COMMONWEALTH OF AUSTRALIA
DEPARTMENT OF EXTERNAL AFFAIRS

AUSTRALIAN NATIONAL ANTARCTIC RESEARCH EXPEDITIONS

ANARE SCIENTIFIC REPORTS

SERIES A (IV) GLACIOLOGY

PUBLICATION No. 97

MASS ECONOMY OF ANTARCTICA: MEASUREMENTS AT MAWSON, 1957

by
M. MELLOR

ISSUED BY THE ANTARCTIC DIVISION
DEPARTMENT OF EXTERNAL AFFAIRS, MELBOURNE
1967

PUBLICATIONS
OF THE
ANTARCTIC DIVISION, DEPARTMENT OF EXTERNAL AFFAIRS

Reports are now published in three categories:

ANARE SCIENTIFIC REPORTS
ANARE INTERIM REPORTS
ANARE DATA REPORTS

These are classified as follows:

Series A	(I)	Narrative
	(II)	Geography
	(III)	Geology
	(IV)	Glaciology
Series B	(I)	Zoology
	(II)	Botany
	(III)	Oceanography
	(IV)	Medical Science
Series C	(I)	Terrestrial Magnetism
	(II)	Cosmic Rays
	(III)	Upper Atmosphere Physics
Series D		Meteorology

In general, it is the policy of the Division to publish in the regular scientific journals all articles suitable for such media. The ANARE SCIENTIFIC REPORTS provide a medium for publication of comprehensive works and for specialized studies not catered for by appropriate journals. The ANARE INTERIM REPORTS cater for scientific and technical articles, descriptions of instruments etc. which, because of the preliminary nature of the work or for other reasons, are likely to be of relatively lower long-term importance. The ANARE DATA REPORTS contain numerical or graphical data with only very limited discussion.

Copies of Reports and a list of Reports published are available from the Publications Officer, Antarctic Division, Department of External Affairs, 568 St. Kilda Road, Melbourne, Victoria 3004, Australia.

COMMONWEALTH OF AUSTRALIA
DEPARTMENT OF EXTERNAL AFFAIRS

AUSTRALIAN NATIONAL ANTARCTIC RESEARCH EXPEDITIONS

ANARE SCIENTIFIC REPORTS

SERIES A (IV) GLACIOLOGY

PUBLICATION No. 97

MASS ECONOMY OF ANTARCTICA: MEASUREMENTS AT MAWSON, 1957

by
M. MELLOR

ISSUED BY THE ANTARCTIC DIVISION
DEPARTMENT OF EXTERNAL AFFAIRS, MELBOURNE
1967

Registered at the G.P.O. Melbourne
for transmission by post as a book.

Printed in Australia by
Brown Prior Anderson Pty Ltd 5 Evans Street Burwood Victoria

CONTENTS

LIST OF FIGURES

MAPS

LIST OF PLATES

Malcolm Mellor was the first professional glaciologist appointed by the Antarctic Division for ANARE and created the Australian glaciological research programme for the IGY.

Many of the studies initiated by him and described in this publication were continued, and some remain active to this day.

It is for this reason that the present paper has seemed eminently worth publishing even ten years after it was written.

MASS ECONOMY OF ANTARCTICA: MEASUREMENTS AT MAWSON, 1957

By

M. Mellor

Antarctic Division, Department of External Affairs, Melbourne
(Now with United States Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire)

ABSTRACT

The results of glaciological measurements made by the author at Mawson over the IGY are presented and analyzed to give the first detailed estimate of the mass budget of this region of Antarctica. By incorporating the data for other regions, an estimate of the mass budget of the whole Antarctic ice cap is made. The estimate is based on data available at the time of writing in 1959.

A brief survey of the physiography and climate of the Mawson region is presented.

Several years' results of ablation measurements indicate that the annual loss of ice varies from about 50g cm^{-2} at the coast to 10g cm^{-2} 20km inland near the firn limit approximately at 500m elevation. Summer melting accounts for most of the annual ablation but there is loss from ice evaporation all through the year.

Snow accumulation data from stake readings and pit studies to 650km inland show an average accumulation rate of about 12g cm^{-2} yr^{-1} over this region. By comparison with other regions the total net annual gain to the Antarctic is estimated.

