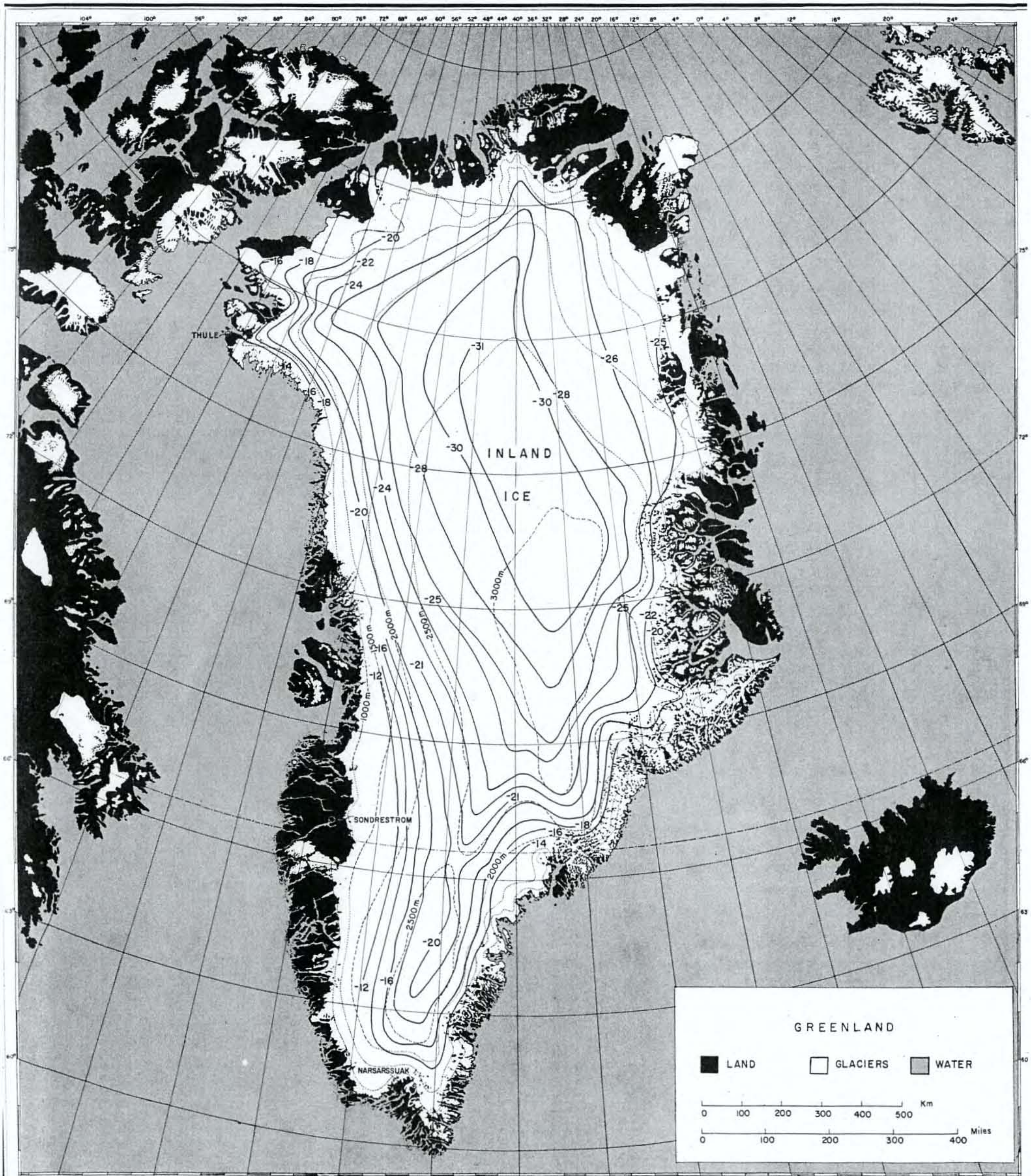

Figure 7. Mean annual snow temperatures.

south of 70°N show an increase of from 1/2C to 3C of the mean air temperature from the period 1907-1920 to the period 1921-1940.[3,4] The ice sheet seems to be somewhat less affected than the coastal land.

