MORPHOLOGY AND REGIME OF THE BRUNT ICE SHELF AND THE ADJACENT INLAND ICE, 1960–61

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ABSTRACT. The morphologies of the Brunt Ice Shelf and the southern part of Riiser-Larsen Isen, which together form an ice shelf approximately 75,000 km.^{2} in area, are described. The *in situ* development of large areas of low ice shelf where there is divergent flow is discussed, and the influence of the slope of the inland ice adjacent to its boundary with the ice shelf, resulting in fracturing of the ice sheet at the point of flotation, is assessed. At the present time the glacierization of the area appears to be in a state of equilibrium. Measurements of the net accumulation of snow at Halley Bay for the period 1945–61, the mean of which is 74·71 cm. yr.$^{-1}$ (38·99 cm. water equivalent), and for shorter periods of time over the remainder of the area, give a good indication of the true precipitation in the region as a whole.

THIS paper is based on glaciological investigations carried out on the Brunt Ice Shelf, the adjacent inland ice and at Tottanfjella during the period January 1960–February 1962.

Halley Bay is situated at the seaward margin of the Brunt Ice Shelf, on the south-east coast of the Weddell Sea. In 1961 the position of the British Antarctic Survey station, which lies just over 1·6 km. inland from Halley Bay, was lat. 75°30′40″S., long. 26°39′42″W. The ice shelf extends for 640 km. from "Dawson-Lambton Glacier" in the south possibly as far north-east as Kapp Norvegia. In one part it is known to stretch seawards from the inland ice for at least 160 km. East of long. 20°W. the ice shelf is known as Riiser-Larsen Isen (Fig. 1).

In 1904 the Scottish National Antarctic Expedition discovered an ice shelf, referred to as the "barrier", in lat. 72°18′S. (Shackleton, 1919, p. 23). This was followed to the south-west for 240 km. to lat. 74°01′S., long. 22°00′W. No rock was found exposed, but only rising slopes of snow and ice with shoaling water off the barrier wall, clearly indicating the presence of land. This area was named Coats Land by W. S. Bruce. Sir Ernest Shackleton reached

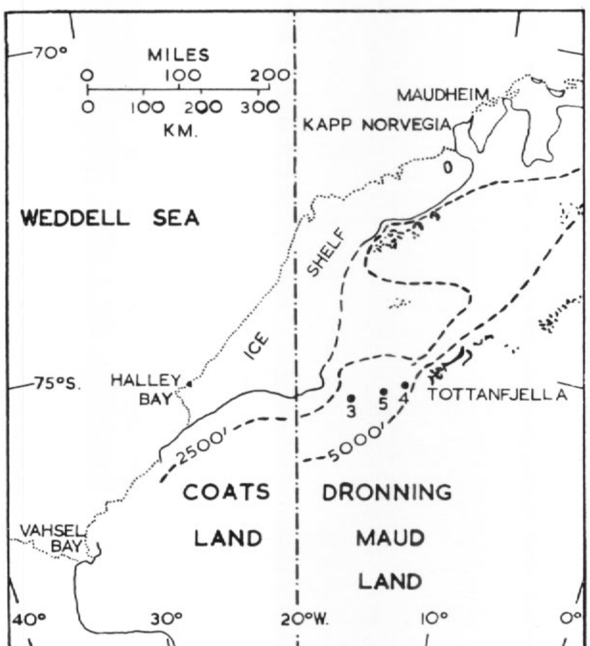

Fig. 1. Map of Coats Land and western Dronning Maud Land showing the positions of pits 3, 4 and 5.

this coast in January 1915 and followed it southward, encountering the "Stancomb-Wills Promontory" at approximately lat. 74°S., long. 25°W. This promontory extended for more than 80 km. into the Weddell Sea and was more than 80 km. across; soundings taken around this vast ice promontory indicated depths of about 360 m. on its landward side and 2,375 m. seaward. Shackleton (1919, p. 26) suggested that the whole mass was afloat and possibly destined to float away at some time in the future. This has undoubtedly taken place and a concentration of grounded icebergs now occupies this area. Proceeding southward along the ice front, Shackleton discovered two great glaciers: "Allan McDonald Glacier" and "Dawson-Lambton Glacier". The former is an ice rise, caused by local grounding of the ice shelf and has now been renamed the McDonald Ice Rumples, while the latter occurs where the inland ice reaches the sea south of the Brunt Ice Shelf. Shackleton described "Dawson-Lambton Glacier" as a huge overflow from the ice sheet, apparently flowing over low hills, the ice being heavily broken. The presence of a 2 m. tide mark on the cliff front and also an apparent elevation of more than 600 m. less than 3 km. from the coast gave ample additional evidence that the ice was aground. Shackleton named the coastline he had traversed the Caird Coast.

Aerial photographs taken by the Norwegian-British-Swedish Antarctic Expedition, 1949–52, revealed that Riiser-Larsen Isen continues unbroken, with the exception of locally grounded areas, as far west as long. 23°W. Also, preliminary maps of the nunatak areas in western Dronning Maud Land have been constructed from observations made on reconnaissance flights south from Maudheim, the expedition base. The most southerly group is a north-east trending range, Heimfrontfjella, the south-western part of which was flown over in 1957 by Sir Vivian Fuchs, the leader of the Commonwealth Trans-Antarctic Expedition, and named Tottanfjella. Because of a discrepancy between the position given to these mountains by these two expeditions, they were originally thought to be separate ranges.

The predominant feature of the landscape of Coats Land and western Dronning Maud Land is the extent of its ice cover and the consequent rarity of rock exposures, which are limited to groups of nunataks in Dronning Maud Land. Topographically, this area can be subdivided into three distinct regions: the ice shelf, the inland ice and the nunatak groups.

The Ice Shelf

The ice shelves of Antarctica are remarkably similar to one another despite the variations in environment and climate which occur over such a wide area. Wright and Priestley (1922, p. 205) have described the environment of an ice shelf as a "zone of balanced forces", where supply, movement and wastage exist together and no single factor predominates over the others. Supply is in the form of direct precipitation, the seaward movement of the land ice and, in some cases, freezing at the lower surface. Movement results from spreading, which follows accumulation, and also from the seaward flow of the land ice. Wastage takes place by calving of icebergs from the ice front, by melting from the lower surface and by ablation.

According to Robin (1958, p. 118), an attempt to attain an equilibrium thickness, dependent largely on the temperature of the locality, is constantly in progress. An ice shelf which is less than the equilibrium thickness will not respond to accumulation by spreading but it will increase in thickness and therefore in elevation above sea-level. Any thickness in excess of equilibrium will result in increased spreading exceeding that caused by accumulation. When an ice shelf is in equilibrium spreading takes place in response to accumulation and thus a constant elevation is maintained.

The extent of the Brunt Ice Shelf is now believed to be considerably greater than was previously imagined, because it was originally thought to be triangular in shape with Halley Bay at one apex (MacDowall, 1957). "Ice shelf A" (Riiser-Larsen Isen), mentioned by Swithinbank (1957*a*, p. 26), is almost certainly the north-eastern part of the Brunt Ice Shelf. An examination of air photographs taken near the ice front between long. 16° and 21°W. shows that it continues as far as long. 23°W. with only small locally grounded areas.

The presence of the McDonald Ice Rumples is a major factor in maintaining this ice shelf in its present position, where it smoothes out the coastline from "Dawson-Lambton Glacier" to Kapp Norvegia. If it were not for this ice rise, the seaward extent of the ice shelf would be greatly reduced and probably limited to the two bays (lat. 72–73°S. and 74–76°S.) in the coastline proper (Fig. 1). The reason for this is that the survival of ice shelves appears to depend on the anchorage provided by locally grounded areas; nowhere do they stretch far seaward unless they are protected (Swithinbank, 1955, p. 65).

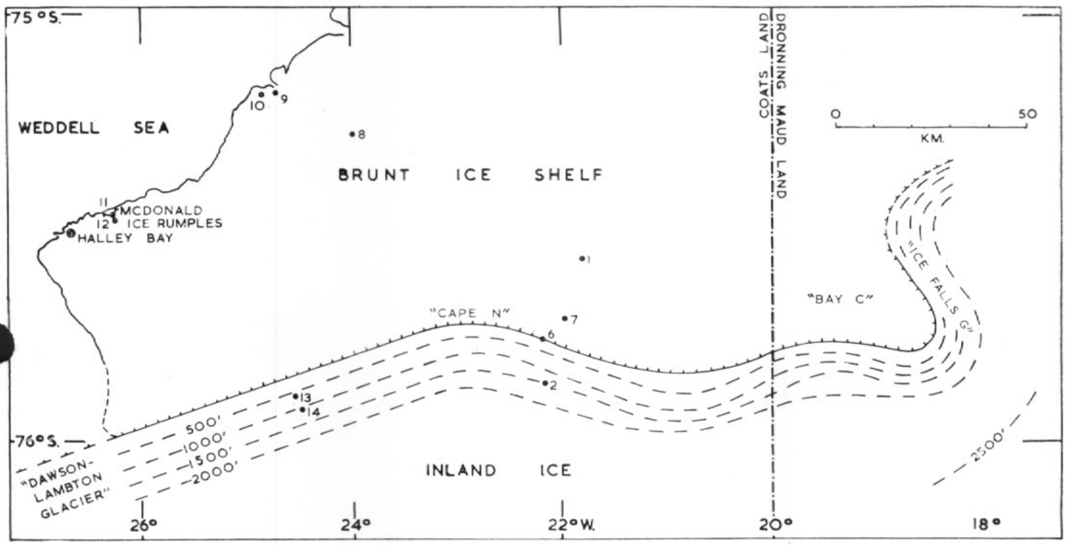

Fig. 2. Map of the Brunt Ice Shelf showing the positions of snow stratigraphy pits.

Topography

South of a line west-north-west from "Cape N" (Fig. 2) the characteristic feature of the ice shelf is the presence of gentle undulations which increase in amplitude and decrease in wave-length as the inland ice is approached. There are occasional disturbances associated with the latter up to 30 km. out in the ice shelf. Below ice falls and ice streams (Fig. 3), which occur regularly along the margin of the inland ice, these disturbances are more pronounced and extend farther into the ice shelf than in the intervening areas. The zone of disturbance is approximately 15 km. wide but between the areas immediately below ice falls it is narrower. Between "Dawson-Lambton Glacier" and "Cape N", an almost continuous chasm or rift separates the ice shelf from the inland ice and only occasionally is it filled in at the foot of some of the ice falls.

Large blocks or "icebergs", elongated parallel to the chasm and calving from the inland ice, form the basis of the disturbances; the areas between the blocks are covered with drift snow lying on a base formed by sea ice (Fig. 4). The topography is largely controlled by the effects of this drift snow deposition, the characteristic form being an asymmetric depression having a depth of approximately 25 m., a length up to 300 m. and a width of about 90 m. On the down-wind side of the long axis of the depression (which is parallel to the ice shelf boundary with the inland ice) a cliff about 20 m. high is usually present.

The katabatic wind which blows down from the inland ice becomes increasingly modified by the normal east-north-easterly prevailing wind over the rest of the ice shelf with increasing distance from the inland ice. This modification at first accounts for the asymmetry of the depressions, and eventually, when the katabatic influence is lost, it accounts for the final infilling and disappearance of these features.

No strand cracks occur on this section of the coastline, because the tidal movement of the ice shelf is taken up mainly in the chasm. Tidal movement of the sea-water has been observed in cracks in the sea-ice floor of the chasm, while sea-water in cracks in the bottom of depressions only a few kilometres from the junction shows no tidal movement.