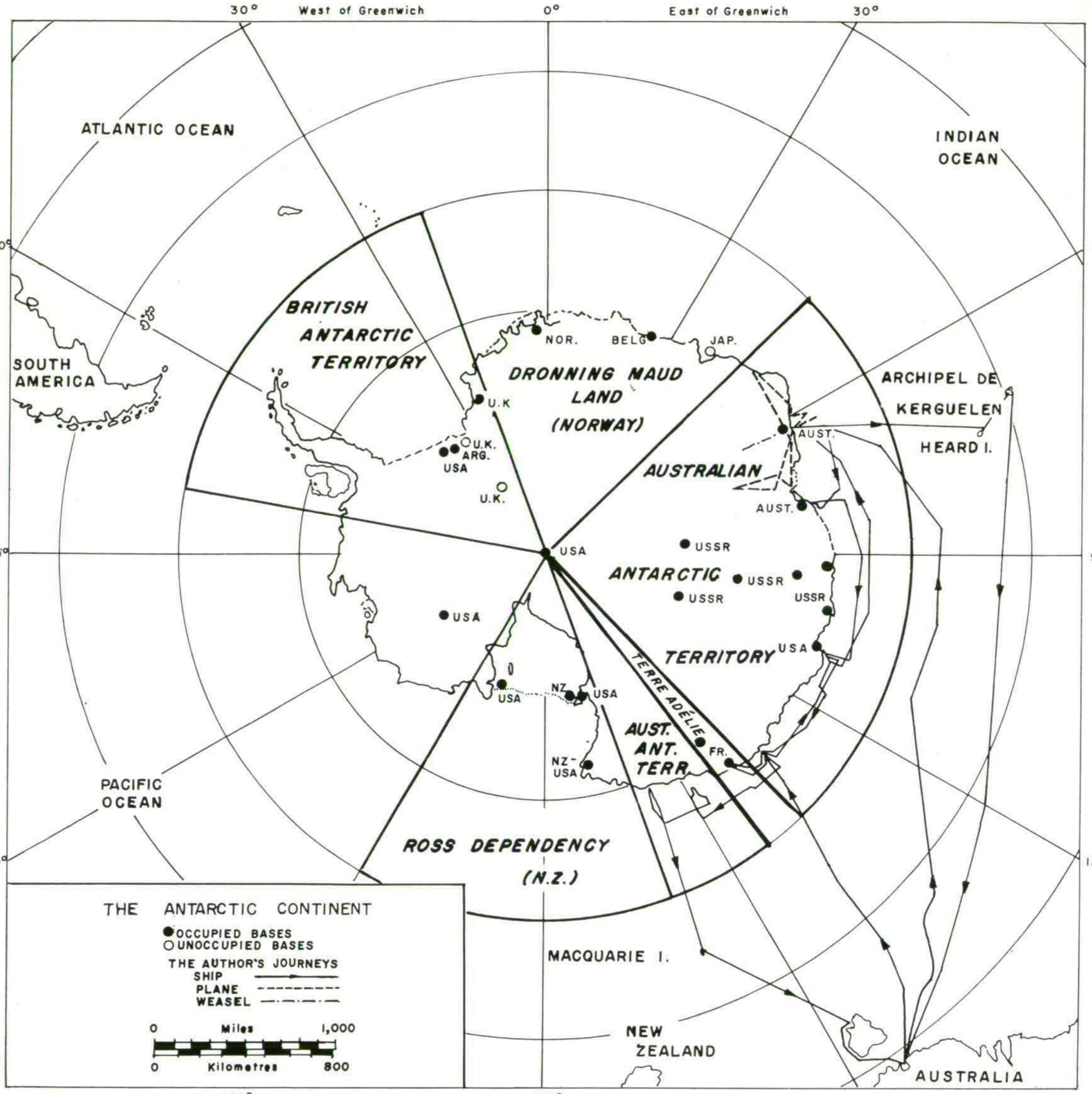
The rates of ice flow were measured at the coastal cliffs near Mawson, as well as a 25km-line 20km inland of Mawson, and several individual glaciers along the coast. From these, and other measurements available of Antarctic coastal ice flow, an estimate is made of the total annual loss to the Antarctic ice cap.

From an examination of photographs of the ice cap coastline near Mawson and other parts of Antarctica, taken at different times, it is concluded that there has

