Some Considerations of Snow Metamorphism in the Antarctic Ice Sheet in the Light of Ice Crystal Studies

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Abstract

Certain aspects of the stratigraphy and petrology of snow and firn were studied at Southice.

Contrasts in shallow lithologies have been attributed to temperature gradients and their sense. These contrasts are maintained in deeper, older firn to at least 8 m, under continuing metamorphism occasioned by vapour transfer. This metamorphism is more intense in less dense layers, a phenomenon due to local gradient differences in adjacent layers having contrasted porosity and conductivity.

Crystal cross-section areas have size distributions which are close to log normal. Quartile or moment measures can be used in studying variations in the nature of size distributions. Some characteristics change progressively with depth.

Average crystal size shows a general increase with depth from 0.006 mm² at the surface to more than 2 mm² at 45 m. This increase is approximately linear except in a surface zone less than 1 m thick where growth rate is transitional and is most extreme next to the surface. Size increases 10 times within even the first centimetre.

Although shallow and deeper firn at Southice commonly show a very strong vertical grain structure like a string of beads, petrofabric analyses show almost random orientation. Important exceptions are wind-crusts and depth-hoar layers.

While crystal size is approximately linear with depth, so too is diameter squared with age, a relationship well known for grain growth in metals and ceramics. Ice crystal growth (under essentially isothermal and undeforming conditions) approximates to:

$$D^2-D_0^2=At\exp(-B/T).$$

The constant B has been evaluated from various observations elsewhere in Antarctica and in Greenland.

I. Introduction

The inland Antarctic station *Southice* was occupied by the Commonwealth Trans-Antarctic Expedition during most of 1957. As a member of the 3 man team operating this station, the author carried out glaciological studies beneath the snow surface to supplement those of Lister (1960) above the surface. The sub-surface work, carried out to a depth of 45 m, included detailed studies of stratigraphy and the petrology of snow and firn. Some preliminary results of this work have already been published (Stephenson and Lister, 1959). A full account is in preparation.

Southice was situated at 81° 57′ S, 28° 50′ W, and an altitude of 1 370 m (4 430 ft). It was 416 km (260 miles) from the coast and 45 km (28 miles) south of the Whichaway Nunataks. Although a nunatak was present only 5 km north-west from the station, this was a very small, low outcrop and *Southice* occupied a very open situation. Its conditions should be well representative of this part of the Antarctic Ice Sheet.

Seismic investigations carried out by Pratt (1960) indicate an ice thickness at Southice of 540 m (1 770 ft). Lister (1960) observed the mean accumulation at stakes from late April until late December 1957 to be 8 cm (water equivalent) and stratigraphic studies suggested that average annual accumulation had been about 10 $g \cdot cm^{-2}$ over the previous decade. The mean annual temperature (as indicated by bore hole temperatures) is close to $-31°C$. Meteorological observations made during the occupation of Southice are summarised and discussed by La Grange (1963) and Lister (1960) has described his own detailed studies of precipitation and drift snow.

II. Methods

A detailed description of the snow stratigraphy was made in a pit to a depth of 13 m. This involved thickness measurements, density determinations of individual strata thicker than 3 cm and visual estimation of grain size (maximum diameter present, minimum and most common diameter). Below 13 m cores were obtained by SIPRE drilling from the pit floor. Density measurements were made of the cores in sections about 10 cm long, without any detailed description of individual strata.

The texture and crystal fabric of selected material was studied in thin sections with a polarising microscope. The sections were prepared by filling and cementing on microscope slides with diethyl phthalate, grinding on abrasive paper, and final dissolving of the cement in tetralin. Measurements of individual crystals were made using a micrometer eyepiece. In each sample the areas of 50 to 200 crystals were represented by length and breadth. Crystal orientations were measured with a 4 axis universal stage on the microscope. The stage hemispheres were not used and the tilt angles were corrected from the relevant empirical curves (for *c*-axis vertical, and for *c*-axis horizontal) according to Rigsby. For the petrofabric work, between 100 and 200 crystals were studied in each specimen. Specimens from the pit walls were easily orientated and with care, specimens from the base of the drill hole could be orientated relative to the position of the drill handle before being broken off by lifting. After raising the drill to the surface and restoring the initial position of the handle, the core could be marked with an appropriate magnetic direction accurate to about 5°.

The petrology studies of shallow snow were made on material from a second pit located several hundred metres from the station. This precaution was taken against the likelyhood of disturbed shallow material closer to the station.

III. Results

Snow facies and lithologies. Surface melting does not occur at Southice, as indicated by the absence of ice layers. The locality is well within the zone of dry snow facies (Benson, 1962; Giovinetto, 1964).