Internal

The energy of flowing is converted into heat at the rate of 2.34×10^{-3} cal per gram per meter drop in elevation. The corresponding temperature increase is 0.47C per 100 meters. Now consider an idealized ice sheet: the surface temperature is equal to the air temperature; the atmospheric lapse rate is equal to the above-mentioned 0.47C per 100 m; the heat from the flow energy is evenly distributed throughout the whole ice mass; and there is no earth heat flow into the ice. The ice will then reach the coast with a temperature equal to the mean air temperature there.

In Greenland only the very southern tip, south of 62°N, would show wet ice: the ice sheet would be cold (i.e., temperature below the melting point). In comparison to this model, the real Greenland ice sheet is colder, because

a) the lapse rate is higher (0.6 to 0.8)

b) near-surface snow or ice temperatures in the melt zone are lower than the mean air temperatures

c) it is probable that most of the flow energy is liberated near the bed.

This means that the ice near the surface is colder than the atmosphere and absorbs heat from it. Thus a negative temperature gradient does not necessarily indicate climatic amelioration.[10]

The earth heat flow of some 40 cal/cm^2 per year can escape through the ice only by conduction if the temperature gradient at the bottom is greater than 2.5C per 100 m. If the gradient is smaller, the bottom of the ice sheet must be wet, and since wet (isothermal) ice does not conduct heat, only a very thin layer can be wet from geothermal heating. A somewhat thicker layer can become wet from internal generation of frictional heat, but it is unlikely to be thicker than a very small fraction of the glacier thickness. A number of icebergs in the fjord at Narssarssuak near the south tip of Greenland were inspected, and none showed any trace of the large ice crystals usually found in bottom wet ice.

The conclusion is that the Greenland ice sheet is cold, except for a very thin bottom layer. It is hoped that very deep core drilling will soon partly clear up the problem of internal temperatures.

In the Thule area, as proved by core drilling, the ice near the edge is frozen to the bed: in this case there is no erosion under 200 ft of ice. An interesting question is whether there can be any erosion under thick ice if the bed is frozen.

SURFACE MOTION

Motion near the edge of the ice sheet has been measured in a number of locations, but since it is completely controlled by the greatly variable local conditions, it throws almost no light on the motion of the ice sheet as a whole. In the interior, where there are no visible fixed landmarks, the determination of motion is difficult and expensive. Accurate "Hiran" fixes were made at six stations in 1956, but are unlikely to be repeated. A fairly large number of astronomical fixes with varying degrees of accuracy have been made, but not yet repeated at sufficiently long time intervals to yield good motion data.

Presently, the best estimate is that surface points in the interior of the ice sheet move with velocities of no more than tens of meters annually.

This subject of motion of the ice sheet is not yet ripe for useful summarizing.

SURFACE CONDITIONS

Bare ice is exposed near the edge of the ice sheet, where all the snow melts away, and in the summer there is also a peripheral strip of wet snow. Saturated slush forms on flats, which are quite extensive, for instance, in the area east of

Sondre Stromfjord. Here slush fields are found as far as 100 km from the ice edge. Most of the peripheral zone is crevassed, but there are few crevassed areas more than 100 km inland, and perhaps none more than 200 km from the nearest land. Most of the common morphological features of glaciated areas can be found in Greenland, but those associated with morainal material are not prominent. Glacial erosion appears to be generally weak, and most of the land area is only thinly covered with debris.

The interior of the ice sheet, $\frac{1}{2}$ to $\frac{2}{3}$ of its area, shows only very surficial summer melt features. It is a monotonous snow landscape with generally low slope gradients. Here there are two interesting morphological features awaiting explanation.

The first is that on many routes one notes, on traveling inland, a succession of relatively steep rises tens of miles apart alternating with relatively flat areas. The second is the wavy surface of the high interior. The wave length is several kilometers and the amplitude several meters. It is unlikely that either of these features is controlled by the shape of the bed. The first may be in part correlatable with accumulation rates, more probably it will have to be explained in terms of ice flow mechanics. The second must be a snowdrift phenomenon.

The snow surface changes from day to day with the weather; strong winds can produce hard sastrugi, very seldom higher than 1 ft. Surface snow density varies from about 0.2 to 0.4 g/cm^3, essentially the same as winter snow elsewhere, except that very light snow is rare because snowfall is usually associated with wind. High winds are rare in the interior at most stations: it blows in excess of 20 mph only some 5% of the time. But very strong winds up to 120 mph in gusts occur in North Greenland at the edges of the ice sheet. Four facies (geological term for different aspects of a formation) of upper layers can conveniently be distinguished[3] (Fig. 8, 9).