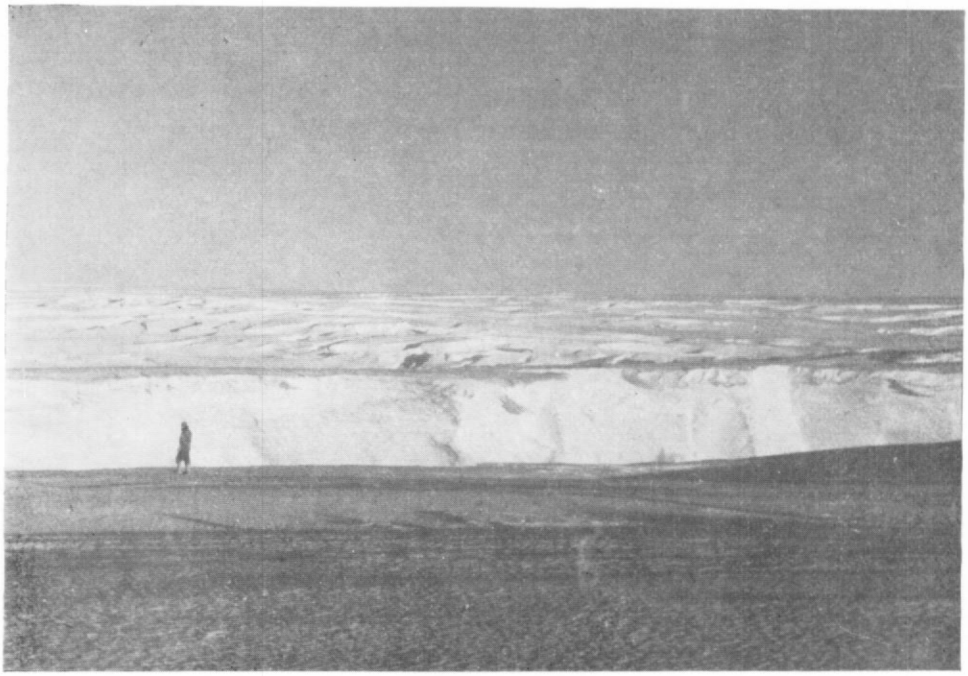

Fig. 3. An ice stream on the inland ice south-east of Halley Bay, viewed from the ice shelf. The chasm described in the text can be seen immediately beyond the figure.

Fig. 4. An unusually large area of snow-drifted sea ice with "icebergs" near the inland ice boundary south-east of Halley Bay.

The McDonald Ice Rumples are near the ice front in this area and undulations radiate from the ice rise into the ice shelf, decreasing in amplitude and increasing in wave-length as the distance from the ice rise increases; this disturbs the regular gently undulating pattern present over the remainder of the area.

The area known as "Bay C" (Fig. 2) is bounded to the east and south by the inland ice, and because of the flow of ice from these conflicting directions it is severely disturbed over its whole area. The slopes of the inland ice immediately adjacent to the junction with the ice shelf in this area are somewhat steeper and have a greater number of ice falls and ice streams than those between "Cape N" and "Dawson-Lambton Glacier". This is very marked in the inland ice east of "Bay C" where the very large "Ice falls G" occur. These are the "huge tumbling glaciers" referred to by Fuchs and Hillary (1958, p. 67).

North of a line west-north-west from "Cape N" and west of "Bay C" the character of the topography is greatly influenced by the change in direction of the true coastline. This swings eastward at "Cape N" and causes the almost continuous chasm bounding the inland ice to the west to split into an *en échelon* series of rifts.

All the features of the ice shelf in this area are orientated north-north-east to south-south-west due to the differential movements of the ice which to the east of "Cape N" is moving northwards, while to the west it is moving north-westward. There are long lines of shear crevasses which compensate for this divergent movement and large areas of low ice shelf (Fig. 5) are actively forming and fulfilling the same purpose.

Fig. 5. An area of low ice shelf surrounded by normal ice shelf north of "Cape N", viewed from the west.

In some areas of low ice shelf cracks have been observed at sea-level, approximately parallel to the long axes of the low areas. These cracks open slowly and at the same time freeze over, constantly producing new sea ice. This happens throughout the year and the absence of any swell or waves serves to protect the newly-formed ice which thickens, and, with the deposition of snow on its upper surface, is transformed into ice shelf. The accumulation over these areas of low ice shelf is much greater than on the normal ice shelf. The areas of low ice shelf, although irregular in outline, generally have their long axes orientated north-north-east to south-south-west. Pressure ridges having this same orientation also occur in this area and these are limited to the areas of low ice shelf, unlike the lines of shear crevasses which occur on both the low and the normal ice shelf.

At one boundary between the low and normal ice shelf, a feature normally characterized by drift slopes and low cliffs, the formation of a chasm, similar in many respects to the one at the inland ice shelf boundary, was observed. This was caused by a section of the normal ice shelf approximately 2 km. long and 1 km. wide calving and "falling back" towards the low ice shelf. This block was tilting slightly, presumably having assumed an equilibrium position after detachment from the parent mass. Strand cracks were observed between the *en échelon* crevasses at "Cape N" but they were absent along the remainder of the inland ice margin.

The front of the Brunt Ice Shelf south of lat. 75°S. can be subdivided into three sections. South of Emperor Bay it is relatively smooth until it meets the mouth of the chasm separating the ice shelf from the inland ice. Occasional high headlands are caused by the buoyant effect of a ram below sea-level, formed by wave action on the ice cliff. Between Emperor Bay and the McDonald Ice Rumples the coastline is severely indented with numerous inlets separated by headlands, some of which are raised by ram development. The inlets are all characterized by down-warping and in many cases they are bounded by crevasses of rift valley type. Two large crevasses or tension re-entrants extend for over 1·6 km. inland south-west of the McDonald Ice Rumples. On the seaward side of the McDonald Ice Rumples an area of low ice shelf has built up from accumulation on sea ice and broken material derived from the ice rise. Here, the ice shelf has been halted or deflected to either side of the ice rise.

The ice front north of the McDonald Ice Rumples is again relatively smooth until three large bays or inlets are reached (Fig. 2). These are associated with the low ice shelf areas north of "Cape N". Active calving of a section of the ice shelf approximately 250 km.^{2} in area, along 29 km. of the ice front, will soon release this section of ice shelf as one large tabular iceberg. It is hinged at its southernmost limits by a developing crevasse system which will detach the block once it has extended another 1·6 km., at which point it will reach the ice front.

This potential iceberg is bounded inland by an area of low ice shelf in which there is a developing crack at sea-level similar to that described above. This crack is linked to the inlet bounding the block to the north. The area between this inlet and the one to the north will eventually suffer a similar fate, while the northernmost of the three inlets is of the same type and is probably linked to a low ice shelf area separated from the ice front by another potential iceberg.

Throughout much of the year the coast north and west of the McDonald Ice Rumples is protected by well-developed sea ice held in place by a complex of grounded icebergs extending 50 km. north of Halley Bay. This is so in the area of the coast bounded in the east by the Brunt Ice Shelf and in the south by "Dawson-Lambton Glacier". Here, small icebergs are frozen into bay ice of several years' growth, forming an ice shelf of the second type described by Wright and Priestley (1922, p. 167).

11 km. east-north-east of Halley Bay is the ice rise known as the McDonald Ice Rumples (Fig. 2). A series of strand cracks is evidence of the grounding of this area. These have the characteristic bridging at the top with raised lips projecting approximately 2 cm. above the general surface level. These upturned lips are confined to the sides of the bridges and are never seen in a central position. A pit dug over one strand crack to investigate both the crack and the snow stratification showed the crack beneath the snow bridge, which occupied the uppermost metre, was 7 cm. wide. Only a cursory examination of this pit was possible, because it was being filled rapidly by drifting snow.

Measurements of the operation of a strand crack were made and the greatest movement recorded was an opening of 2·9 cm. in 7 hr. The tidal origin of these features was indicated by an opening and closing in every 12 hr.; these movements were heard and often felt from within the tent pitched near them. Their intensity is most marked midway between the two tides, the maximum activity being on the falling tide midway between the extreme positions.

The strand cracks follow the trend of a broad shallow depression and they occur on the side of it nearest to the ice rise. They are approximately 2 km. from the disturbance and within 500 m. of the bottom of the depression. This trends parallel to the margins of the ice rise and is followed by a further series of troughs and anticlines of increasing amplitude and decreasing wave-length, which culminate in the highest point. This is about 15 m. above the general level of the ice shelf.

The major deformational features of the ice rise, which covers an area approximately 2 km. long and 1 km. wide, are the series of sub-parallel anticlines and the evenly-spaced crevasses which cut them at right angles (Fig. 6). These crevasses vary from a few centimetres to 100 m. in width and they are widest at the anticlinal crests. The anticlines are the result of

Fig. 6. The major anticline of the McDonald Ice Rumples, looking west-south-west from its highest point. Rifts or tension re-entrants can be seen in the distance.

horizontal compressional stresses and in some instances the strain-rate increases to a point where it cannot be absorbed by plastic deformation, and the anticlinal arch breaks up into irregular blocks. Horizontal tensional stresses are responsible for the transverse crevasses.

Shear crevasses are also present and they are caused by the divergent flow of the ice shelf around the grounded area; the strand cracks pass seaward into a belt of these crevasses. A large rift 450 m. long and 100 m. wide has developed in the deformed area, which trends roughly parallel to the tension crevasses. This is approximately 30 m. deep and is floored with sea ice in which tide cracks were observed.

Zumberge (1958, p. 794–99) has also recorded deformed ice features in the Ross Ice Shelf between Roosevelt Island and the Bay of Whales which are comparable with those of the McDonald Ice Rumples.

Discussion

A factor emerging from the study of the Brunt Ice Shelf and the contiguous Riiser-Larsen Isen is the significance of the slope of the inland ice adjacent to its boundary with the ice shelf. It appears that once this slope has attained a certain attitude then the flow of ice from the land cannot be accomplished without fracturing along the line of flotation. This failure to accommodate the necessary bending of the ice results in the formation of "icebergs" (p. 15).

South-east of the northern part of Riiser-Larsen Isen the boundary with the inland ice shows a distinct break or cliff with associated disturbances of a type similar to those described earlier, i.e. asymmetrical depressions. The slope of the inland ice adjacent to this area is steeper than that around any of the other ice shelves described by Swithinbank (1957*a*, p. 27), who has attributed these features to an increase in slope. However, it is less steep than that of the inland ice adjacent to the boundary between "Dawson-Lambton Glacier" and the "Ice falls G", east of "Bay C". Throughout this area ice falls and ice streams occur regularly, often extending above the 300 m. contour. The average gradient for the first 500 m. is approximately 1 in 12 and it is undoubtedly this slope which gives rise to fracturing along the line of flotation, thus causing calving of the large blocks or "icebergs".

The amount of slope necessary to initiate the break-up of the ice at the margin of the land depends upon the thickness of the ice at that point and also upon its speed of flow, both absolute and relative to adjacent areas. Swithinbank (1957*a*, p. 32) has described the disturbance caused by the relatively fast-moving "Ice stream A" at its entry into "Ice shelf D" in western Dronning Maud Land, and this example clearly illustrates the effects of differing rates of movement in adjacent ice masses when crossing the line of flotation at the inland ice margin. The result is a much greater degree of fracturing than would otherwise have been encountered with uniformly flowing ice and thus the formation of "icebergs" similar to those described on p. 15. Along the inland ice margin adjacent to the Brunt Ice Shelf and near "Bay C" the alternation of relatively fast-flowing ice streams and ice falls with smaller relatively undisturbed intervening areas enhances the effect of the slope itself and increases the extent of the break-up suffered by the ice on flotation. In addition, the material derived from ice streams and ice falls may often be excessively broken by crevassing before it reaches the ice shelf boundary.

The development of a chasm along much of this line of fracture and the separation of the individual blocks or "icebergs" is probably brought about by the mechanism described by Swithinbank (1957*a*, p. 27), which is similar in character to that suggested by Debenham (1948, p. 210). The continued freezing of water in the tide cracks in the initial fracture forces the ice apart, while the sides of the chasm undergo continuous relative movement in a vertical plane. As most of the tidal movement is accommodated in or near the main line of fracture, this process ceases when the "icebergs" have moved away from this zone. The blocks then cease to move relative to one another and the ice shelf proceeds to consolidate.

The chasm described on p. 17 is initially different in origin to the chasm associated with the inland ice margin. The former was caused by part of a large area of normal ice shelf attempting to achieve equilibrium and its detachment probably commenced initially by crevasse development during the break-up of the ice into "icebergs" on flotation at the inland ice boundary. In this case the equilibrium position was attained because of the absence of or the low resistance from the adjoining low ice shelf area.