Various snow and firn lithologies can be described under *wind crusts, granular layers* (widely varied in relative coarseness and density), *sastrugi layers* (harder, massive lenticular layers) and *depth-hoar layers.* No radiation crusts like those described by Giovinetto (1960) at the South Pole were recognised at Southice. Almost all layers at less than 7.5 m depth show a vertical structure, with the grains aligned like strings of beads. Below this depth the apparent vertical structure steadily disappears by the lateral interference of adjacent grains, the transition occurring at densities from 0.52 to 0.55 g·cm^{-3}. This is close to the "critical density" for ice grains, 0.55, this being the maximum density possible for close packing of ice spheres of equal size (Anderson and Benson, 1963).

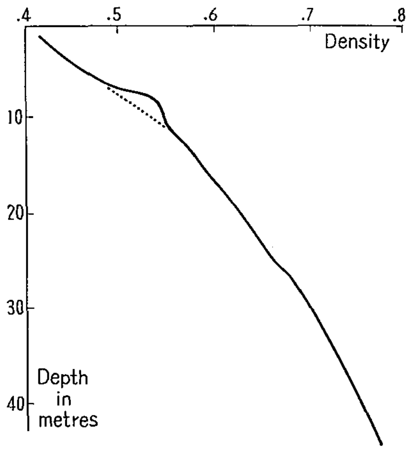

Fig. 1. The mean density profile at Southice

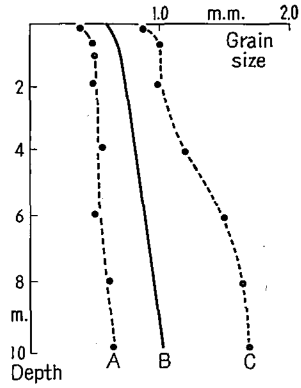

Fig. 2. Grain size features at Southice (Common grain size: A, very fine granular layers; B, average for all layers; C, coarse granular layers)

Density and grain size profiles. The full mean density profile is shown in Fig. 1 and features of the grain size profile are given in Fig. 2. Some important aspects of the profiles (not all demonstrated on these figures) should be emphasised:

1. There are general alternations of coarser less dense granular layers, and finer denser granular layers. These represent material beneath summer and winter surfaces respectively, and permit the estimation of annual layering and accumulation as near 10 g·yr^{-1}.

2. Average grain size increases most swiftly near the surface. Below 50 cm slow increase in the average occurs, with differences for various lithologies. The largest increase occurs in the coarser (and less dense) layers, and these layers also become apparently thicker especially in the region 4–6 m. By contrast, the grain sizes of finer (and denser) layers show much smaller change with depth. The general increase in grain size is similar to that described by Lister (1961) at Northice in Greenland.

3. Depth-hoar layers occur at somewhat irregular intervals. They apparently develop in thickness with increasing depth to a maximum 7 cm at 580 cm, where the largest depth-hoar grain sizes occur. Below this depth, these open, loosely-textured layers collapse, producing the more rapid increase in average density seen on the profile near 7 m (Fig. 1).

Crystal size measurements. The clear distinction made by Schytt (1958 b) between grain size and crystal size has been followed. Grain size refers to snow and firn grains, whereas crystal size refers to individual crystals (seen under a petrographic microscope) several of which usually compose each grain.

Microscopic measurements were made on some specimens sectioned horizontally, and on others vertically. Nearly all the horizontal measurements were made of specimens below 1 m, and nearly all the vertical ones above this depth. The reasons for making the vertical measurements of the shallower material were the wide variation in lithology, and the opportunity to measure crystal size in several adjacent layers (or parts of the same layer) in a single thin section. Measurements suggested that there were only small differences between average crystal size in horizontal and vertical sections.

Crystal size is expressed as the average for a number of crystals. The possible error involved in this average crystal size depends on the number of measurements and the standard deviation of the individual measurements. This was analysed in a number of specimens, examples of which are given in Table 1.

Table 1

Depth	Lithology	H or V	N	Crystal size	σ	Poss. error
0.15 mm	Wind-crust	H	50	0.00652	0.00504	0.00143
"	"	V	50	0.00773	0.00575	0.00163
8 cm	Depth-hoar	H	52	0.146	0.1165	0.0325
18.2 cm	Granular lyr.	V	52	0.1185	0.0852	0.0237
29 cm	Fine gran. lyr.	V	52	0.1365	0.1018	0.0284
47 cm	Coarse gran. lyr.	V	52	0.376	0.264	0.0738
70.3 cm	"	V	52	0.214	0.143	0.0400
1.92 m	Granular lyr.	H	50	0.379	0.326	0.093
			100	0.383	0.345	0.050
			230	0.348	0.384	0.050
"	"	V	50	0.450	0.442	0.126
			100	0.428	0.507	0.100
			200	0.450	0.476	0.066
12.11 m	Gran. lyr.	H	100	1.26	1.25	0.25
20.57 m	Fine gran. lyr.	H	52	0.834	0.575	0.16
41 m	Wind-crust	H	51	3.25	2.17	0.61
45.6 m	Gran. lyr.	H	100	2.22	1.89	0.37

H or V: Horizontal or vertical section
N: Number of crystals measured
Crystal size: Average crystal area (mm^2)
σ: Standard deviation
Poss. error: $t \times \frac{\sigma}{\sqrt{N}}$

From these and other examples, possible errors in general can be assumed less than:

30% for 50 crystals,
20% for 100 crystals,
15% for 200 crystals.