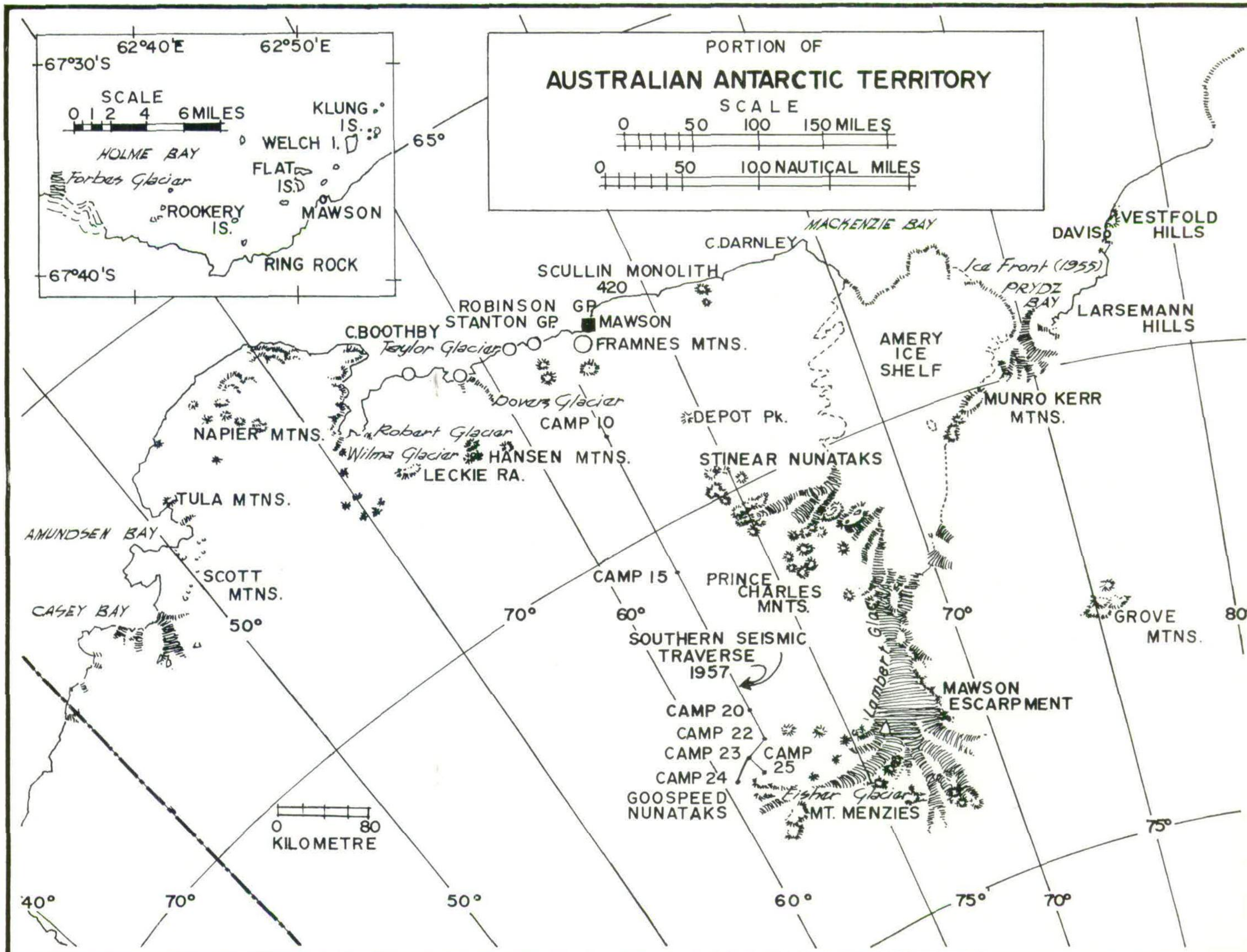
been negligible net change in the ice margins over the last 20 years, and probably over a very much longer period.

From a summary of the individual factors, a total annual net mass gain for the Mawson region is evaluated. Here a net gain over loss of at least 30 per cent is estimated. By including results for other parts of Antarctica, an estimate of total gain of $1 \cdot 6 \times 10^{18}$g yr^{-1}, is calculated, compared to a total loss of $0 \cdot 7 \times 10^{18}$g yr^{-1}. Because of the large possible errors in the present estimates, however, much more data are required before a firm conclusion can be drawn.

To study the amount of Antarctic snow removal by blizzards, a number of different snow traps were developed and a series of measurements carried out to determine the vertical profile of drift density for different occasions and windspeeds. Two different mechanisms were recognized: a saltation layer to 50cm and turbulent transfer above this. From these results the annual loss of snow is calculated.



MAP 1
The Antarctic continent.



Map 2
Portion of Australian Antarctic Territory.

O. INTRODUCTION

The investigations which have gradually revealed to us the nature and workings of the great ice cap of Antarctica were initiated by explorers of the nineteenth century who determined the broad outlines of the continent. At the beginning of the present century the first land-based expeditions enabled scientists to describe the glacial processes and even undertake some measurements.

In 1902, Drygalski measured the movement of the glacier ice at Gaussberg, and precipitation measurements were made at Snow Hill island by Nordenskjold's Swedish expedition in the same year. During Shackleton's expedition of 1907-09, David and Priestley made glaciological observations, and during Scott's last expedition, 1910-13, Wright and Priestley undertook snow and ice studies which led to the publication in 1922 of their classical book on glaciology. Mawson's Australasian Antarctic Expedition, 1911-14, made glaciological studies, and for the first time in Antarctica an attempt was made to measure the amount of snow blown off the ice cap and out to sea.

Between the two world wars there was little scientific investigation of the ice cap, apart from the work done by Byrd's expedition in 1939-41. There was, however, an important step forward when systematic aerial photography was introduced to Antarctica by Lars Christensen in 1936-37 and Ritscher in 1939.

After the Second World War, in 1947, Byrd made an intensive survey of a large part of Antarctica and the huge collection of air photographs secured during this operation gave an overall integrated picture of the ice cap and its physical features. In this post-war period the chain of bases first established in 1943-44 by the Falkland Islands Dependencies Survey was extended, but their outlying situation on the Graham Land peninsula was unsuited to studies of the main ice cap. From 1950 to 1952 the Norwegian-British-Swedish Antarctic Expedition worked in Dronning Maud Land. This expedition was of considerable significance, since it established the principle of close international co-operation and also went into the field with a scientific programme already planned in detail with the aid of air photographs. The expedition used mechanical transport with success and made the first seismic ice depth measurements on the inland ice. The well executed field work has been followed up by the publication of excellent scientific reports.

In 1950, Expéditions Polaires Françaises set up a base, Port Martin, on the coast of Terre Adélie and this base was occupied for two years before being destroyed by fire. The observations made in Terre Adélie provided the material for an important work on glaciology by Loewe (1956), as well as for a number of shorter papers. A small French base was maintained at Pointe Géologie in 1952, but in 1953 France temporarily withdrew from Antarctica.