The depression associated with the boundary between the ice shelf and the inland ice sheet, which is mentioned by Thiel and Ostenso (1961, p. 825) and Swithinbank (1957*a*, p. 17), is usually not apparent at the inland margin of the Brunt Ice Shelf. The extensive break-up of the ice following flotation successfully masks any semblance of this feature over much of this area. However, at "Cape N" the strand cracks between the *en échelon* chasms are situated on the landward side of a depression. Whether this depression has resulted from constant bending in response to tides, with consequent thinning, as suggested by Robin (1958, p. 124), or whether it has resulted from the rigidity of the ice causing the absence of local hydrostatic equilibrium is uncertain. The limited occurrence of such a regular depression again illustrates the influence of the degree of slope of the ice sheet adjacent to the Brunt Ice Shelf. Throughout almost the whole area thinning by tidal action (similar to that which takes place when a metal strip is bent in order to break it) is unnecessary, because the tensile strength of the ice is insufficient to accommodate the necessary change in attitude from its grounded position to its position afloat, and consequently fracturing occurs on flotation.

Another factor influencing the development of crevasses in floating ice shelf is described by Thiel and Ostenso (1961, p. 830). This is a change in the direction of movement of the ice on reaching the ice shelf, from perpendicular to the margin of the grounded ice to, in the case they have described, parallel to it. The change in direction of the flow of the ice after crossing the line of flotation and the subsequent stresses involved are also exhibited to a lesser degree between "Cape N" and the "Ice falls G". Examples of crevasse development due to divergent flow are shown on some of the Commonwealth Trans-Antarctic Expedition air photographs.

One of the Commonwealth Trans-Antarctic Expedition air photographs shows an ice stream, which commences at an altitude of approximately 500 m. above sea-level on the inland ice sheet and flows down into the ice shelf between "Bay C" and "Cape N". Little break-up on flotation can be seen but a series of sub-parallel depressions, indicating the ice is afloat, is well developed in the ice shelf. Debenham (1948, p. 211) has suggested that the enlarging action of sea-water freezing in vertical cracks is the cause of these features but he has not explained how the cracks can penetrate to the bottom of the thick free-floating ice shelf when the maximum depth of crevasses is usually estimated to be of the order of 30 m. The name "hinge line valleys" has been given to these features by various American scientists, in particular Poulter (1947) who attributed the hinging to tidal bending where the flat ice shelf slides off moraine and floats freely. Robin (1958, p. 119), dissatisfied with these explanations, believed that the depressions originate when divergent flow is too great to be accommodated by spreading due to accumulation. Why valleys develop in preference to a general uniform thinning is not yet known but, according to Robin, "the presence of a valley must have the effect of relieving the tensional stress to a limited extent to one or both sides of the valley, as there frequently appears to be flat stretches of ice shelf between such valleys", and "the rate of sinking of a valley near Maudheim is much greater than the calculated thinning for that zone, so presumably the required thinning of the adjacent ice shelf is concentrated into formation of the valley".

The fact that these depressions are clearly an active feature supports Robin's contention, otherwise they would have been filled by the accumulation of snow. It would appear, however, that they originate at the margin of the inland ice where thinning may be attributed to tidal bending and divergence due to tension within the ice shelf. This dual role of the marginal depression ceases once another depression is formed and tension becomes the sole operating mechanism. To the north-west of this area tension within the ice shelf increases and break-up along the line of the depressions takes place. It seems that the name "divergence or tension valleys", suggested by Robin (1958, p. 122), is preferable to "hinge line valleys". The latter perhaps should be restricted to depressions associated with grounded ice margins.

Siple observed that the most landward "hinge line valley" may be a straight-walled canyon with a sea-ice floor ([Roscoe], 1953, p. 58). This corresponds to the chasm described on p. 15.

Wright and Priestley (1922, p. 164–69) have described the development of ice shelf by two processes. The first type forms originally from land ice extensions with or without interstitial sea ice, while the second forms mainly or entirely by the accumulation of snow upon sea ice which has persisted for several seasons. Both of these types of ice shelf development are well illustrated by the Brunt Ice Shelf; the former process is the more prominent and also exhibits the inclusion of interstitial sea ice where the sea-ice base between the "icebergs" formed at the inland margin is covered with drift snow and precipitation. The second process operates on the seaward side of the McDonald Ice Rumples and also in the sheltered bay formed by the coastline of "Dawson-Lambton Glacier" and the south-west front of the ice shelf. At the McDonald Ice Rumples conditions are directly comparable with those in the Bay of Whales in the Ross Ice Shelf described by Gould (1935, p. 1379, 1389–91). A third process has also been observed, in which divergent flow of large areas of normal ice shelf results in the formation of ice shelf *in situ* (p. 17).

The first stage in the development of an ice shelf of this type is the break-up of an area of normal ice shelf under tension. Continued separation of the resultant blocks gives rise to an area of open water existing in sheltered conditions and this freezes over rapidly. The situation of this new ice is conducive to heavy accumulation relative to the normal, both from direct precipitation and from drifting snow, and it thickens as a result of this accumulation and also by continued freezing on its lower surface. However, while this is taking place increased separation of the marginal blocks continues and more ice is formed over the exposed areas. The ultimate end product of this process should be an area of ice shelf adjoining the original divergent blocks, which at its outer margins is equal in thickness to the adjacent normal ice shelf and which descends gradually to the sea ice and the crack from which it originated. This assumes that the normal ice shelf does not exceed its equilibrium thickness, a condition which is unlikely as equilibrium was probably attained by spreading before break-up took place. The attainment of an equilibrium state in the new low ice shelf has not been encountered, because its increase to a sufficient thickness is prevented by calving of the area on reaching the ice front.

In the region north-west of "Cape N" this low ice shelf forms almost 50 per cent of the total area, the reason for this being the conflicting directions of supply of land ice mentioned previously, and the consequent inability of the ice shelf to compensate for the inadequate supply of material solely by spreading.

Swithinbank (1957*a*, p. 33) has observed the presence in air photographs of rifts due to differential movement in "Ice shelf D" (western Dronning Maud Land) but he has not suggested whether they are developing or closing. Blackburn (1937, p. 598), describing crevassed areas on the Ross Ice Shelf, mentioned an area of low elevation at lat. 83°18′S., long. 157°W., but he made no suggestions as to its origin.

Pressure ridges are often encountered in areas of low ice shelf. These often exceed 3 m. in height and can be up to 100 m. across. They are apparently caused by interruptions in the divergence of areas of normal ice shelf, either by an increase in speed of one block relative to another seaward, or by some other movement of the blocks contrary to their normal divergence.

The boundary between normal and low ice shelf is characteristically marked by extensive drift deposition. In many cases shear crevasses often mark this line, since they accommodate lateral movement between the divergent areas of the normal ice shelf bounding the low area. Shear crevasses also occur occasionally within the low ice shelf, and in cases where there is a line of shear crevasses in the normal ice shelf they may be continued in the low area.

The Inland Ice

The inland ice adjacent to the Brunt Ice Shelf and Riiser-Larsen Isen south of lat. 75°S. is characterized by many ice falls and ice streams. They usually occur below a height of 450 m., but "Ice falls G" east of "Bay C" are an exception and appear to extend up to approximately 600 m. Throughout this area the slope down to the ice shelf is persistently steep and to the south it passes into the ice stream of "Dawson-Lambton Glacier". Except at "Cape N" and below certain of the ice falls, the slope of the inland ice usually terminates in a cliff which forms the inland side of the chasm separating the ice shelf from the grounded ice. At "Cape N" strand cracks are present between the *en échelon* series of chasms, which define the margin of the inland ice.

Beyond the zone of ice falls and ice streams adjacent to the ice shelf, the inland ice rises in a series of gentle undulations. Occasionally there are small ice falls reflecting the underlying topography and these form the only breaks in the featureless monotony of the landscape.

Additional information provided by Blundell and Winterton (1962) has revealed more detail of the surface topography in the area between Riiser-Larsen Isen and the nunatak groups, Milorgknausane, "Nunataks L", Tottanfjella and Heimfrontfjella. Immediately south of Milorgknausane a deep valley, only 150 m. above sea-level less than 5 km. from the range, trends from north-east to south-west (Fig. 1). On the southern side of this valley, south-south-east from Milorgknausane, "Nunataks L" attain a maximum height of 1,230 m. above sea-level. They lie on the north side of a broad spur which extends westwards from Tottanfjella towards the ice shelf north of "Bay C".

At Tottanfjella (Figs. 7, 8 and 9) blue ice fields were observed on the lee side of the col separating "Peak H" from "Peak A" (Fig. 10) and also in the lee of "Peaks E". In each case the ice surface was smooth and the ice itself was blue, impermeable and contained many enclosed air bubbles. Vertical tension cracks at right angles to one another were present at each location. These were never more than a few millimetres wide and they appeared to be orientated parallel to and at right angles to the direction of ice movement. Scattered surface moraine over each of these ice fields indicates a negative regime, while many of the boulders are partly surrounded by scoops caused by wind action and radiation.

Schytt (1961, p. 188) has described the mode of occurrence of blue ice fields in western Dronning Maud Land as follows: "We know that in past times the ice sheet was a few hundred metres thicker than it is today. During the gradual thinning process the accumulation *in situ* was able to preserve the general crystalline structure of the ice; there is no reason to doubt that the thicker ice sheet of several hundred or even several thousand years ago was, as far as crystal structure is concerned, very similar to the ice of today. In certain locations, however, the accumulation was insufficient and lower levels of the ice sheet became exposed. When the retreat came to an end, and a new general equilibrium was established, the ice field could continue to exist because of the local climatic conditions on the leeward side of the high mountain ridge." This localized negative regime of the present time may be attributed

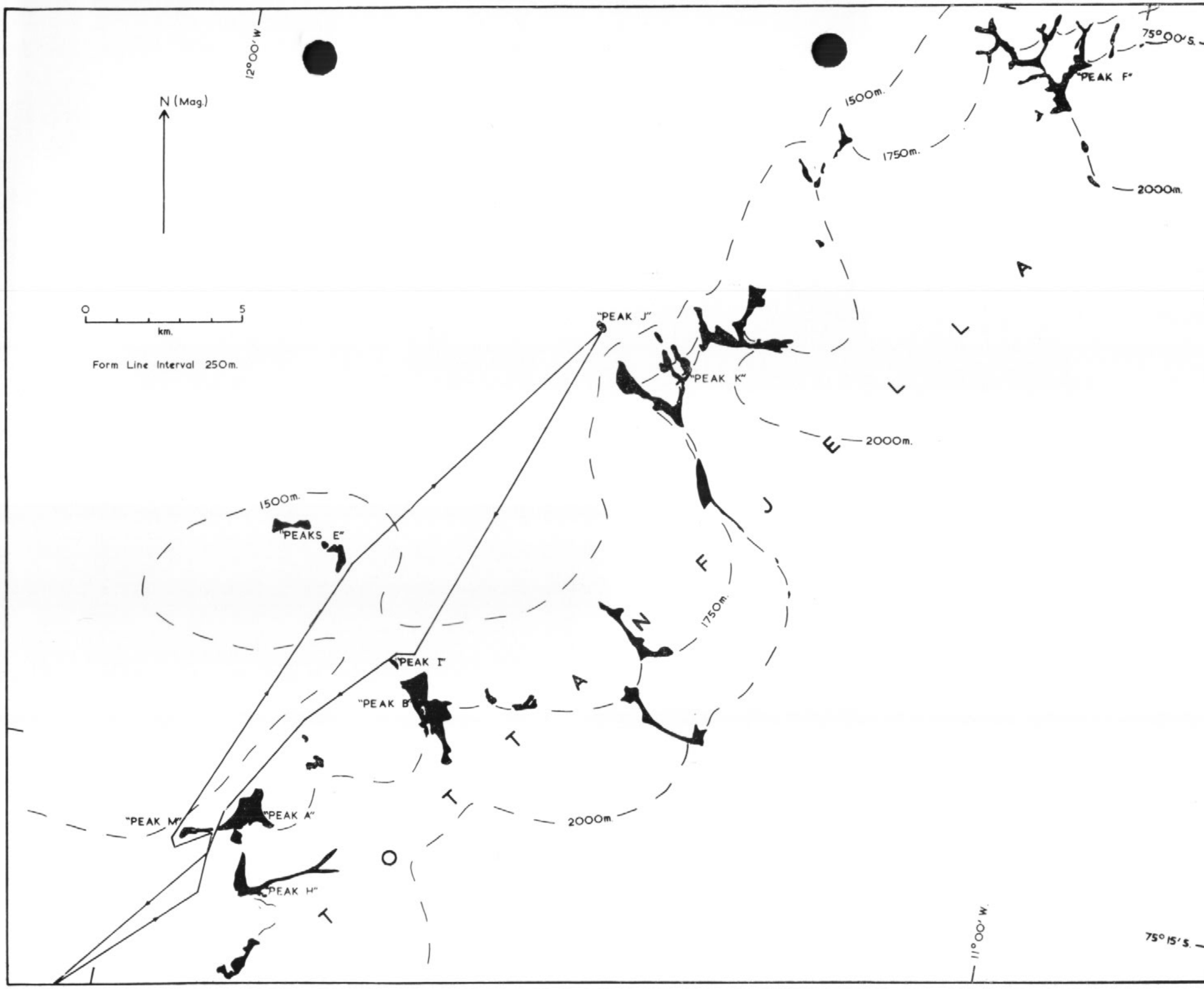

Fig. 7. Topographical sketch map of Tottanfjella, Dronning Maud Land. The sledge route is also shown.