Horizontal and vertical crystal sizes. In a specimen cut at 0.15 mm and one at

1.92 m both horizontal and vertical crystal sizes were measured; in each, average vertical crystal size is larger. "Students" t tests were applied to the two averages in each case to test the significance of their differences, with the following results:

0.15 mm t has a probability of more than 0.2 (1 in 5),

1.92 m t has a probability of less than 0.02 (1 in 50).

The difference in horizontal and vertical crystal sizes next to the surface is not significant, but that at 1.92 m is significant. It may be that the relative difference increases with depth.

The difference between horizontal and vertical crystal sizes at 1.92 m is surprisingly small since this granular layer had a strongly developed vertical grain structure. In comparison with the possible errors in average crystal size measurement the differences are small enough to assume that depth profiles for horizontal and vertical crystal sizes are essentially the same.

To establish the nature of size differences between horizontal and vertical sections sphericity was considered. An expression for sphericity is D/l where l is the longest dimension of the crystal and D is the diameter of the circle with area equal to the crystal cross-section. For the specimen at 1.92 m average sphericities of 100 crystals are:

Horizontal 0.894,

Vertical 0.856.

In this case, the difference in size can be attributed to sphericity. The crystals are slightly more elongate in vertical section, parallel to the vertical grain structure.

The nature of crystal size distributions. The ice crystals seen in thin section provide a random selection of cross-section areas. Their average is smaller than that of the full cross-sections of the three dimensional crystals. While it is possible to estimate spatial grain size distribution, (as is done in metallurgy by Johnson, 1946) it has been shown that the planar crystal size distribution (from random cross-sections) is representative of spatial size distribution features (Feltham, 1957), both distributions in metals being approximately log-normal. The problem of ice crystal size measurement is identical with that in metallurgy, and size distributions have therefore been considered without recalculation to spatial size in the present work.

An assemblage of ice crystals bears some analogies to particles in a clastic sedimentary rock. (Major differences exist in that diagenetic metamorphism of snow to firn has completely modified the original grains.) The parameters commonly used for statistical analysis of sediments (Krumbein and Pettijohn, 1938) can be adopted for comparison of ice crystal size distributions. The frequencies of ice crystal sizes occurring within geometrically progressive size ranges were determined. Since only 52 area measurements were made in most specimens, the frequencies were weighted for their actual size, rather than be used as simple numerical frequencies. This procedure can be compared with the weighing of different size fractions from a sand sample, rather than counting the number of grains in each fraction. Cumulative frequency curves were constructed and the parameters *median*, and *sorting*, *skewness*, and *kurtosis* coefficients were derived from graphical measurements. Examples are given in Fig. 3. (The *median* is the size at which half the sample is coarser and the other half finer. *Sorting* coefficient is an expression for size dispersion in a sample, specimens with more uniform sizes having lower coefficients.

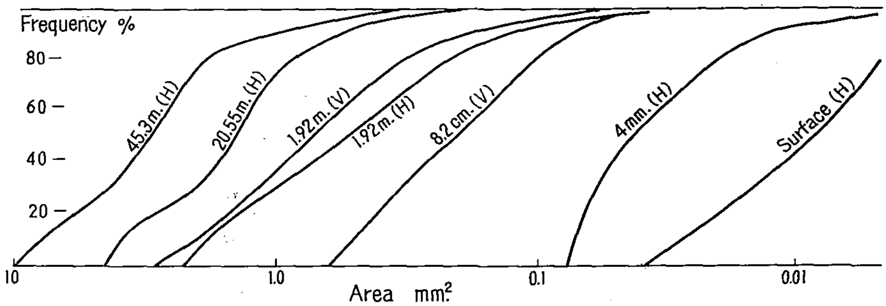

Fig. 3. Typical cumulative frequency curves for crystal areas (Frequency analysis parameters for these specimens are given in Table 2)

Table 2

Depth	Type	N*	Average	Median	Sorting	Skewness	Kurtosis
Surface	Drift	104	0.00657	0.0088	1.77	1.09	0.23
4 mm	Wind-crust	52	0.0278	0.04	1.66	0.87	0.33
8.2 cm	Depth-hoar	52	0.150	0.21	1.77	0.94	0.29
1.92 m	Gran. lyr.	230	0.348 (H)	0.56	2.13	1.01	0.28
"	" "	200	0.450 (V)	0.76	1.86	0.78	0.27
20.55 m	Wind-crust	53	1.17	2.24	1.49	1.05	0.21
45.3 m	Gran. lyr.	100	2.22	3.1	1.52	1.06	0.22

* N is the number of crystal areas measured.