In 1954 the Australian National Antarctic Research Expeditions, under the direction of Phillip Law, set up a permanent Antarctic base, Mawson, on the coast of Mac.Robertson Land. During the first year of occupation, measurements of accumulation and ablation were initiated by R. Dovers and firn temperatures were measured on the ice cap. In the following two years observations were extended to include measurements of ice movement, the work being carried out by P. Crohn.

The International Geophysical Year of 1957-58, a development from the Polar Years of 1882-83 and 1932-33, called for the setting up of many Antarctic bases by different nations (see Map 1) and for a detailed study of glaciology. For the first time, simultaneous measurements were to be made in many different parts of the continent and improved means of transportation would permit wide areas to be covered by each expedition.

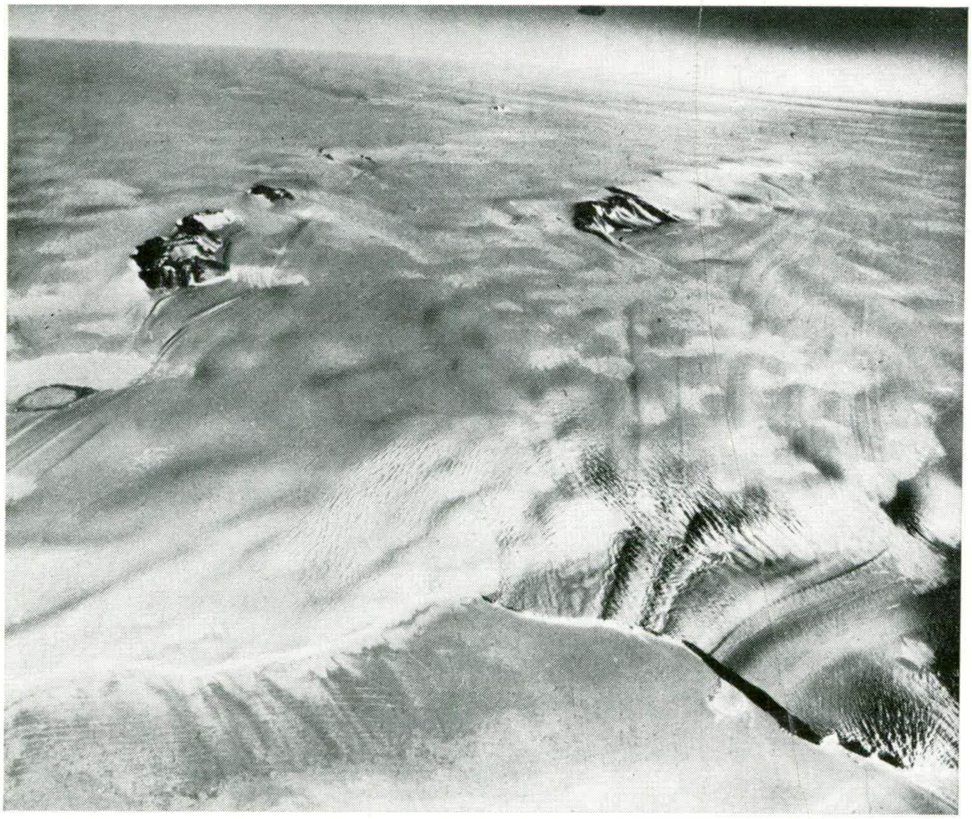
In 1956 the author joined the Australian National Research Expeditions as glaciologist and in 1957 wintered at Mawson. Equipment and facilities for glaciological work were limited, but a wide range of topics was studied. The results obtained in the field were analyzed in Melbourne in conjunction with the Meteorology Department, Melbourne University, during the austral winter of 1958, and in the summer of 1959 the author was given the opportunity of returning to Antarctica in M.V. *Magga Dan* to make further studies and to install at Wilkes station some equipment developed from experience gained in 1957.

Some results have already been published, but in the present paper the observations which lead to an outline of the mass economy of the ice cap have been brought together. After giving a general account of the nature and behaviour of the ice cap, the accumulation of snow on the surface of the ice is described and the transformation of snow into ice is touched upon. Measurements of accumulation from Mac.Robertson Land are compared with similar ones from other parts of Antarctica and values leading to an accumulation estimate for the whole ice sheet are introduced. The agencies responsible for removal of ice from the continent are then discussed and measurements of ice flow, ablation, and wind-blown snow are given, together with descriptions of the measuring techniques used. A still undetermined factor in the Antarctic mass budget, oceanic melting beneath ice shelves, is briefly discussed. Since the mass budget, towards which the observations lead, is intended to show whether or not the ice cap is in equilibrium at present, direct observations on fluctuations of ice level are given in the chapter dealing with past and contemporary variations in the positions of the ice margins. Finally, mass budgets for the region studied and for the whole of Antarctica are drawn up. The average precipitation over Antarctica is estimated and the ice import and export estimates are compared, showing a large surplus of accumulation over ice loss. This apparent indication of a growing ice cap is regarded with some doubt and it is concluded that more data are needed before this result can be accepted.