1. The ablation facies - extends from the edge of the ice to the firn line. Above the firn line, there is a permanent snow cover, i.e., the snow does not all melt away. Bare ice characterizes the area of the ablation facies, which covers no more than 15% of the ice sheet area.

2. The soaked facies - extends from the firn line to the saturation line. Summer melting produces sufficient water to soak at least the whole snow accumulation of the past year. Thus all the snow of the soaked facies has been wet at least once. Average density of the upper 5 meters is greater than 0.50 g/cm^3. The soaked facies belt covers about 10% of the ice sheet area.

3. The percolation facies - extends from the saturation line to the dry snow line. Summer melting is insufficient to soak all of the past year's accumulation. Percolation in pipe-like vertical channels can, however, penetrate to and past the previous summer layers and spread to form ice lenses and layers in older snow. The average density of the upper 5 meters varies from 0.43 to 0.39 g/cm^3. The percolation facies covers some 45% of the ice sheet area.

4. The dry-snow facies - is characteristic of the whole area lying above the dry-snow line. Melting and percolation are negligible; the snow remains permanently dry, and the average density of the upper 5 meters is less than 0.375 g/cm^3. Snow of the dry facies covers 30% of the ice sheet area.

Figure 10 shows a pit dug in the dry snow (dry-snow facies). The pit walls have been brushed to reveal the succession of hard and soft layers.

SNOW DENSIFICATION

Figure 11 shows a typical 10-meter snow profile[1] at 77°N 56°W near the upper limit of the percolation facies. There are almost always some low-density layers deposited in the summer, and down to about 7 meters the general increase in density is almost linear with depth. The slope of the depth-density line varies with location. The smaller the rate of accumulation, and the higher the temperature, the greater the density increase over a given depth increment.