Skewness involves the symmetry of the size distributions of crystals coarser and finer than the median. *Kurtosis* coefficient is an indication of the predominance of the most commonly-occurring crystal sizes.)

These parameters can be used to compare the nature of crystal size distributions in different snow and firn lithologies. Alternatively, different types of distribution could be recognized and used as a basis for classification. Analysis of results is not yet completed, but the following generalizations can be drawn for horizontal sections:

1. Distinctions between different lithologies on the basis of size distributions are not clear cut. The most characteristic lithology, wind-crust, shows better sorting coefficient than granular layers, in most cases, but there are exceptions. The average sorting coefficient for 9 wind-crusts is 1.55 ± 0.14, while the average for 16 granular layers is 1.71 ± 0.10. Other parameters do not show any apparent differences.

2. Sorting coefficient is the only parameter of size distribution which would appear to change consistently with depth. Below the surface there is an apparent increase in the sorting index, followed by a slow decrease. This is suggested by the following averages: Surface to 8.0 cm -1.66 ± 0.19 (5 specimens); 1 to 10 m -1.79 ± 0.19 (8); 10 to 21 m -1.67 ± 0.21 (6); 30 to 46 m -1.57 ± 0.20 (7). The measurements made by Fuchs (1959) in firn at Site 2 in Greenland suggest opposite changes, with sorting index continuing to increase. However, his measurements were vertical ones.

The log-normal distribution of sizes can be tested by drawing graphs on probability

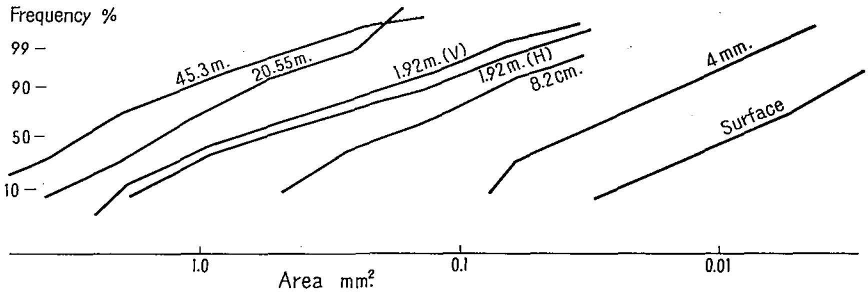

Fig. 4. Cumulative frequency graphs for crystal areas constructed on probability paper

paper. Most specimens do give essentially linear graphs (Fig. 4). These demonstrate the variety of distributions in different specimens, but show they are all approximately log-normal.

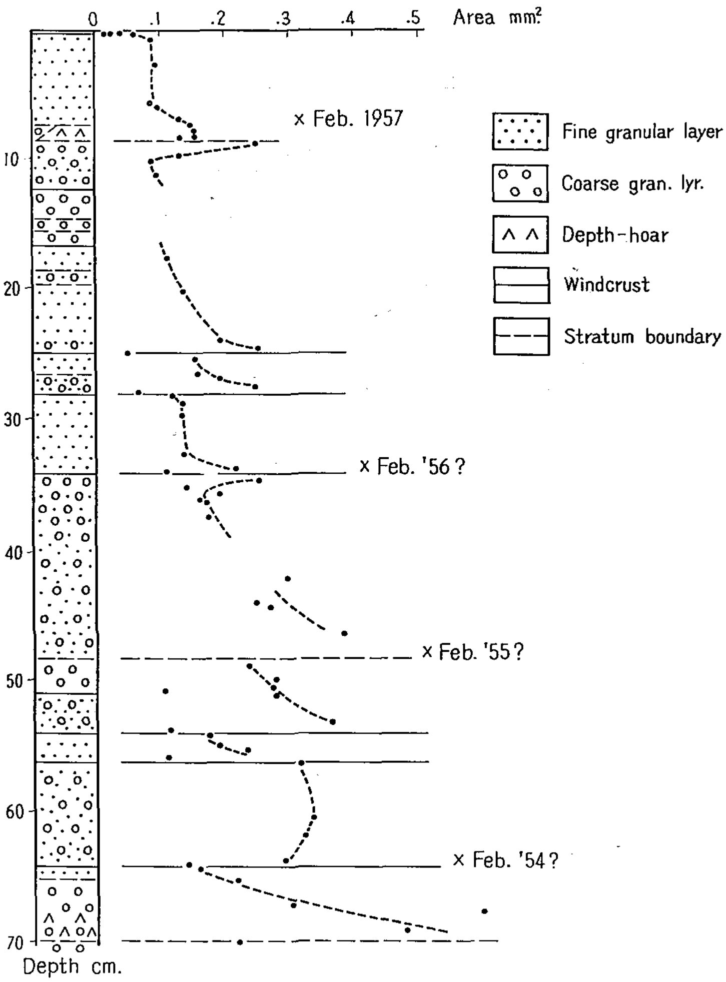

Fig. 5. Crystal sizes and stratigraphic details of shallow snow at Southice (vertical sections)

Depth profile for crystal size above 70 cm. Figure 5 shows the distribution of crystal size and other stratigraphic details in an undisturbed profile. These details were measured in samples from a pit well beyond the surface influence of the station buildings and tunnels at Southice. All the values given are for vertical sections.