1. PHYSIOGRAPHY AND CLIMATE

The ice cap of Antarctica, consisting of a great dome of slowly-moving ice, the shape of which is discussed in a later section, rests on a rock base which, although depressed below sea level by the ice load in many areas, seems to be of a continental type.

The ice deforms in a manner intermediate between plastic flow and viscous flow and moves radially outwards under its own weight as new snow accumulates on the surface. Temperatures in the interior are very low and there is little precipitation in the heart of the continent. The bulk of the snow falling on the ice cap results from cyclonic activity in the peripheral regions, and the main accumulation zone is a circumferential belt which is probably no more than 800km in width.



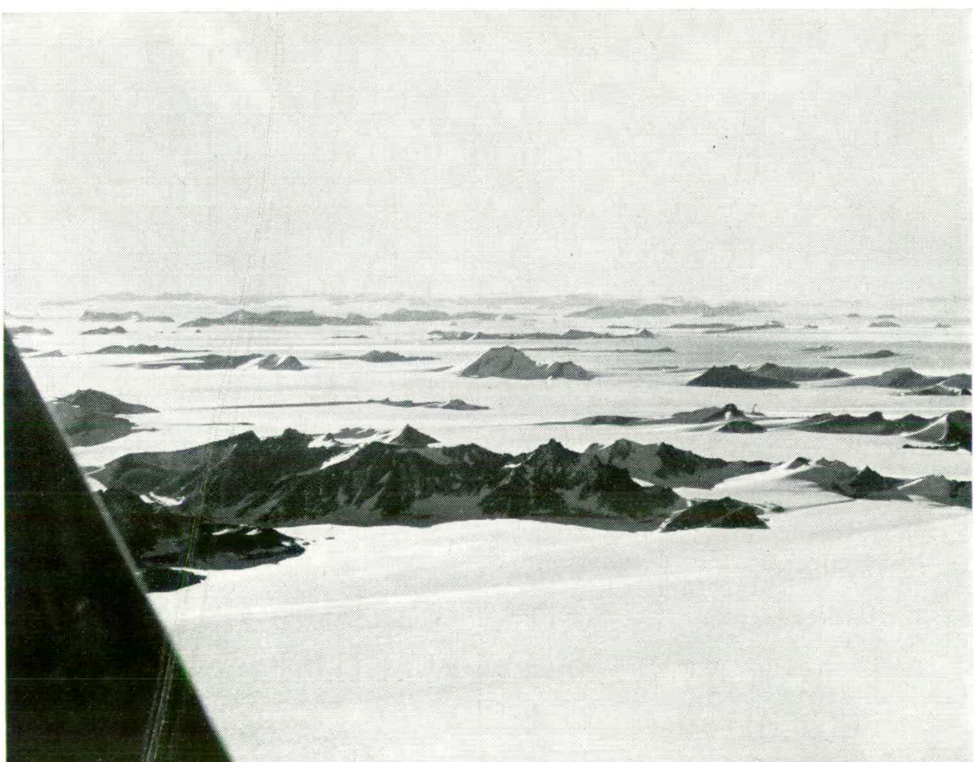
ANARE photo

PLATE 1

Coastal slopes of the ice cap east of Mawson. The humpy nature of the thin coastal ice is apparent and the general snow-free surface can be seen contrasting with the few snow patches. "Flow lines" stand out clearly on the hard ice. A melt-water lake can be seen on the left of the photograph. The mountains are Mount Rivett and Mount Marsden.

Of the snow which falls on the ice cap, a certain fraction is soon swept towards the coast by the persistent downslope winds, eventually being lost to the sea, whilst the rest is redistributed over the surface by winds before being buried and compacted by succeeding snowfalls. Pressure and migratory recrystallization transform the firn and, when the air spaces between the grains become separated and sealed as settlement increases the density, bubbly glacier ice is formed. This ice then moves towards the edges of the ice sheet under its own weight, the main deformation being concentrated in the lowest layers of the ice until the smaller depth of the fringes is reached. In the outer regions of the ice cap evaporation occurs and there may be melting followed by direct run-off to the sea. The main ice loss is effected by iceberg calving from the radially flowing ice sheet, from faster-moving ice streams, or from floating ice shelves.

In the coastal regions between longitude 45°E and longitude 80°E, the idealized picture of a radially moving, flat-based ice cap is considerably modified by the nature of the subglacial topography. Bays and valleys in the subglacial land mass produce indentations of the ice coastline and flow directions are changed, fast-flowing ice streams generally occupying coastal valleys. Ice shelves may be located in the bays, receiving inflows from the surrounding continental ice and



ANARE photo 5392B G. Johansen

PLATE 2

Jagged mountains projecting through the surface of the ice cap in Enderby Land.

surface nourishment from direct precipitation and wind-blown snow. Seismic, gravity and survey work south of Mawson shows that the subglacial terrain in Mac.Robertson Land is rugged, with some areas 500m below sea level, whilst 3,800m mountains break through the ice surface 700km inland. For a distance of several hundred kilometres from the coast the contours of the ice surface are largely determined by the underlying land features.