Fig. 8. Looking south-westward over Tottanfjella with "Peak K" in the foreground. (Photograph by courtesy of the Commonwealth Trans-Antarctic Expedition, 1955–58.)

Fig. 9. View to the south-west from the immense snow drift in the lee of "Peak J".

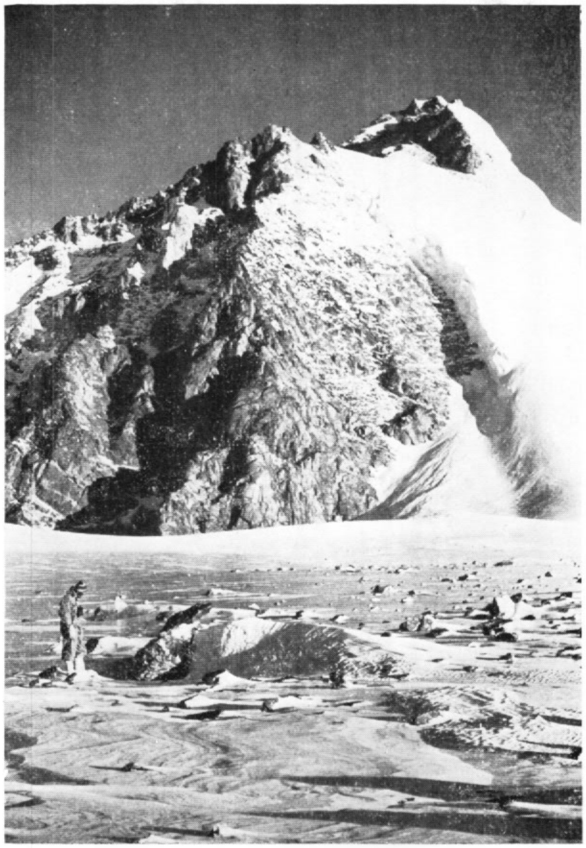

Fig. 10. "Peak H" (2,150 m. above sea-level), Tottanfjella, Dronning Maud Land, viewed from the west. A small blue ice field (about 0·5 km. across) can be seen beyond the figure in the foreground.

to ablation through powerful wind erosion, high evaporation caused by föhn winds and by heat radiation from adjacent rocks.

The evidence provided by the lichen distribution and the occurrence of glacial erratics shows that considerable ice retreat has taken place in the past, but that the thinning of the ice sheet has not been in progress for several decades or even centuries. Net accumulation is predominant over this area, except in places where local topography has caused exceptional wind conditions, but no major advance can be seen in the ice caps of the promontories which form part of the range (Ardus, 1964, p. 18), since they are very thin and inactive. It would thus seem that the inland ice sheet in this area is at present in a state of equilibrium.

Surface Features

The modification of the snow surface by wind produces characteristic snow drifts, ripple marks and skavler or sastrugi. The latter are most characteristic, being present throughout the year in most types of snow. They are up to 30 cm. high on the ice shelf but on the inland ice within 80 km. of Tottanfjella they are sometimes almost 1 m. high. For newly-formed skavler, procession at rates of up to 15 cm./hr. in wind speeds of 11 m./sec. have been observed.

The east-north-east orientation of the skavler throughout this area is only disturbed on the slopes of the inland ice adjacent to the ice shelf. Here, the katabatic winds are more persistent than the prevailing winds and the skavler are orientated perpendicular to the boundary of the inland ice with the ice shelf.

Barchan snow drifts occur rarely and the more usual form is the simple drift. Ploughshare ice (Wright and Priestley, 1922, p. 272) occurs on steep north-facing slopes among the disturbances adjacent to the inland ice and at the McDonald Ice Rumples.

Unlike the features described above, *penitentes* occurring in Emperor Bay are limited in location and time. They develop in summer often to a height of over 0·5 m. where the effects of insolation on the snow are magnified by the concentration of emperor penguin droppings.

Regime

In the determination of the mass balance of the Brunt Ice Shelf the net snow accumulation is by far the largest item on the credit side of the annual budget. The mass balance is the quantitative relationship between supply and removal over the whole ice mass, while the regime is concerned only with the condition of the upper surface. The factors controlling the regime are accumulation, either in the form of solid precipitation or drift snow, and ablation caused by deflation or evaporation; it is thus a measure of the amounts of water added to or subtracted from the upper surface. If a negative or balanced regime is established in an area, ice retreat results, but if there is a positive regime it is necessary to establish whether movement can remove an equivalent volume of ice before a condition of advance or retreat can be determined. An exception to this is the ice shelf nourished by subglacial accumulation, a condition exhibited by the McMurdo Ice Shelf (Debenham, 1920, p. 51; 1948, p. 205; Swithinbank and others, 1961, p. 764).

In contrast to glaciers in most parts of the world, Antarctica is almost entirely an area of accumulation, maintaining its equilibrium by the calving of icebergs rather than by ablation. Prior to the Norwegian-British-Swedish Antarctic Expedition, 1949–52, little attention had been paid to regime in Antarctica, and few detailed measurements had been made on ice in areas unaffected by topography. Accumulation measurements by means of stakes were made during the National Antarctic Expedition, 1901–04 (Dines, 1908, p. 471) and during the United States Antarctic Service Expedition, 1939–41 (Dorsey, 1945, p. 350; Wade, 1945, p. 167), but these were either limited in time or influenced by local wind effects. The glaciologists of the Norwegian-British-Swedish Antarctic Expedition, who hoped to use methods which had been developed in work on temperate glaciers (Ahlmann, 1946, p. 315), established the techniques which are now largely employed in measurements of regime in Antarctica.

In the present work regime determination was carried out by measurement of the amount of accumulation against stakes set vertically in the snow and also by the examination of the stratigraphy in a series of pits. The latter provided a measure of the water equivalent of the accumulation and details of the annual increments of accumulation dating back to 1944.

Accumulation

Two types of stakes were used for snow accumulation measurements: bamboo and 3·8 cm. square-section ramin wood. Jointed alloy stakes, used in the study of the relative movement of the ice shelf, were found to be unsuitable because when separating sections to erect a theodolite directly over a stake it was found that the stakes moved relative to the snow surface. The metal stakes also had a tendency to loosen in the snow during warm summer spells. Most of the accumulation stakes were erected vertically in the snow but in addition six sets of ramin wood "goal posts" were used. The level of the snow surface beneath the cross-bar of each of these was measured at ten points.

Throughout 1960 a pattern of nine stakes south-east of the station huts was used as a monthly control, while detailed variations in the surface level were measured against a pair of "goal posts". In 1961 a new line of ten control stakes, aligned north—south 1,000 yd. (914·4 m.) east of the station at intervals of 100 yd. (91·44 m.), and four additional sets of "goal posts" were used. These new stakes were all allowed to settle before measurements were taken and there was a four-month overlap of records before the 1960 patterns were abandoned. These stakes had to be abandoned because new living accommodation was erected at the station and the area influenced by the existing large drift therefore increased. A new cold laboratory was also established in May 1961 close to the 1960 control stake pattern. In order to provide the detail of changes in surface level for constructing the accumulation trace (Fig. 11), the readings taken at the "goal posts" were corrected graphically to fit the monthly measurements at the control stakes.

Because density variations in the surface layers are almost insignificant from one place to another owing to the nature of the topography, the mean accumulation was measured in cm. of snow. The water equivalent of this was calculated later from density measurements made in a pit.

TABLE I. NET ACCUMULATION STAKES ON BEARING 130° MAG. FROM HALLEY BAY

Distance from Halley Bay (km.)	*Net Accumulation of Snow* 16 March–6 November 1961 (cm.)
8·1	82·5
11·3	64·9
14·5	71·6
17·7	79·9
21·1	98·5
24·2	99·9
27·4	73·1
30·6	80·0
33·8	120·0
34·9	111·7
37·0	141·1
40·5	93·2
43·6	79·3
44·1	107·2
45·2	122·4
45·7	67·0

Swithinbank (1957*b*, p. 55) clearly established the influence of any surface slope, however slight, on the resulting accumulation. He also found that proximity to the ice front, on a level or almost level surface, led to a smaller accumulation. In view of this, the stakes used for accumulation measurements at Halley Bay were all erected in a level area more than 2·5 km. from the ice front and up-wind of the large drift caused by the station buildings. Other observations made on the Brunt Ice Shelf confirmed Swithinbank's conclusions. In undulations, which are quite common on the ice shelf and which increase in amplitude and decrease in wave-length towards the inland ice, the amounts of accumulation exceed that of the intervening areas and over level expanses. A line of 16 stakes, erected at intervals on a bearing of approximately 130° mag. from Halley Bay to within 16 km. of the inland ice, confirms this fact, since the stakes recording relatively high accumulation were situated in depressions (Table I). Similarly, excessive amounts of accumulation were recorded in the depressions leading to bays in the ice front and also in areas of low ice shelf described earlier. In contrast, close to the ice front, especially on high headleads, the permanence of the surface is very pronounced, being hard, wind-crusted and heavily metamorphosed. This ice front effect is also developed near the margins of blocks of normal ice shelf in areas where low ice shelf is developing.