The following aspects of this profile should be emphasised:

1. In the upper centimetre there is an extreme increase in crystal size (area increases 10 times).

2. Below 1 cm there is a general but much slower increase in crystal size. The most characteristic and contrasted lithologies, wind-crust and depth-hoar layer confirm this increase.

3. Certain groups of layers show consistent variations. In some, a general increase in size downwards occurs, with abrupt reversal into the following layer. In others, virtually no increase occurs, except for a rise at the base in some cases. It is suggested that those groups of layers with consistent trends are composed of a succession of deposits laid down in a relatively short time, and then subjected to metamorphism beneath the surface, without new deposition for some time. Several such developments occur within a year. The significance of the two styles of variations is not clear, but there is some stratigraphic evidence that they may have developed beneath summer and winter surfaces respectively.

4. There is no clear distinction on the basis of crystal size alone between firn layers metamorphosed beneath a summer surface ("summer layers") and those beneath a winter surface ("winter layers"). The correlation with grain-size is probably very low, judging from the wide variation of crystal size in single layers.

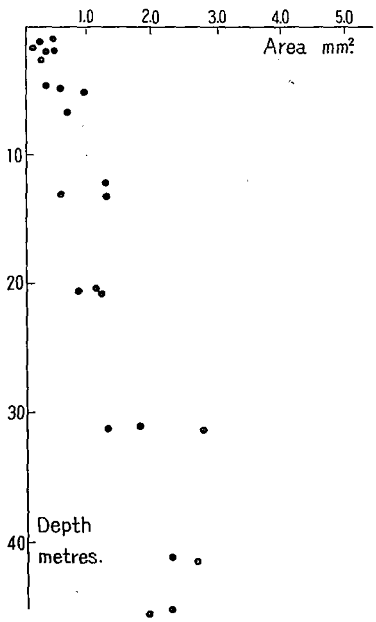

Fig. 6. Deeper crystal sizes at Southice

Depth profiles for crystal size below 1 m. This is shown in Fig. 6, in which nearly all values were obtained from horizontal sections. These features are noteworthy:

1. A general increase in size occurs with depth. This is approximately linear, and appears to coincide with the values obtained at Maudheim by Schytt (1958 b).

2. The scatter of values appears to increase somewhat with depth (being approximately 0.5 mm² near 1 m, 0.8 near 12 m, and 1.5 near 31 m).

3. The average rate of increase of crystal size to 45 m (0.05 mm² per metre) is less than a sixth that in the first 70 cm below the surface (0.34 per metre).

Coordination of the crystal size profiles. There is a gap in observations between 70 cm and 1 m. In addition, the measurements of material below 1 m are not numerous enough to establish details of the rate of increase in size. However, although this remains unconfirmed, it is supposed that the relationship is a transitional one: Following an extremely rapid increase in the first centimetre there is a much slower rate which declines gradually until reaching a more persistent low value.

Petrofabric analysis. A variety of lithologies were examined at various depths, and some typical contoured diagrams for directions of ice crystals *c*-axes have been published elsewhere (Stephenson and Lister, 1959). With the exception of wind-crusts and depth-hoar layers, almost all specimens gave very weak or virtually random orientations for *c*-axes. Both wind-crusts and depth-hoar layers have preferred orientation in a vertical direction, giving a simple polar maximum. The strength of this reached 20% in some wind-crusts and 9% in some depth-hoar layers. These strong petterns were already developed in wind-crusts only recently formed at the surface itself, and in depth hoar layers formed during the 1957 winter within 10 cm of the surface. Corresponding layers at depth, even at 45 m, still showed such orientations persisting.

In analyses of other lithologies, very weak patterns were sometimes present. Never more than 4%, these could usually be explained as of depositional origin. Plate-shaped drift grains are often parallel to the basal plane and on deposition tend to give a weak vertical pattern. Columnar drift grains are often elongated parallel to the *c*-axis, and on deposition tend to give weak patterns perpendicular to and parallel to wind direction. (Such patterns were observed for the two main prevailing winds.)

IV. Processes of Metamorphism

The essential process of snow diagenesis is one of sintering. A great part of accumulation at Southice takes place as drift snow, which is composed of grains that were originally snow crystals but have more rounded corners due to evaporation and abrasion. When it accumulates and is left undisturbed for a sufficient time, drift becomes fixed by sintering which forms neck connections between adjacent grains, and proceeds to simplify the outlines of such "dumbell" and more complex grain combinations. The mechanisms and rates of isothermal sintering in ice have been investigated in elegant experiments by Kingery (1960), Kuroiwa (1961) and Hobbs and Mason (1964). At lower temperatures the first two authors favour surface diffusion as the main mechanism, while Hobbs and Mason believe that evaporation condensation is the major process.