Along most of the coastline, continental ice meets the sea directly without intervening ice shelf and, except for floating glacier tongues, the ice is land-based. Ice shelves are generally situated in bays, and sometimes tortuous inlets between ice-covered islands are filled with shelf ice. The coasts of Enderby Land and Kemp Land have several small shelves but the largest one in the 45°E to 80°E sector is the 50,000km^2 Amery Ice Shelf, which lies at the head of the Mackenzie Bay—Prydz Bay indentation in Mac.Robertson Land. The line of the continental ice cliffs is occasionally broken by rock exposures which range in size from small rocky knolls to extensive ice free hills, such as the Vestfold Hills in Princess Elizabeth Land (Plates 3 and 4). Many sections of the coast are of the skerry type and groups of small islands lying within 15km of the mainland are common (cf. Plates 11 and 12).

ANARE photo 7177

PLATE 3

Part of the Vestfold Hills seen from the air in summer. Davis station is on the seaward edge of the main mass almost in the centre of the photograph.

ANARE photo 3073/11 Phillip Law

PLATE 4

Scullin Monolith, Mac.Robertson Land. This great coastal rock rises steeply from the sea and diverts the flow of continental ice round its sides to east and west. A girdle of spray ice can be seen running round the base of the cliffs.

New sea ice starts to form in March, but this ice is frequently broken up and dispersed by gales between March and July along most of the Mac.Robertson Land coast. The sea ice continues to grow thicker until late September at Mawson, and the maximum thickness after one winter's growth is about 150cm. Strong melting during December brings about a rapid deterioration of the ice and the final break-out generally occurs in late January. It is unusual for the ice around Mac.Robertson Land to persist unbroken through the summer, so that ice of more than one year's growth is uncommon. On some parts of the coast stable sea ice never forms at all, for new ice can be broken up persistently by katabatic winds, cyclonic gales, or ocean currents (see Mellor 1959d).

In Mac.Robertson Land the surface of the continental ice rises steeply from the sea to a height of 1,500m about 80km from the edge and an altitude of 2,500m is reached at 300km inland. Further inland, slopes are more gentle and maximum plateau heights 600km from the coast are around 3,000m. The steep marginal slopes are generally swept by strong katabatic winds which remove most of the snow below the 600m level and affect the accumulation to about the 2,000m level. The snow surfaces of windy areas are covered with sastrugi, sharp-edged ridges oriented along the main katabatic wind directions and produced during redistribution of snow by the wind (Plate 6). In Mac.Robertson Land the wind maintains in a snow-free condition a coastal belt which suffers ice loss from its hard ice surface by melting during the summer and by evaporation throughout the year (Plates 1 and 12). This net ablation zone, lying between the coast and the firn limit (which has a maximum height around 900m), is not usually more than 20km in width, although there are exceptions to this rule where valleys channel the katabatic winds. Although ice shelves retain a snow cover throughout the year, the continental ice slopes bounding the shelves generally have the appearance of ablation areas, and it is surmised that a considerable proportion of the surface nourishment of the Amery Ice Shelf comes from snow blown off the continental slopes and deposited on the flat shelf as the gravity winds lose velocity (Mellor and McKinnon 1959).

Apart from very small sedimentary areas, the exposed mountains are of old basement complex, metamorphic rocks intruded with more recent igneous material. There is generally a dramatic relief (Plate 2), and ample evidence of a former more intense glaciation exists. The flanks of the larger mountains are moraine-covered, ice-free, U-shaped valleys, and roche moutonée profiles are common and raised beaches have been tentatively identified.

There is ample evidence from aerial photographs and visual observations that coastal regions of Australian Antarctic Territory further to the east are broadly similar to those in the sector 45° to 80° East longitude. Although there are some parts where the firn limit is virtually at sea level, there are also net ablation zones at intervals to the easternmost limit of the Territory and in many areas melt lakes and great surface drainage systems indicate heavy summer ablation (Plate 13). Rock exposures near Wilkes station, Lewis Island, Dumont d'Urville station and in Oates Land give the same indications of a former more extended glaciation as those found around Mac.Robertson Land.

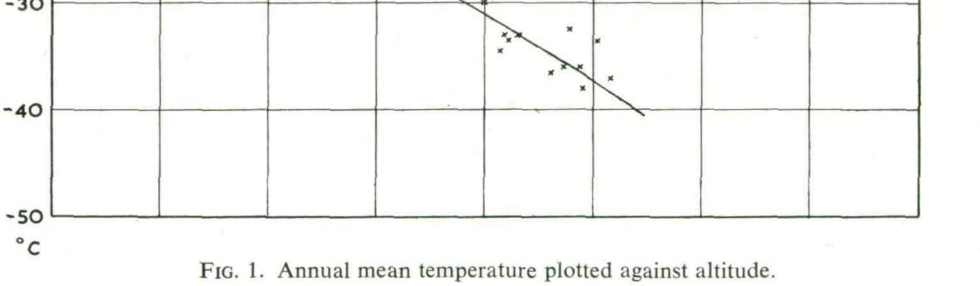


FIG. 1. Annual mean temperature plotted against altitude.