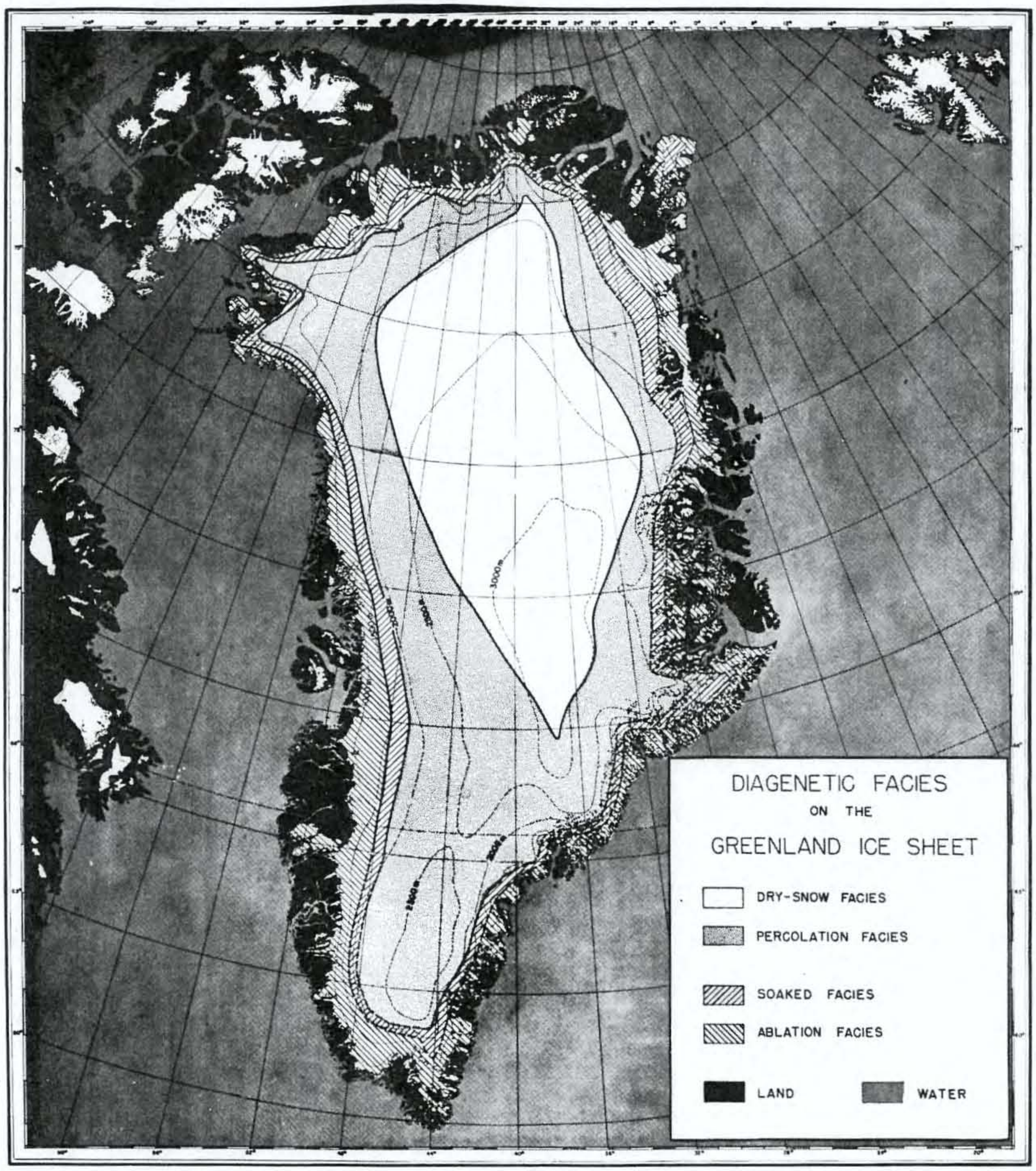

Figure 8. Distribution of facies (Benson, 1960, Fig. 48). The location of facies boundaries is most certain along the west side of the crest between 66 and 79°N. Away from this region, the dry-snow line may be extrapolated with confidence because it is independent of accumulation. The dry-snow line on this map was drawn along the intersection between a plane sloping 1.15 m/km to the north, and the surface of the ice sheet. Location of the firn line and saturation line in south and east Greenland is highly speculative.

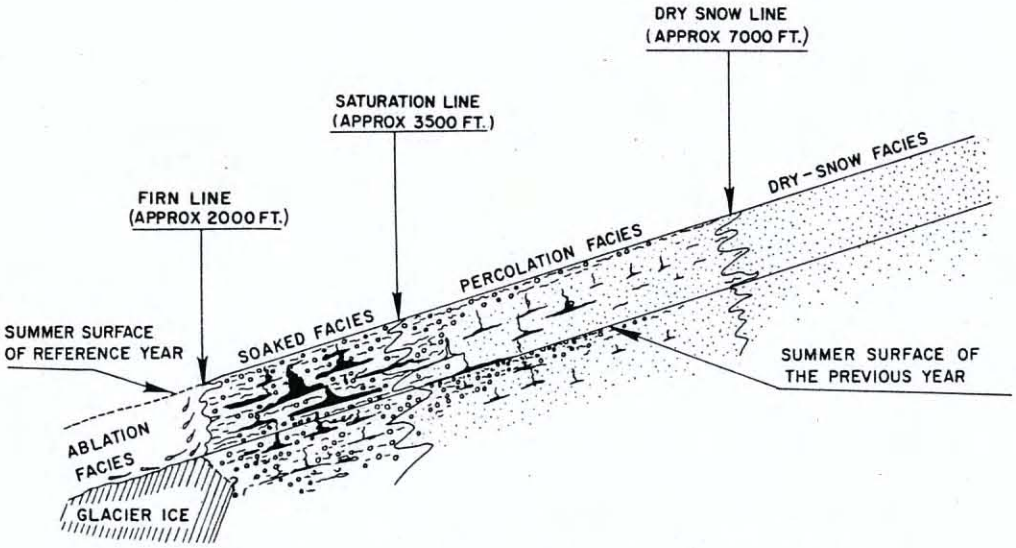

Figure 9. Schematic of facies (Benson, 1960, Fig. 15).