Yosida (1963) has shown theoretically that isothermal sintering (even with stress localisation) cannot account for the rapid growth in grain size which occurs in any shallow snow cover. He has indicated the major importance of temperature gradients, which promote vapour transfer. In Antarctica, snow temperature gradients are always present because of the big difference in winter and summer temperatures. Yosida and Kojima (1956) demonstrated by experiment the effect on snow grains of evaporation, vapour transfer and condensation under a temperature gradient. Evaporation and transfer take place from surfaces facing in the direction of decreasing temperature, and condensation occurs on surfaces facing the opposite direction; the result of this "hand to hand" transfer is to cause a "migration" effect up-gradient in the grains themselves. The process also changes various characteristics including density, permeability, grain type and grain size. de Quervain (1958, 1963) has demonstrated in a series of careful experiments the nature of these changes, and has also discussed the theory of metamorphism in several models of snow structure.

Summer and winter "Layers". Lister (1961) has emphasised that near the surface, grain size is inversely proportional to density. He (Lister, 1956) and other workers have shown the contrast between snow metamorphosed beneath the surface in summer (larger grains and lower density), and that metamorphosed beneath the surface in winter (smaller grains, more tightly packed). Apart from the possible effects of stronger wind packing and smaller primary grain sizes in winter, the contrasts can be attributed to a negative temperature gradient in summer, (with the possibilities of a superficial reversal caused by strong solar radiation), and a positive one in winter. Evaporation in the surface layers is active in summer introducing lower densities and, together with the influence of higher temperatures, larger grain sizes. In winter, evaporation occurs at depth, and condensation in the surface layers, producing higher densities, and increased grain sizes which are relatively smaller than the summer ones. Schytt (1958 a) has emphasised that winter deposits which happen to remain near the surface until summer acquire all the characteristics of summer layers. In such circumstances stratigraphic identification of summer and winter layers may correlate poorly with evidence from oxygen isotope ratios. If the annual accumulation is small, then the distinction between the two types of layers becomes blurred, the overall structure being a result of the stronger summer influence.

At Southice it was noticed that the 1957 winter accumulations underwent a metamorphism during the winter caused by the generally positive temperature gradient, with such effects as an appreciable increase in grain size. However, a vertical structure like that found in older layers did not develop. For this reason, it can be supposed that the higher temperatures and possibly the negative gradient of summer are necessary to produce it.

Wind-crusts and depth-hoar layers. Both these were observed to form during the winter at Southice, under presumed strong positive temperature gradients. The textures of these lithologies are interesting, especially since they develop preferred vertical orientations.

Wind-crusts form at the surface and are ice crusts less than 1 mm thick which occur singly, or in multiples closely adjacent. The texture of a freshly formed wind-crust is of polygonal crystals in horizontal section, the original single crystal grains having grown laterally towards one another near their equators forming thin bridging aprons. As these aprons develop, the crystal boundaries in vertical section are seen to be nearly vertical. Clearly, the metamorphism is one of vapour condensation in a critical horizontal plane at or very close beneath the surface. The extreme conditions also produce reorientation of the crystal structure. This reorientation does not involve any grain movement nor can it be discounted as simply preferential growth of only those crystals suitably orientated originally. Schytt (1958 a) showed that wind-crusts at Maudheim had formed during long precipitation-free periods under cold surface conditions. It is likely that additional requirements for their formation are a negative temperature gradient with a most abrupt fall in temperature at the surface itself. As their name implies, their formation is especially favoured by prolonged winds. While strong winds mix the air above the surface, smoothing out any temperature gradient, nevertheless at the surface itself, a very appreciable gradient can be expected across the top few millimetres.

The origin of multiple wind-crusts can be offered in terms of slight accumulation on the surface between crust formations. However, it seems more likely that the consistency of spacing often seen in multiple wind-crusts indicates modification of the temperature gradient details by altered external conditions, or by interference by the initial crust itself — a steadily forming wind-crust would have increasing thermal conductivity, and would become progressively more effective as a seal to air and vapour movement.

Depth-hoar layers were seen to form superficially at Southice during late winter at depths of less than 10 cm. The formation of depth-hoar layers has been explained by earlier workers as a sublimation zone under a negative temperature gradient, and de Quervain (1958) has demonstrated formation of hoar crystals experimentally. Giddings and LaChapelle (1962) have succeeded in predicting actual rates of formation theoretically. The depth localisation of the depth-hoar layer must be influenced by the details of the temperature gradient, as a local steepening at this level. Giddings and LaChapelle (1962) emphasise the contrast in conductivity between air and ice, which accounts for appreciable differences in conductivity in snow layers of dissimilar porosity. It is for this reason that depth-hoar so commonly develops at "summer surfaces" which are overlain by denser winter deposits. Depth-hoar layers are very commonly discontinuous laterally, in many cases being patchy or lenticular. This character may be attributed to irregularities on the snow surface at the time of formation or to original lateral inhomogeneities in strata and their boundaries. The texture of superficial depth-hoar layers studied at Southice was a granular one, containing a number of individual hoar crystals which consisted of simple disc-shaped crystals, orientated horizontally. In many cases, these disc-shaped crystals can be seen to have developed from grains similar to others nearby with random orientation. In some way the depth-hoar metamorphism causes a crystallographic reorientation without grain rotation and apparently without selective growth of only those crystals having suitable orientation.