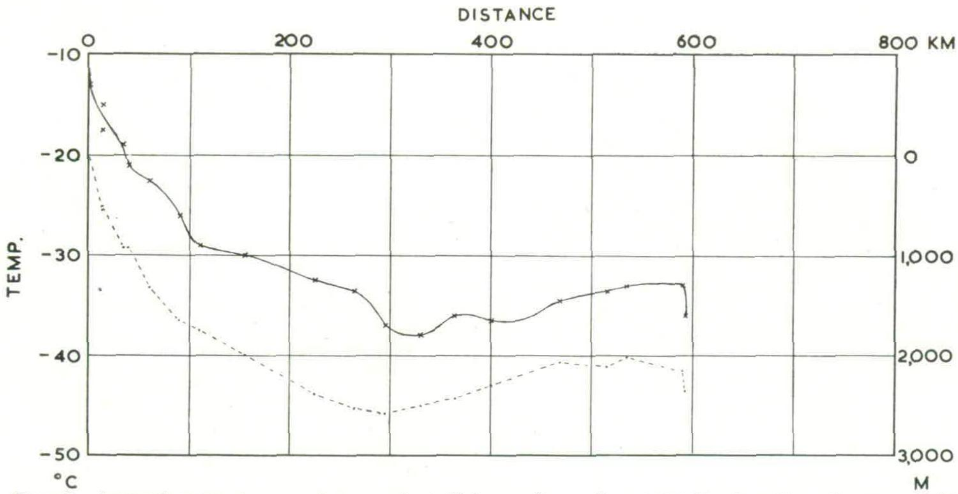


FIG. 2. Annual mean temperature against distance from the coast (broken line showns surface elevation).

Annual variations of air temperature do not penetrate deeper than about 15m below the surface of the firn and so ice temperatures measured at that depth in boreholes give approximations to present annual mean air temperatures on the ice cap. Figs. 1 and 2 show these annual mean air temperatures at the ice surface as functions of altitude and distance from the coast respectively (see Mellor 1958a). It can be seen that there is a rapid fall of temperature with increasing altitude, the rate being about 1·0°C per 100m at the 1,000m level and changing to about 1·3°C per 100m at the 2,500m level. Further examination of the graphs suggests that the effect of latitude on the temperatures is small in comparison with the height effect.

The annual mean of observed air temperatures at Mawson is rather higher than would be expected from extrapolation of the curve in Fig. 1. This is probably due to the fact that Mawson stands on an isolated area of rock which absorbs more solar radiation than the surrounding ice, thus producing a slight local warming of the air. Climatic summaries for Mawson and Davis are given in Table 1.

TABLE 1.I

Mawson 1954

Month	Mean pressure (station level)	Mean temp. (°C)	Wind velocity (m/sec)
Jan.	—	—	—
Feb.	987·5	—	6·2
Mar.	990·2	—7·7	10·3
Apr.	981·7	—17·7	6·7
May	989·4	—17·6	10·3
Jun.	991·3	—14·4	10·8
Jul.	998·7	—18·3	10·3
Aug.	989·8	—19·7	9·3
Sep.	979·2	—21·0	8·8
Oct.	980·2	—14·8	10·3
Nov.	989·5	—5·2	10·3
Dec.	993·8	+0·1	6·2

TABLE 1.II

Mawson 1955

Month	Mean pressure (station level)	Mean temp. (°C)	Wind velocity (m/sec)
Jan.	984·7	—0·3	7·7
Feb.	987·2	—7·2	9·3
Mar.	989·6	—12·3	9·8
Apr.	992·7	—14·5	10·3
May	987·9	—14·1	15·5
Jun.	984·1	—18·2	12·4
Jul.	986·9	—17·8	15·5
Aug.	1000·1	—15·4	10·8
Sep.	978·4	—16·7	11·8
Oct.	987·0	—11·3	12·9
Nov.	989·7	—3·4	13·9
Dec.	992·4	—0·3	9·8

TABLE 1.III

Mawson 1957

Month	Mean pressure (mean sea level)	Mean temp. (°C)
Jan.	—	—
Feb.	—	—
Mar.	991	—9·9
Apr.	991	—13·1
May	996	—15·3
Jun.	997	—12·8
Jul.	991	—17·7
Aug.	991	—15·3
Sep.	—	—
Oct.	984	—9·1
Nov.	995	—3·1
Dec.	996	—0·6

TABLE 1.IV

Mawson 1958

Month	Mean pressure (mean sea level)	Mean temp. (°C)
Jan.	992	—0·3
Feb.	994	—5·7
Mar.	991	—7·1
Apr.	991	—14·2
May	997	—14·7
Jun.	993	—15·8
Jul.	988	—15·7
Aug.	983	—20·0
Sep.	982	—18·3
Oct.	988	—14·1
Nov.	989	—6·3
Dec.	983	—0·8