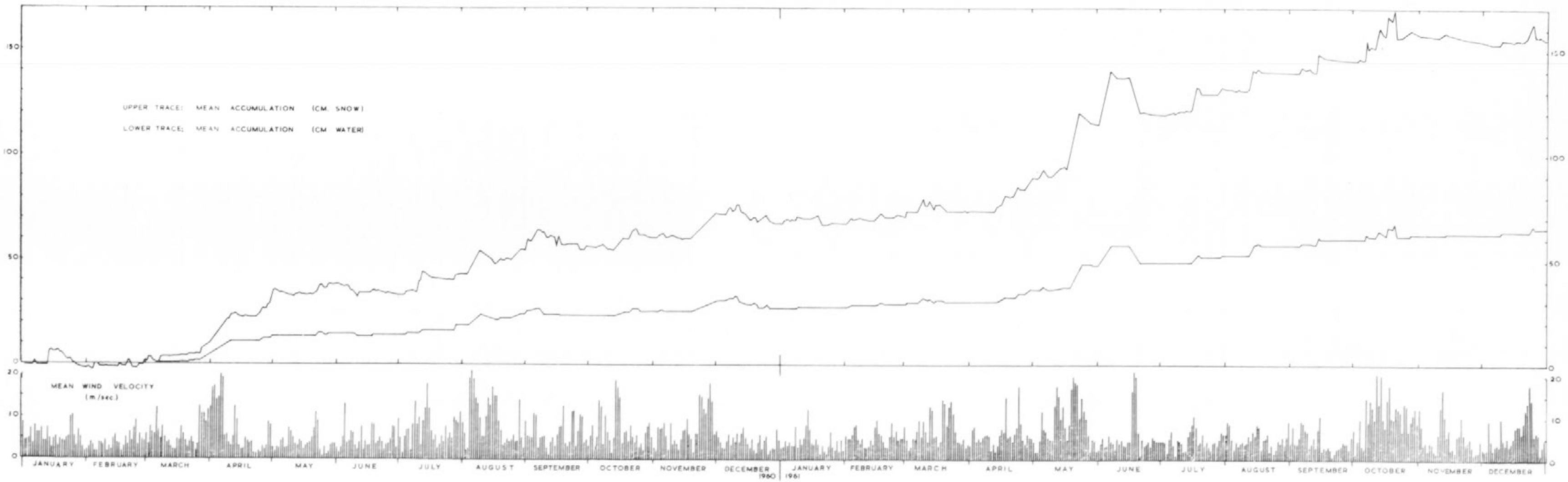

Fig. 11. The snow and water equivalent accumulation traces and the daily mean wind speeds at Halley Bay for the years 1960 and 1961.

In contrast to these variations caused by topographical changes, the stake measurements shown in Table II illustrate the good agreement with the results obtained at Halley Bay. These stakes were established up-wind from the latter in level areas, and therefore similarly subject to a net accumulation representative of the ice shelf as a whole.

Minor surface irregularities in the form of skavler and snow drifts can cause large discrepancies between stake measurements; this is very evident in measurements taken relative

TABLE II. NET ACCUMULATION STAKES UP-WIND FROM HALLEY BAY

Date	*Bearing from Halley Bay* (°mag.)	*Distance* (km.)	*Net Accumulation of Snow* (cm.)
8 November 1959–22 November 1960	068·5	8·8	60
22 November 1960–20 December 1961			71
9 November 1959–7 December 1960	067·0	15·3	69
7 December 1960–20 December 1961			74
10 November 1959–7 December 1960	069·5	20·1	69
7 December 1960–20 December 1961			83
11 November 1959–10 November 1960	084·0	17·2	56

TABLE III. MONTHLY NET ACCUMULATION AT HALLEY BAY

Month	*Net Accumulation* (cm.) *1960*	*1961*	*1962*
January	−1·45	−0·5	+0·7
February	+2·90	+3·2	
March	+9·25	+2·9	
April	+24·75	+15·8	
May	+2·85	+25·3	
June	−4·80	+4·7	
July	+10·10	+12·1	
August	+11·20	+7·7	
September	+1·80	+6·1	
October	+4·10	+12·6	
November	+11·60	−3·1	
December	−4·40	+0·7	
TOTAL	67·90	87·5	
COMBINED TOTAL	155·4		

to the cross-bars of the "goal post" stakes. This micro-morphology results in areas of erosion and areas of deposition being separated by only a few centimetres during a storm.

In the construction of the diagram illustrating the mean net accumulation (Fig. 11) it is necessary to consider any possible errors that may have arisen. One fact which should be borne in mind is that the diagram represents the changes in level between the times at which measurements were taken, and is therefore not a complete record of the continuous variations taking place at the surface. The height of the surface relative to the stakes was measured to the nearest millimetre and the mean of each pattern was calculated. The results obtained

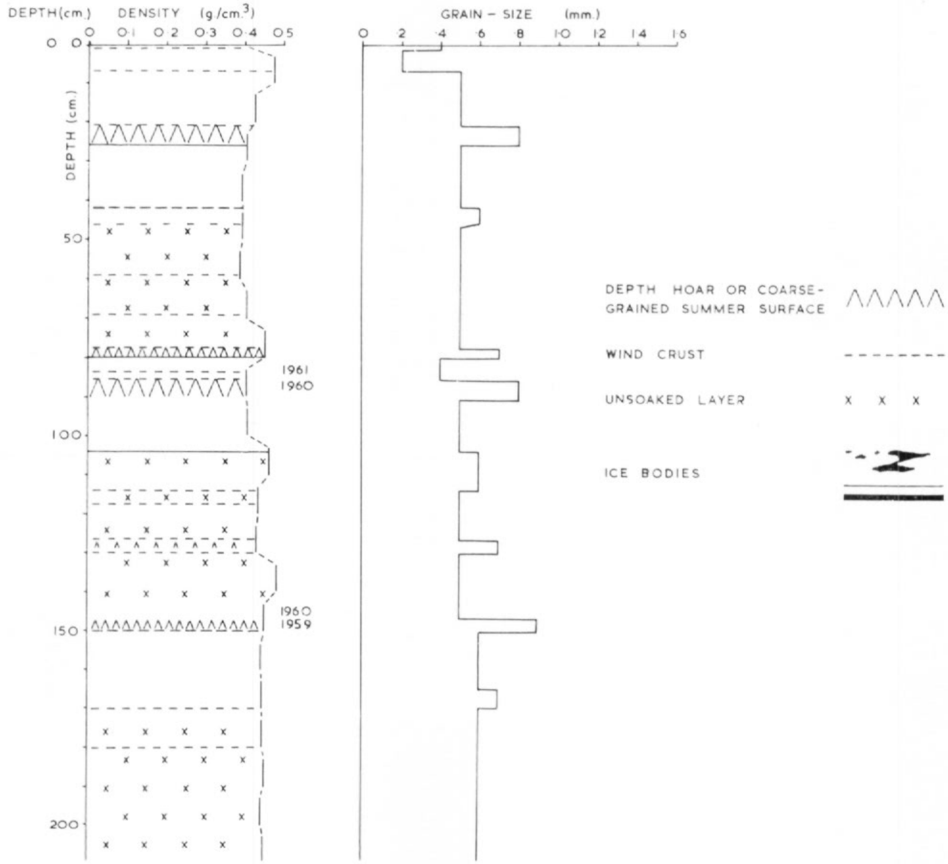

Fig. 12. The snow stratigraphy, density and grain-size profiles of the pit dug at Halley Bay on 31 January 1962.

from the "goal posts" were corrected graphically to fit the monthly figures obtained from the control stakes, and these monthly net accumulation figures are shown in Table III. The amount involved in the proportional correction of the figures obtained from "goal posts" to those from the control stakes was usually very small, and their mean values for the 1960 annual increment differed by only 3 mm.

Any error incurred by the sinking of the stakes relative to the snow surface must be negligible, as all stakes were allowed to settle before being used. A newly-erected bamboo stake, placed close to a stake of the same type which had been in position for two years, sank 4·8 cm. relative to the older stake in 96 days. After this period the stakes remained fixed relative to one another. Although no measurements were made during the 96 days, it can be safely assumed that settling of the new stake was greatest just after its erection.

As most stakes were embedded in the 1959 accumulation, the settling of the 1960 and 1961 accumulation took place without influencing the position of the stakes, and therefore no correction factor is necessary.

To calculate the water equivalent of the net accumulation for 1960 and 1961 a pit was dug to a depth of 2 m. on 31 January 1962. Density measurements were made and the resulting profile (Fig. 12) was assumed to be representative of conditions in the area of the stake measurements. The snow trace (Fig. 11) was then treated in the manner described by Swithinbank (1957*b*, p. 60), points being established where no accumulation had taken place for several days and the water equivalent calculated from the density profile of Fig. 12. From each of these points the water equivalent was traced to the previous point, usually in proportion to the amounts of accumulation. However, in many cases a lowering of the surface level has been interpreted as settling of the snow and not as net ablation. From measurements made in the deep pit (p. 32) a rate of settling of 0·145 cm./day for the uppermost 2 m. has been calculated, approximately 80 per cent of this taking place in the uppermost metre. This compares favourably with a calculated rate of settling of 0·141 cm./day obtained from a large number of density measurements taken at the surface, the mean of which was 0·32 g./cm.^{3}, and the known density at 2 m. obtained from the density profile of Fig. 12, 0·45 g./cm.^{3}. Undoubtedly, the greater amount of settling took place in newly-deposited snow at or near the surface, but occasionally a drop in surface level took place which was too great to be explained by settling alone. Deflation was undoubtedly the cause of this ablation, since evaporation is an insignificant factor in the regime.

When considering the regime in relation to time it is found to be markedly positive, despite the stepped profile of the accumulation trace. The majority of falls in the surface, which are conspicuous, can be attributed to settling, while the measurable ablation which occurs is caused by deflation of the relatively soft snow of recent deposition. The permanence of the surface over long periods was largely due to its hardness caused by wind compaction, and this permanence demonstrates the insignificance of direct evaporation, a fact noted by Wade (1945, p. 169), who used the term ablation only for evaporation, Swithinbank (1957*b*, p. 64) and Lister (1960, p. 38).

The frequency at which stakes are read has a marked effect on the conclusions which can be drawn from the results. Hourly readings give a complex trace when accumulation is plotted against time, while monthly readings give a simple one. Because of this it is always necessary to consider accumulation and ablation in relation to a specific time period. The annual net accumulation of snow at Halley Bay is of the order of 75 cm., while the annual net ablation is zero. However, the amount of ablation in any one day may be quite large and unless a continuous recording of the variations in surface level is kept it is meaningless to sum the positive or negative changes shown at the stakes. These positive and negative variations in the surface level at a given point may be almost continuous, especially during a blizzard. Also, the micro-morphology and the broader surface morphology can cause erosion and deposition in relatively close proximity to one another at the same instant. Both Wade (1945, p. 169) and Howard (1948, p. 920) have made this mistake, producing an accumulation/ablation ratio dependent upon their daily data.

Changes in the surface level of any magnitude are largely related to the occurrence of storm winds from the east-north-east. The consistency of the wind direction is very marked and winds from any other direction are rare; those from the south-south-west are the only other ones of any consequence and then they are always weak. The mean daily wind speeds, obtained from each day's recorded trace, are shown in conjunction with the accumulation trace. In Fig. 11 the long storms all of which came from the east are very conspicuous. Each of these, in March-April, August and November 1960, and in March, May, October and December 1961, resulted in accumulation. However, the amount of accumulation cannot be correlated directly with wind speed because on some occasions the result of a blizzard was net ablation. In spite of this, it can be seen that much of the accumulation is related to storm winds from the east-north-east and therefore to the depressions which cause them. Lamb (1952, p. 33) has deduced that cyclonic situations in Antarctica result from the occurrence of westerly winds around the coast, from the presence of low pressures in high latitudes, especially in the Ross and Weddell Sea areas, and from the development of cloud types, other than stratus and strato-cumulus with indications of subsidence above, including extensive frontal up-gliding cloud systems and instability types. He has estimated that for the Halley Bay area the average frontal frequency is approximately every 10 days, and he has suggested that the distribution of precipitation must be related to the longer-period frequency distribution of frontal and cyclonic activity, to the moisture content of air masses carried inland and to the unknown orography of the interior. The commencement of the storm on 22 November 1960 was marked by the passage of a frontal cloud formation.

The accumulation measurements at Halley Bay were made on an almost level area of great extent which is subjected to uniform winds and uniform precipitation. The relationship between accumulation and probable precipitation, in the absence of any other appreciable source (hoar frost deposition is insignificant) is very marked, and Swithinbank's (1957*b*, p. 73) conclusion, that under these circumstances, and therefore over most large areas of flat terrain in Antarctica, the amount of accumulation must approximate to the amount of precipitation, seems justified. Windborne drift snow is regarded as an endless belt, being lifted, moved along and then re-deposited elsewhere with no net change in surface level taking place. Lister (1960, p. 37) has concluded that, if air is supersaturated or undersaturated with drift particles at a particular wind speed, either deposition or erosion is caused at the snow surface.

Stratigraphy

Prior to the Norwegian-British-Swedish Antarctic Expedition, 1949–52, all attempts to establish an annual stratification in Antarctica had failed, but annual strata had been identified and measured in the Arctic. Schytt (1958*a*, p. 9–19) has given a very comprehensive review of these previous stratigraphical observations and has described the determination of annual strata on the Maudheim Ice Shelf and the adjacent inland ice.