Such mechanisms for the formation of surface crusts and depth-hoar have been proposed by previous workers including Bader (1939).

Metamorphism in deeper firn. At depths below the superficial layers most drastically affected by summer and winter metamorphisms, firn layers are subjected to similar but in general less extreme temperature gradients. The sense of these commonly reverses with depth due to the delays in warm summer wave and cold winter wave penetration. Vapour transfer metamorphism must continue under the influence of these gradients, with decreasing effect at greater depths. The effect of alternations of lithologies with different porosity and structure (having different conductivities) would cause variations in temperature gradient. The locally steeper gradients should occur in the less dense layers and the resulting metamorphism enhance contrasts between adjacent layers, perhaps balancing to some extent the faster compaction of the less dense layers. It is suggested that this continuing process (which is independent of the gradient sense) is responsible for the widening divergence of grain size in fine granular and coarser granular layers described in a previous section. It is also responsible for the continued development of depth-hoar layers down to the depth of their collapse. Taylor (1965) has discussed certain aspects of temperature gradient influence in snow metamorphism, and he has also emphasised how it determines certain other structural details.

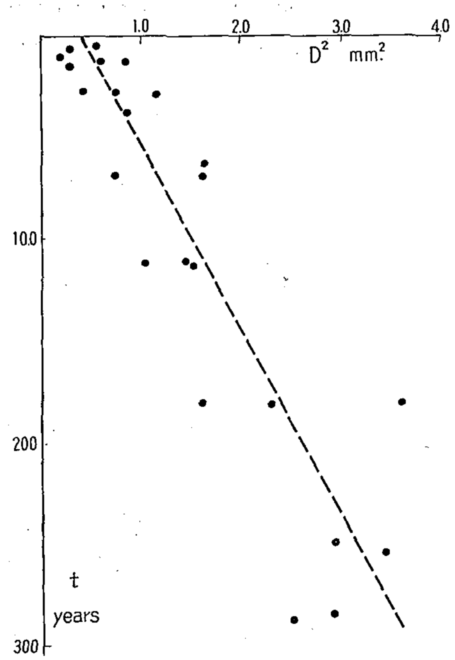

Fig. 7. Crystal size—age relationship

Crystal growth. The growth of ice crystals in firn is analogous with grain growth in metals and ceramics. The mechanism of this process of crystal boundary migration has been lucidly discussed by Smith (1949) in terms of imbalanced boundary energies.

Although the variation of crystal size is shown earlier in this paper as increase in area with depth, it can also be represented as D^2 varying with age, the same parameters used in metal and ceramic grain growth studies (where D is the diameter of the circle with area equal to the crystal cross-section). A graph prepared is shown in Fig. 7, age at various depths being integrated from the density profile and an accumulation rate of $10\,g{\cdot}cm^{-2}$. The results are again essentially linear.

Crystal sizes in the shallow profile show a faster average rate of increase than those of the deep profile, but as argued in an earlier section, they are assumed to be transitional. The reason for the declining growth rate with depth may be that superficial material experiences higher temperatures in summer than does deeper material. Grain growth in metals is considerably faster at higher temperature; this is also the case for ice, as is indicated later in this paper.

The extremely rapid crystal growth found in the first centimetre of the shallow profile may be influenced by such factors as the following:

1. The surface material is considerably younger than the material below 1 cm.

2. The primary crystal size in drift snow is relatively so small that as a crystal aggregate it is less "stable" than an aggregate of larger sizes. Examples of accelerated growth rates in very fine grained metals and ceramics would be parallel.

3. The initial crystal growth may not conform to a linear relationship between D^2 and time. For example, if D is plotted against depth (to approximate time details) a linear relationship is suggested for the interval between the surface and 1 cm (Fig. 8). This approximates to:

$$D = 0.1 + 0.03\,x\,,$$

(D, diameter in mm and x depth in mm)

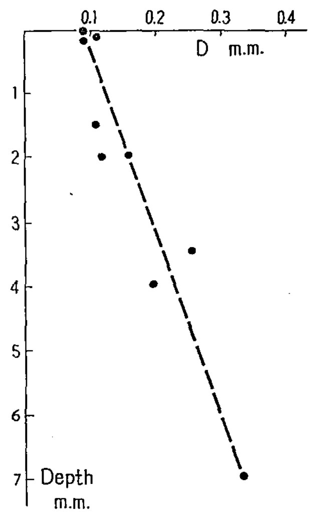

Fig. 8. Crystal size—depth relationship in the first centimetre beneath the surface

Possible mechanisms for the apparently regular variation of crystal sizes demonstrated in groups of layers near the surface (Fig. 5) are not known. They appear to contradict zones of maximum temperature, and regions of most intense evaporation and vapour transfer. The very coarse crystal size in depth-hoar layers is to be expected with the extreme evaporation metamorphism. Crystal size remains small in shallow wind-crusts because of restriction by the limiting surfaces of the layers, analogous to grain growth restriction in thin metal plate or wire. At depth, wind-crusts show relatively larger crystal sizes than most other lithologies; this is not understood.