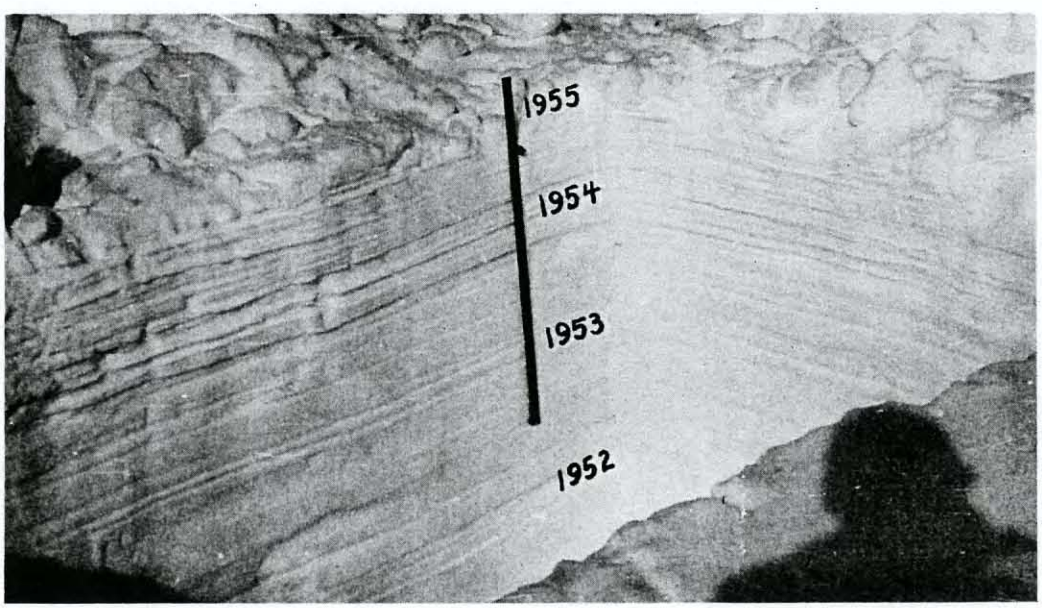

Figure 10. Pit in dry snow (Benson, 1960, Fig. 4b). The walls have been brushed to reveal succession of hard and soft layers.

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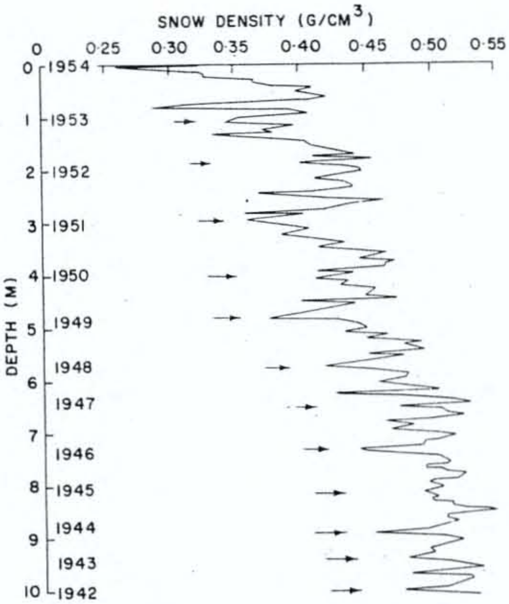

Figure 11. Snow density profile at 77°N 56°W (Bader, 1958, Fig. 4).

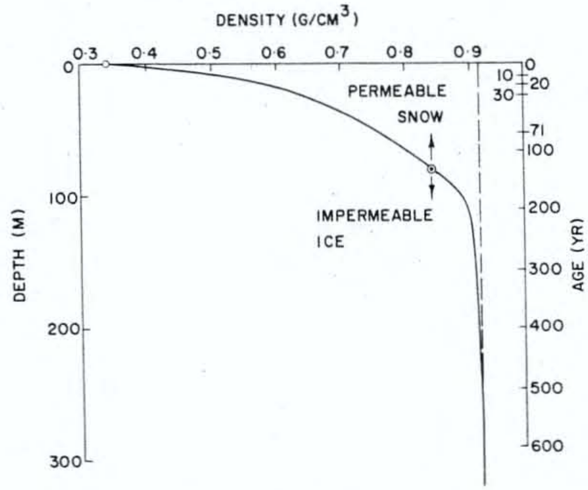

Figure 12. Depth-density curve 77°N 56°W (Bader, 1958, Fig. 3).

Figure 12 shows the increase of density with depth[1] as determined on cores from a drill hole at 77°N 56°W. At about 80 m depth. the pores in the snow have become isolated and there is impermeable ice with a density of about 0. 8 g/cm^3, and an air bubble volume of about 10%. Further densification takes place at the expense of air bubble volume, causing a rise in bubble presssure.[7] The resulting glacier ice is fine-grained (1 mm to 1 or 2 cm) and milky in appearance due to inclusion of hundreds of bubbles per cm^3.