Unlike temperate glaciers, most of Antarctica has no distinct accumulation and ablation seasons; over a large proportion of its area it has little or nothing to act as a source of material for the production of dirt or marker bands. The determination of the annual stratification is therefore relatively difficult.

The variations in the type of snowflake which accumulate on the surface have little connection with the textures of the snow strata below. Throughout most of the year snowflakes of the simple type are predominant, but in summer more complex forms occur and on several occasions in summer graupel, or soft hail, was observed, being derived from cumulo-nimbus cloud developments. The latter were associated with winds from the south-south-west.

The investigations of snow stratification at Halley Bay were carried out in pits. Two major pits were dug, one to a depth of 2 m. on 31 January 1962 and a deep pit which was commenced in May 1961. Both of these pits were south-east of the station buildings in areas unaffected by drifting. A cold laboratory, a floorless hut approximately 2·5 m. square and 2 m. high, was erected on the proposed site for the deep pit in May 1961 and the pit was dug from within the hut, snow being removed via a hatch in the roof. A depth of 12·2 m. was reached and cores were extracted to a depth of 15 m. Stratigraphical observations and grain-size measurements were made on the pit walls and in the blocks cut for density determinations. The grain-size was measured with a scale and magnifier. Below 12·2 m. only density measurements were made because core recovery was poor and the maximum unbroken length rarely exceeded 3 cm.

The pit dug on 31 January 1962 revealed strings which had been pegged to the surface on various occasions during the preceding two years, and also a winter horizon which was defined by the presence of cocoa powder which had been scattered thinly on the surface during the season of darkness in 1960. These markers served to indicate the age of the strata in which they occurred. In this pit, as in those dug while sledging, density measurements were made by weighing the contents of a snow-sampler. For the deep pit the method of measuring the volume of firn blocks was that described by Schytt (1958*a*, p. 24). These were weighed and the density computed. In all pits individual strata often showed a horizontal variation and fossil sastrugi were occasionally observed.

The identification of annual layers was facilitated by the occurrence of depth hoar layers (Seligman, 1962, p. 62–73) or highly metamorphosed firn, which were interpreted as summer surfaces. It was impossible to differentiate between snow deposited in summer and snow

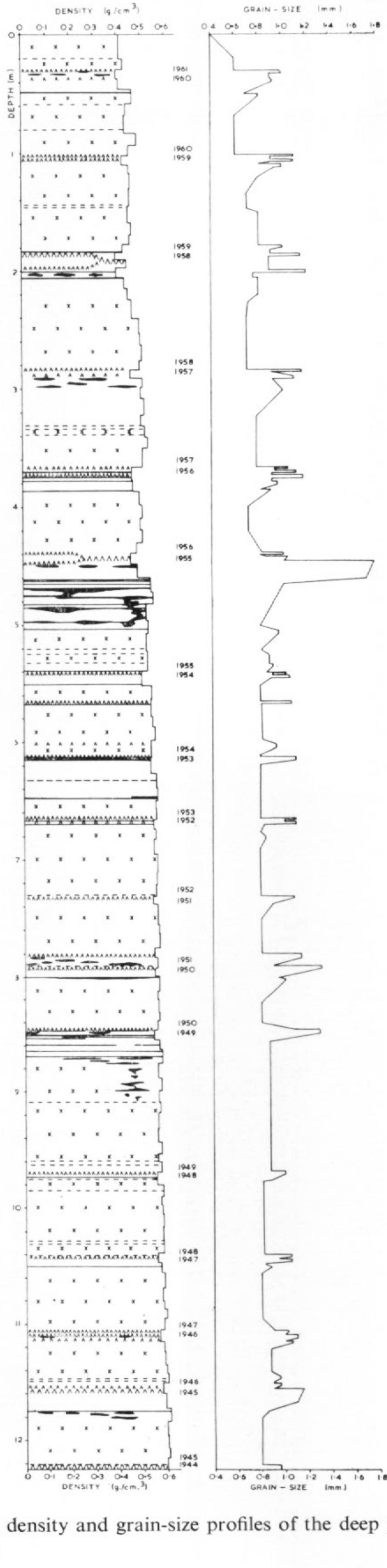

Fig. 13. The snow stratigraphy, density and grain-size profiles of the deep pit at Halley Bay. The key to the symbols is given in Fig. 12.

deposited in winter which had become coarser-grained near the surface in summer. Because of the doubt as to when snow was deposited, summer surfaces are not referred to as summer layers, a term implying deposition in summer, despite having a certain depth. In some cases summers are interpreted as having more than one long spell of good weather, since more than one depth hoar layer is present.

Ideal conditions for the formation of depth hoar occur when densely packed, fine-grained autumn snow overlies the relatively warm, coarse-grained summer snow (Quervain, 1963, p. 380–81). The temperature of the autumn snow is lower than that below and therefore

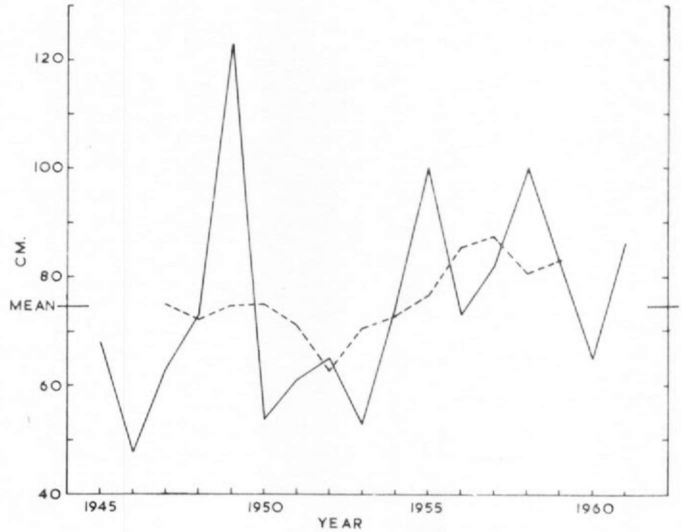

Fig. 14. The annual accumulation for the period 1945–61 with superimposed 5-year running means.

a vapour pressure gradient is established, ranging from high in the warm snow below to low in the cold layer above. Evaporation from the warmer level and condensation of water vapour in the colder newer snow results; this normal diffusion of water molecules is assisted by convection currents. Condensation is aided by the cold, fine-grained, densely packed snow inhibiting the movement of the water vapour, while ice layers and wind crusts, which often occur below the depth hoar layers, prohibit the flow of water vapour from below. The depth hoar horizon is therefore subject to evaporation with no compensating condensation. In addition, it is also situated where the temperature gradient is at a maximum and therefore where metamorphism is most intense. The net result of these processes is a stratum characterized by large, often hexagonal crystals, possessing little cohesion, and by relatively low density values.

As well as the alternation of the dense fine-grained winter snow with the coarser snow and depth hoar of the summers, the alternation of soaked and unsoaked firn is very characteristic of the stratigraphic profile. The soaked firn contains lenses, layers, laminae and pellets of ice. The ice layers sometimes occur where the downward percolation of water has been arrested by the presence of a wind crust. In some years very little melting has taken place but in others almost the whole of the annual accumulation may be soaked by downward percolation of summer surface melt water. The occurrence of unsoaked strata directly above a summer surface indicates that the former is probably an autumn deposit which grades upwards into the winter accumulation.

The coarsening of the summer snow is the direct result of the metamorphic processes described above and therefore the varying grain-size alone gives little indication of the time of year in which individual snow strata were deposited.

As the alternating character of the stratigraphy depends upon variations in the weather, the time between the formation of two summer surfaces (an accumulation year) does not necessarily coincide with a calendar year. In one instance a spell of good weather giving rise to a summer surface may occur early in the summer, whereas in another it may be late. This may lead to somewhat misleading comparisons between individual annual increments, but long-term averages are relatively unaffected.

TABLE IV. ANNUAL ACCUMULATION MEASUREMENTS

Deep pit (commenced May 1961)

Year	*Depth* (cm.)	*Accumulation* (cm.)	*Water Equivalent* (cm.)	*Density* (g./cm.^{3})
1961	0–37	86	35·85	0·416
1960	37–102	65	29·01	0·447
1959	102–184	82	36·72	0·448
1958	184–284	100	46·16	0·462
1957	284–366	82	41·21	0·502
1956	366–439	73	35·32	0·484
1955	439–539	100	51·50	0·515
1954	539–613	74	39·56	0·536
1953	613–666	53	29·29	0·552
1952	666–731	65	36·09	0·556
1951	731–792	61	34·23	0·561
1950	792–846	54	30·41	0·563
1949	846–969	123	69·82	0·566
1948	969–1,042	73	42·02	0·577
1947	1,042–1,105	63	36·60	0·581
1946	1,105–1,153	48	28·15	0·587
1945	1,153–1,221	68	40·84	0·601
1944	1,221	—	—	—
TOTAL		1,270	662·78	—
MEAN		74·71	38·99	0·523

In the following discussion of the annual accumulation the stratigraphy is considered in relation to each accumulation year. This stratigraphy is illustrated diagrammatically in Figs. 12 and 13, the former representing the strata observed in the 2 m. pit of 31 January 1962 and the latter that of the deep pit. The annual accumulation measurements are given in Tables IV and V (Fig. 14), and for 1960 and 1961 these are compared with the stake measurements and the other pit at Halley Bay in Table VI. The top of the deep pit is equivalent to a depth of 49 cm. in the 2 m. pit.

TABLE V. 5-YEAR RUNNING MEANS OF ANNUAL ACCUMULATION MEASUREMENTS

Year	*Accumulation* (cm.)	*Water Equivalent* (cm.)	*Density* (g./cm.³)
1959	83·00	37·79	0·455
1958	80·40	37·68	0·469
1957	87·40	42·18	0·482
1956	85·80	42·75	0·498
1955	76·40	39·38	0·515
1954	73·00	38·35	0·525
1953	70·60	38·13	0·540
1952	61·40	33·92	0·552
1951	71·20	39·97	0·561
1950	75·20	42·51	0·565
1949	74·81	42·62	0·569
1948	72·20	41·40	0·573
1947	75·00	43·49	0·579
TOTAL	986·41	520·17	—
MEAN	75·88	40·01	0·527

TABLE VI. COMPARATIVE ACCUMULATION MEASUREMENTS

Year		*Stake Pattern*	*Pit of 31 January 1962*	*Deep Pit*
1961	Accumulation	87·5 cm.	86·0 cm.	
	Density		0·416 g./cm.³	
	Water equivalent		35·85 cm.	
1960	Accumulation	67·9 cm.	61·0 cm.	65 cm.
	Density		0·443 g./cm.³	0·447 g./cm.³
	Water equivalent		27·03 cm.	29·01 cm.

1961. Two surfaces gave rise to depth hoar horizons in the summer of 1960–61. The first developed after the storm of November 1960 and was present until January 1961, while the second, in which there are holes similar to that illustrated by Schytt (1958*b*, pl. 3b), was present in February and March. A string pegged to the surface in January, in the manner described by Lister (1956, p. 235), was found between these two layers. The top of the lower depth hoar layer at 86 cm. marks the bottom of the 1961 accumulation, thereby agreeing well with the accumulation of 87·5 cm. measured at the stake patterns. In the deep pit only one marked depth hoar layer was present below the 1961 accumulation with several lenses of ice beneath, while a coarse highly metamorphosed band matches the second depth hoar layer of the smaller pit. An ice lamina at 1·03 m. in the small pit and at 49 cm. in the deep pit indicates the extent of melt-water percolation; below this the firn is unsoaked.

1960. The strings present at depths of 1·09 and 1·03 m. in the small pit are separated by snow accumulated in the storm between their respective establishment on 1 and 11 August. That laid down on the second date remained partially exposed at the surface until November. Cocoa powder which was scattered on the surface on 15 May occurred at 1·17 m. The depth hoar layers at a depth of 1·47 m. in the small pit and at 1·02 and 1·05 m. in the deep pit are assumed to represent the 1959–60 summer surface. Little melting appears to have taken place during this summer as the occurrence of ice bodies is very limited and the unsoaked firn is only 7 cm. below the depth hoar horizon.