V. Comparison of Crystal Growth Rates with Other Examples

For comparison, other measurements of crystal sizes in Antarctica (Bentley and others, 1964) and Greenland (Fuchs, 1959; Langway, 1963) have been considered in terms of D^2 and age (integrated graphically from density profiles and mean accumulations). In most cases the distribution of values approximates to a linear one. Growth rates are slower for lower temperatures. Later studies in grain growth rates in metals have confirmed a relationship found by Cole and others (1954) in steel and a theoretical derivation for it is discussed by Feltham (1957). In its simplest form the relationship is temperature dependent:

$$D^2 - D_0^2 = At \exp(-B/T).$$

where D is diameter at time t,

D_0 initial diameter,

A and B are constants,

T is absolute temperature.

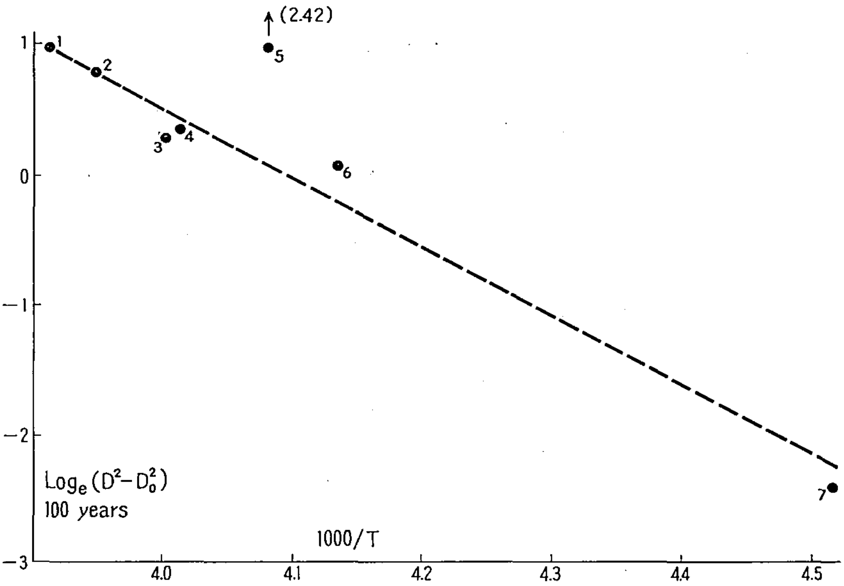

Fig. 9. The evaluation of the constant B in $D^2 - D_0^2 = At \exp(-B/T)$ (1—Maudheim; 2—Wilkes; 3—Site 2, Greenland; 4—Byrd; 5—Little America V; 6—Southice; 7—South Pole)

To test the temperature dependence of various growth rates found in polar firn and ice, a plot has been made of $\log_e(D^2-D_0^2)$ against $1/T$ for the time interval 100 years (Fig. 9). The result is apparently consistent giving a value for B of 5 450. (This constant is independent of the units chosen, but D was measured in mm and t in years.) While the result is an approximate one it appears to confirm that isothermal crystal growth rates in firn and ice are analogous to those in metals and ceramics.

Some results are anomalous. For example, growth rates at Little America are very much faster than would be anticipated. However, the rate of densification and other features at Little America such as petrofabric structure are also anomalous, and Gow (1963) attributes these to deformation. This may also account for the faster crystal growth.

VI. Conclusions

Contrasts in snow and shallow firn lithologies are metamorphic ones, governed by temperature gradients and their sense. Deeper metamorphic processes, also occasioned by vapour transfer, appear to be active to at least 8 m, and tend to maintain the shallow contrasts in spite of compaction.

Available measurements of crystal size in deeper firn and ice suggest that ice obeys laws similar to those in metals and ceramics. Many, more detailed studies need to be carried out to confirm or modify this, and a great number of details remain unaccounted for.

It appears that not enough long-term studies of progressive natural metamorphism in snow have been undertaken. Given adequate data, it should become possible to confirm process rates theoretically.

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In correspondence with A. J. Gow in early 1966, it became evident that he too was studying ice crystal growth rates and their possible temperature dependence. It is acknowledged that he reached independent conclusions similar to mine. He also gave me unpublished crystal size data for the South Pole.

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