1959. 65 cm. of even-grained unsoaked firn occur beneath the 7 cm. of firn containing ice pellets below the 1959–60 summer surface. At a depth of 1·78 m. in the deep pit there is an increase in grain-size and below this is a thin ice layer. This overlies a depth hoar layer which shows horizontal variation where fossil sastrugi have caused it to split and form two separate strata, which are in places up to 14 cm. apart between 1·85 and 1·99 m. Where these depth hoar layers converge their total thickness increases to over 3 cm. The presence of the sastrugi indicates that a long spell of warm weather was broken by a storm, after which the original surface was in part covered by fresh deposition and in part left clear.

1958. Underlying the lower, or combined, depth hoar stratum ice layers occur at 2·01 and 2·05 m. Between these layers lenses of ice occur, while rare ice pellets were observed down to 2·11 m., beneath which, unsoaked firn is present to the depth hoar layer at 2·84 m.

1957. Two depth hoar layers at 2·84 and 2·90 m. indicate the occurrence of separate warm spells, while the extensive evidence of melting indicates the overall warmness of the 1957–58 summer season. Ice lenses and pellets are numerous between 2·91 and 2·98 m. but the latter become increasingly rare until the unsoaked firn is reached at 3·30 m. Three wind crusts occur in the winter snow at 3·31, 3·34 and 3·39 m.

1956. The 1956–57 summer is marked by the development of three depth hoar layers. The uppermost pair at 3·66 and 3·69 m. are separated by a relatively coarse layer and are underlain by a thin ice layer. The third depth hoar layer is at 3·74 m. below which are three ice laminae at 3·76, 3·78 and 3·87 m. This whole assemblage has a very open texture and seems to indicate a warm summer with a little melting during which two storms caused some accumulation. Beneath this a homogeneous stratum of unsoaked winter snow continues down to the previous summer surface at 4·39 m.

1955. As in the summer of 1958–59, the depth hoar stratum of the 1955–56 summer is divided in places into two individual layers up to 8 cm. apart. Beneath this depth hoar the firn is characterized by an exceptional coarseness and the presence of a large proportion of ice bodies. This marked downward percolation of melt water soaked 65 per cent of the year's accumulation, indicating that the summer must have been exceptionally warm. Below an ice lamina at 5·04 m. the unsoaked firn continues down to a depth of 5·33 m., where an increase in grain-size and the occurrence of occasional ice pellets has been interpreted as a late summer surface.

1954. An ice lamina at 5·39 m. underlain by four coarse layers, the uppermost of which possesses more marked depth hoar characteristics, has been assumed to represent the summer surface. An ice lamina at 5·51 m. marks the top of unsoaked firn which continues down to another highly metamorphosed horizon at 5·67 m. This lies immediately above another ice lamina and must represent an early summer surface. The 1954–55 summer therefore appears to have had three markedly warm spells, the middle one of which was predominant, separated by periods of accumulation. The firn below this spring surface is unsoaked but a relatively coarse, open-textured layer at 6·00 m. is interpreted as an autumn surface.

1953. The depth hoar horizon marking the 1953–54 summer surface at 6·13 m. is immediately underlain by three ice layers. The firn is soaked down to 6·48 m. with an ice lamina at 6·33 m. and an ice layer immediately above the unsoaked firn. This layer changed its horizon at several places in the pit walls.

1952. Apart from thin ice laminae associated with the depth hoar horizon at 6·66 m., the accumulation of 1952 is completely unsoaked homogeneous firn, indicating that the summer of 1952–53 was relatively cold in contrast with that of the following year.

1951. The summer surface of 1951–52 is marked by a poorly developed stratum of depth hoar immediately below a wind crust at 7.30 m. No melting could be discerned and the implication is that no surface was preserved for long during the summer; the one which developed into a depth hoar horizon was inhibited by the strong wind crust lying immediately above which acted as a barrier to the transfer of water vapour. The highly metamorphosed stratum at 7·82 m., below which numerous ice lenses have developed, is regarded as a late summer surface, the depth hoar layer at 7·92 m. being interpreted as the major 1950–51 summer surface.

1950. The wind crust under which the main depth hoar horizon occurs has been enhanced by the addition of melt water from above and is now an ice lamella. Further melting took place below this depth hoar layer and an ice layer of varying thickness occurs at 8·01 m. This stratigraphy suggests two long warm spells, in both of which melting took place, separated by a stormy period in which the firn between 7·82 and 7·90 m. accumulated. The melt water of the second warm period appears to have soaked this accumulation down to, but not through, the wind crust immediately above the depth hoar horizon. Below the ice layer at 8·01 m. a homogeneous unit of winter snow continues down to the previous summer surface.

1949. The depth hoar stratum at 8·46 m., which represents the 1949–50 summer surface, overlies a heavily soaked unit containing many ice bodies. These are evenly distributed down as far as 8·69 m. but in one place melt water has percolated down to a wind crust at 9·09 m. This isolated soaking of the firn occurs alongside unsoaked even-grained winter snow which continues with no evidence of any warm periods down to the 1948–49 summer surface at 9·69 m. Therefore, the large accumulation (1·23 m.) attributed to 1949 seems justified, because the presence of any other depth hoar strata has not been obscured. The single depth hoar stratum and the extensive development of ice bodies beneath it imply one long warm period during the summer of 1949–50.

1948. A poorly developed depth hoar stratum defines the summer surface of 1948–49; little melting is associated with it and most of the year's accumulation is unsoaked.

1947. The summer surface at 10·42 m. underlies a wind crust on which the unsoaked 1948 winter snow occurs. Little melting has taken place, the only discernible ice body being a thin lamina at 10·50 m. below which the firn is homogeneous and even-grained down to the previous depth hoar stratum at 11·05 m.

1946. A thick highly metamorphosed layer containing isolated lenses of ice forms the 1946–47 summer surface between 11·05 and 11·13 m. Below this unsoaked firn overlies two late summer surfaces at 11·43 and 11·49 m. Each of these is relatively highly metamorphosed and overlain by a thin wind crust.

1945. Below the autumn surfaces described above, the strata remain very coarse down to 11·73 m., with a well-developed depth hoar horizon between 11·53 and 11·58 m. Further evidence of the warmth of the summer is the presence of an ice layer at 11·73 m. with many fairly large ice lenses below it. The unsoaked winter snow continues to a depth of 12·21 m. where the 1944–45 summer surface occurs.

1944. Only the uppermost part of the 1944 accumulation is exposed in the deep pit; this is a depth hoar layer overlain by a wind crust. It was not possible to discover whether any melting had taken place.

TABLE VII. RATES OF SETTLING IN FIRN

Actual Depth 20 October 1961 (m.)	*Interval* 20 October 1961 (cm.)	*Interval* 25 January 1962 (cm.)	*Rate of Settling* (cm. m.$^{-1}$ yr.$^{-1}$)
0			
	100	91·3	43·35
1			
	100	97·6	8·68
2			
	100	98·1	7·16
3			
	100	99·0	3·77
4			
	100	98·8	4·52
5			
	100	99·1	3·49
6			
	100	98·9	4·14
7			
	100	100·0	0·00
8			
	100	99·4	2·26
9			
	100	99·9	0·38
10			

Settling

The measurements of settling of the firn in the deep pit were made to a depth of 10 m. by measuring the changes in distance between markers placed at 1 m. intervals in the wall of the pit. The markers were 11·5 cm. oval section nails driven into the firn with their major axes in a horizontal plane. The distances recorded were between crosses filed on the heads of the nails and the period between measurements was 97 days.

The measurements and the rates of settling for each metre interval are shown in Table VII. The overall rate of settling was found to be 7·77 cm. m.$^{-1}$ yr.$^{-1}$, but most of this took place in the upper 2 m. and the settling rate for the 2–10 m. interval is 3·22 cm. m.$^{-1}$ yr.$^{-1}$.

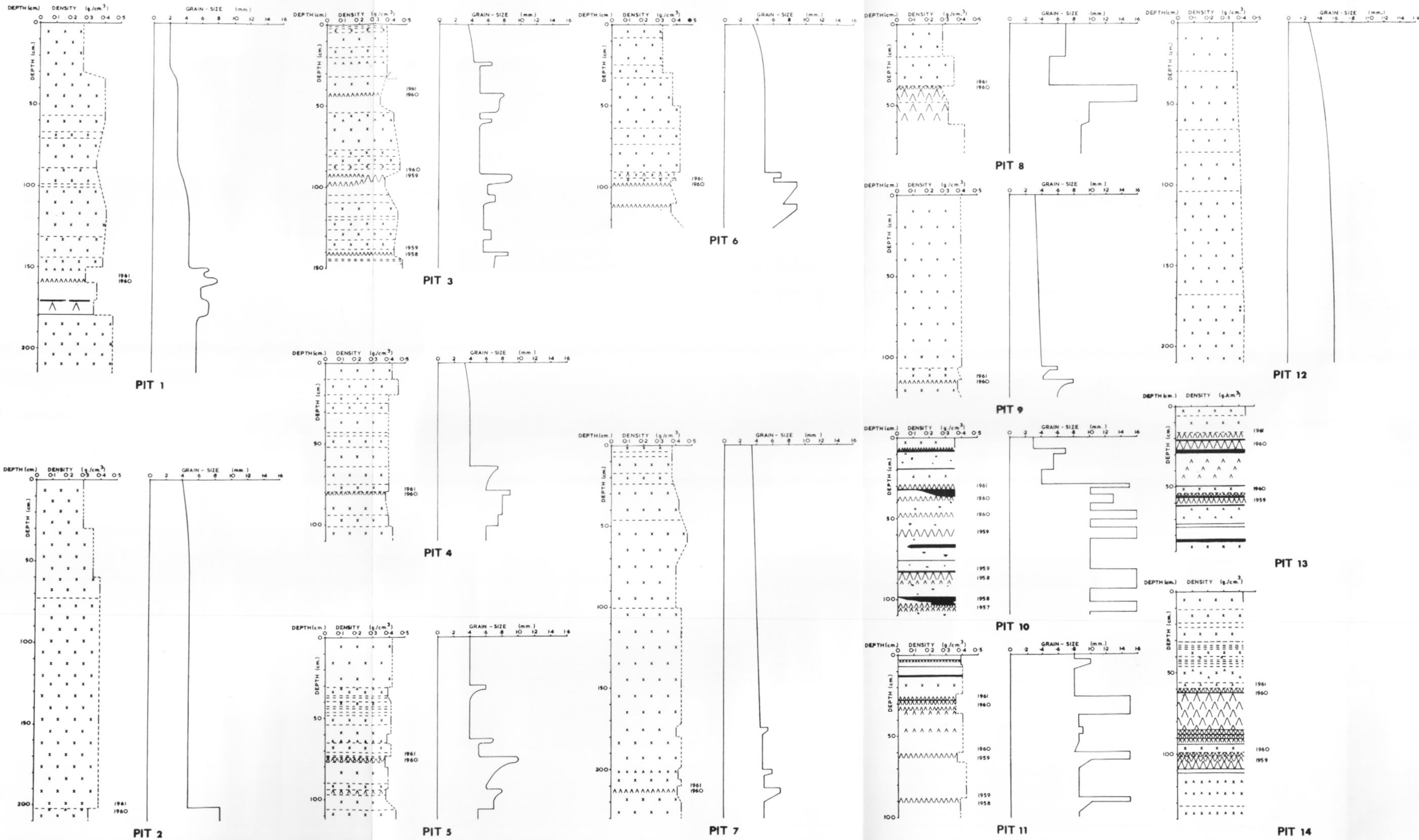

Fig. 15. The snow stratigraphy, density and grain-size profiles of pits 1–14 which were studied during sledge journeys. The key to the symbols is given in Fig. 12.

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Snow stratigraphy studies during sledge journeys

In order to determine annual snow accumulation values for different parts of the Brunt Ice Shelf and the adjacent inland ice the stratigraphy of a number of pits was examined. The observations made in 14 pits are described and interpreted below. The positions of pits 1–14 are shown in Figs. 1 and 2, while the stratigraphy, density and mean grain diameter for each pit are illustrated diagrammatically in Fig. 15.

Pit 1 (27 October 1961). On a flat area of low ice shelf 200 m. east of the junction with normal ice shelf. The stratum down to 33 cm. was deposited during a storm on 26 October. Below this there is unsoaked snow with occasional wind crusts to the two 1960–61 summer surfaces at 1·51 and 1·57 m. These depth hoar layers are associated with some melting, the resulting melt water having percolated to the ice lamina at 1·79 m., and ice lenses up to 0·3 cm. thick occur at 1·71 m. The two depth hoar layers correspond to those observed at depths of 0·78 and 0·86 m. in the pit dug at Halley Bay on 31 January 1962. The top of the depth hoar layer at 1·57 m. may therefore be assumed to represent the 1960–61 summer surface, while that at 1·51 m. probably developed during March and April 1961. The increased net accumulation rate shown by pit 1, almost twice that at Halley Bay despite the shorter time period, was found to be characteristic of all low ice shelf areas.

Pit 2 (3 November 1961). In a slight hollow between gentle undulations 458 m. above sea-level on the inland ice 20 km. south-east of "Cape N". The 1960–61 summer surface occurs at a depth of 2·02 m. Above this unsoaked firn of regular grain-size extends to the surface and no melting has taken place. The summer surface does not possess notable depth hoar characteristics, being marked by an increase in grain-size and a decrease in density. The large amount of accumulation recorded appears to have been caused by the situation on the side of a hollow facing up-wind.

Pit 3 (9 November 1961). 850 m. above sea-level on the west side of a gentle undulation. Well-developed depth hoar horizons representing the major summer surfaces occur at 0·43, 0·92 and 1·40 m. In addition, coarse horizons at 0·23 and 0·58 m. indicate warm periods during the autumn and spring of the 1960–61 summer. Coarse horizons between 1·32 and 1·26 m. and between 1·09 and 1·05 m. also represent warm periods during the autumn and spring of 1959. The depth hoar horizon of the 1959–60 summer, which occurs at depths between 0·92 and 0·99 m., in places bifurcates to form two separate strata due to the interruption of a long spell of warm weather by a storm, resulting in accumulation over part but not all of the surface.

Pit 4 (18 November 1961). 1,360 m. above sea-level 37·5 km. from Tottanfjella in an area sloping gently to the west. The main summer surface of 1960–61 is indicated by the presence of a depth hoar horizon at 0·79 m. A coarse-grained stratum at 0·62 m. represents the surface present during a warm spell in the autumn.

Pit 5 (19 November 1961). 1,040 m. above sea-level in a flat area with moderate sastrugi. A depth hoar horizon at 0·74 m. is assumed to be the 1960–61 summer surface, while coarse layers at 0·64 and 0·94 m. represent autumn and spring surfaces. The coarse horizon at 0·31 m. probably represents an early spring surface, prior to the 1961–62 summer, with the numerous wind crusts developed beneath it being caused by winter storm winds.

Pit 6 (27 November 1961). At the boundary of the inland ice and the Brunt Ice Shelf between the *en échelon* chasms developed in this area. The pit was dug on the summit of a gentle rise with strand cracks operating near the site. Two depth hoar layers at depths of 0·97 and 1·01 m. represent the 1960–61 summer surfaces, the first of which probably developed in spring or early summer, while a coarse open layer at 0·91 m. is probably the result of a warm period in late summer or autumn.

Pit 7 (29 November 1961). On a gentle slope adjacent to and up-wind from one of the *en échelon* chasms at "Cape N". Two 1960–61 summer surface horizons occur at depths of 2·01 and 2·12 m., the second of which has the more strongly developed depth hoar characteristics and is presumably the main summer surface, while the first probably formed slightly later. The coarse layer between 1·75 and 1·77 m. has probably resulted from a warm autumn spell preceding the winter accumulation.

Pit 8 (2 December 1961). On the west side of a pressure ridge in an area of low ice shelf, 1·6 km. east from the junction with normal ice shelf to the west. The pit was situated on the crest of a gentle undulation. The 0·37 m. of accumulation above the depth hoar layer at this depth, which is thought to be that of the 1960–61 summer, is atypical of areas of low shelf and is presumably due to the situation of the pit on a slight rise. The depth hoar horizon at 0·37 m. is thought to represent a surface which had been exposed for a long period before being covered by the following winter snow. It possesses idiomorphic crystals up to 2·5 mm. in diameter, and overlies very coarse open strata in which there are many idiomorphic crystals and occasional holes. The accumulation below 0·71 m. is presumably that of the 1960 winter.

Pit 9 (5 December 1961). On a flat area of low ice shelf adjacent to the large block of normal ice shelf described on p. 18, and close to the inlet which bounds it to the north. Large snowdrifts and sastrugi of moderate size occur in the area. The stratigraphy of this pit implies fairly continuous deposition since the formation of the late summer surface at 0·97 m. A second coarse layer at 1·04 m., possessing better developed depth hoar characteristics, is assumed to represent the major 1960–61 summer surface.

Pit 10 (5 December 1961). On the crest of a gentle rise on the normal ice shelf close to pit 9. Apart from occasional transient snow drifts on which small sastrugi had developed, the surface was highly metamorphosed, soft, coarse and fairly open. Apart from the upper 0·29 m., the strata of this pit have been extensively metamorphosed with the consequent development of a large number of ice bodies. The 7 cm. of fine-grained unsoaked snow at the top of the pit represent the cross-section of one of the snow drifts mentioned above. Depth hoar layers occur at 0·07, 0·29, 0·36, 0·47, 0·57, 0·84 and 1·03 m., those below and including that at 0·47 m. being extremely well developed, with a very open, hanging structure and many idiomorphic crystals up to 3 mm. in diameter. The average grain diameter below 0·29 m., excluding the depth hoar layers, is 1·0 mm.

The depth hoar horizon at 7 cm. depth is interpreted as a spring surface of the 1961–62 summer, while the accumulation below this to a depth of 0·29 m. is the 1961 winter accumulation. This had already been soaked down to the ice layer at 0·19 m. by early summer melting. The two depth hoar horizons at 0·29 and 0·36 m. are regarded as late and early 1960–61 summer surfaces, considerable melting having taken place at the upper one with the formation of a large ice body up to 4 cm. thick. The two summer surfaces, assumed to be of 1959–60, are associated with thick ice layers throughout the preceding winter accumulation. Similarly, the depth hoar horizon at 0·84 m., assumed to be the 1958–59 summer surface, overlies a soaked winter unit with a thick ice layer at 0·99 m., which in places thickens and extends down to the 1957–58 summer surface at 1·03 m.

Pits 11 and 12 (8 December 1961). Pits 11 and 12 were dug at the McDonald Ice Rumples where the grounded ice shelf has undergone deformation, horizontal compression having caused the ice shelf to fold into anticlines cut by transverse crevasses. Pit 11 was situated on the crest of the anticline next but one to the main ridge, with pit 12 in the inner adjacent depression.

Pit 11. The depth hoar horizon at 3 cm. represents the surface present during the long warm period in November 1961, melting associated with this surface having formed two ice layers at 0·07 and 0·13 m. The accumulation below the latter was unsoaked down to the previous summer surface at 0·25 m. Three depth hoar horizons at 0·25, 0·29 and 0·33 m., the last one possessing markedly idiomorphic crystals up to 4·0 cm. in diameter, represent summer surfaces of 1960–61, while open layers of relatively coarse grain at 0·36 and 0·47 m. suggest that warm periods occurred during the spring and late autumn of 1960. The 1959–60 and 1958–59 summer surfaces are at 0·60 and 0·89 m. and in each case their depth hoar characteristics are strongly developed.

Pit 12. In contrast with pit 11, with its overall appearance of having been subjected to much metamorphism, the uniform appearance of the unsoaked strata in pit 12 was very marked. Grain-size shows a steady but slight increase with depth, as does density, although in the latter case a slight change occurs below the uppermost wind crust at 0·30 m. This suggests that the accumulation down to this depth was of recent deposition. No previous summer surfaces were found down to the bottom of this pit at 2·10 m., the total accumulation being of the year 1961.

Pit 13 (17 November 1961). Dug by M. Bethel and E. Thornton. On the inland ice 3·2 km. from the boundary with the Brunt Ice Shelf, 110 m. above sea-level. Depth hoar horizons at 0·17 and 0·23 m., each overlying an ice layer, are assumed to represent the 1960–61 summer surfaces. Melt water has percolated through much of the 1960 winter snow down to an ice layer at a depth of 0·50 m. The strata between 0·17 and 0·50 m. are all metamorphosed and fairly coarse-grained. Two more depth hoar layers at 0·54 and 0·58 m., each underlain by an ice layer and the former immediately beneath a wind crust, represent the 1959–60 summer surfaces. The strata beneath are again relatively coarse down to 0·75 m., and beneath this reduced grain diameter and increased hardness imply winter snow. This has been soaked to at least a depth of 0·84 m., where there is an ice layer.

Pit 14 (20 November 1961). Dug by M. Bethel and E. Thornton. On the inland ice 9·7 km. from the Brunt Ice Shelf at an altitude of 430 m. above sea-level. This pit is characterized by the extreme coarseness of the strata below a depth of 0·59 m. Between 0·62 and 0·85 m. very coarse depth hoar crystals overlie an even coarser 3 cm. thick depth hoar layer. This, in turn, overlies a sequence of fairly coarse strata down to the bottom of the pit at 1·40 m. The presence of numerous wind crusts down to the depth hoar horizon at 0·59 m. is indicative of the frequency and strength of the katabatic winds which descend from the inland ice to the ice shelf. The 1961 winter snow overlies the strongly metamorphosed 1960–61 summer snow in and below which there are thin ice layers. It is possible that the coarse horizon at 1·01 m. represents the 1959–60 summer surface overlying the 1959 accumulation which is also metamorphosed.

CONCLUSIONS

The Brunt Ice Shelf and Riiser-Larsen Isen to the north-east form an ice shelf approximately 75,000 km.[2] in area, which extends from "Dawson-Lambton Glacier" north-eastward to Kapp Norvegia. The extent of this ice shelf is largely due to the protection afforded by the McDonald Ice Rumples. North and north-west of "Cape N" large areas of low ice shelf are actively forming within the ice shelf as a direct result of divergent flow which has surpassed the stage of tension valley development. This process has not been described elsewhere, although it is essentially a variation of the ice shelf formed originally from land ice extensions, with or without interstitial sea ice, described by Wright and Priestley (1922, p. 164).

The amount of slope of the inland ice adjacent to the ice shelf has a great influence on the morphology of the ice shelf in the area adjacent to the inland ice boundary. The development of the chasm or rift separating the ice shelf from the inland ice, which is characteristic of the Brunt Ice Shelf, and the formation of "icebergs" are attributed to the steepness of the slope of the inland ice causing fracturing along the line of flotation. This steepness of slope is enhanced by varying rates of movement between ice streams and intervening areas.

Inland, Tottanfjella have attained their present form following the erosion of an escarpment formed by block-faulting in conditions similar to those of the present day, except for a former greater thickness of ice cover. The ice cover now appears to be in a state of equilibrium. North of Tottanfjella a broad valley trends westward to Riiser-Larsen Isen. Milorgknausane and "Nunataks L" are situated on the northern and southern slopes of this valley.

Measurements of accumulation, both in pits and against stakes, show a great uniformity over the area, except where the influence of local topographical conditions predominate, e.g. in areas of low ice shelf, near the ice front and in undulations. Because of snow drifting, east-facing slopes receive excessive accumulation, while west-facing ones receive reduced amounts. However, at Tottanfjella these conditions are reversed on very steep slopes: the east-facing ones are usually bare, while the lee sides of mountains are heavily glacierized. At Halley Bay the net accumulation is considered to be representative of the true precipitation and the annual measurements for 1945–61 are probably a good measure of the precipitation over the whole area.

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