TABLE 1.V

Davis 1957

Month	Mean pressure (mean sea level)	Mean temp. (°C)	Wind velocity (m/sec)
Jan.	—	—	—
Feb.	988·6	−3·1	4·0
Mar.	986·9	−8·5	4·6
Apr.	986·9	−9·4	4·5
May	994·4	−14·8	4·7
Jun.	996·6	−11·7	5·3
Jul.	986·7	−17·6	4·3
Aug.	988·0	−15·3	4·9
Sep.	981·6	−16·7	3·5
Oct.	982·6	−8·9	6·5
Nov.	991·7	−6·7	4·2
Dec.	994·6	−0·2	4·3

TABLE 1.VI

Davis 1958

Month	Mean pressure (mean sea level)	Mean temp. (°C)	Wind velocity (m/sec)
Jan.	992·2	−0·3	—
Feb.	—	—	—
Mar.	—	−7·6	5·2
Apr.	—	−11·9	—
May	995	−13·5	3·4
Jun.	995	−16·1	5·4
Jul.	987	−14·6	7·0
Aug.	981	−19·0	—
Sep.	981	−18·7	—
Oct.	986	−13·8	—
Nov.	985	−6·4	—
Dec.	982	0·0	—

At the time of writing, summaries of climatic factors were not then published for Mawson for 1956, 1957 and 1958, and for Davis for 1957 and 1958. The figures for those years listed in the foregoing table were based on "Monthly Climatic Data for the World" and ANARE newsletters.

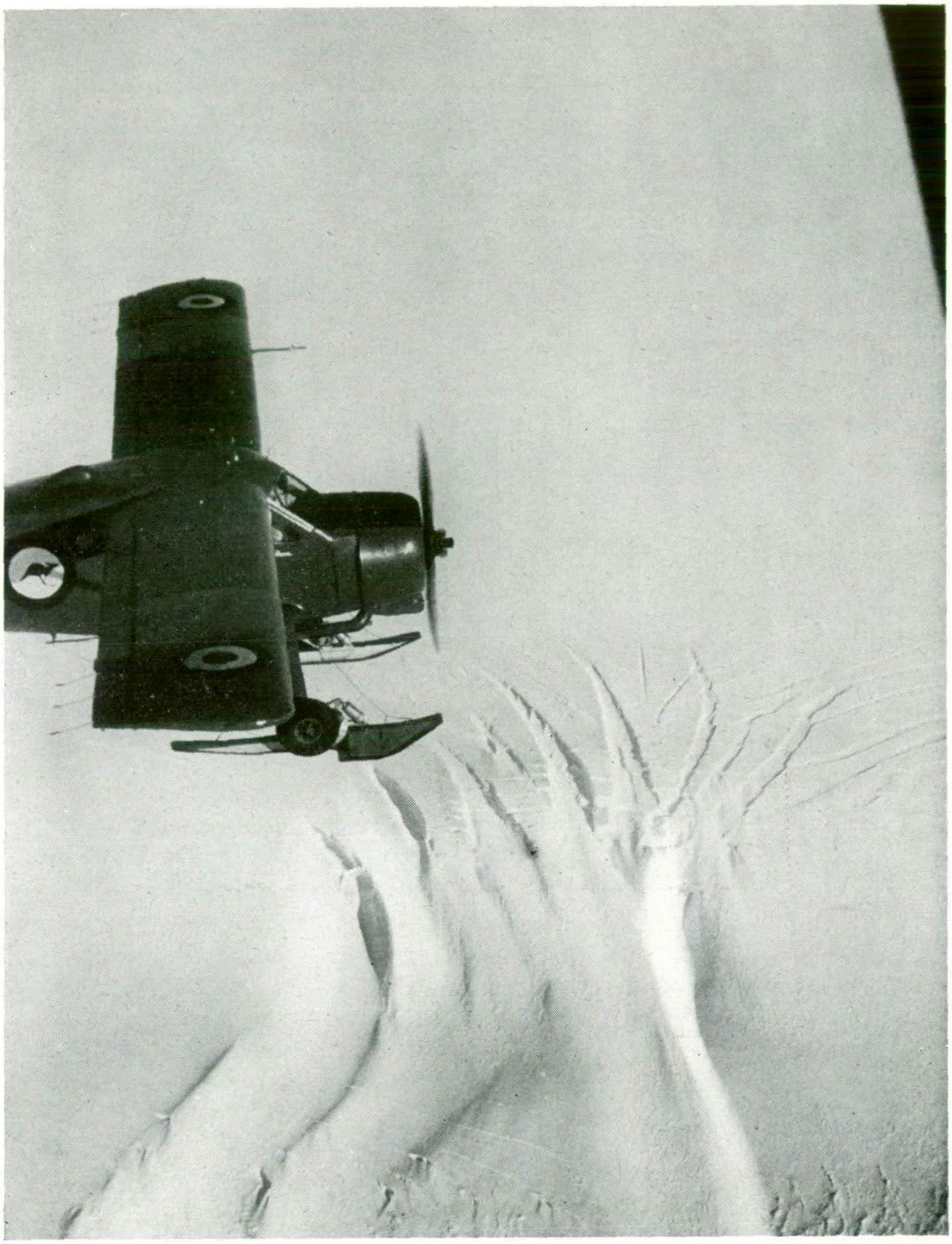
2. SNOW ACCUMULATION

2.1. ACCUMULATION MEASUREMENTS

The measurement of snow accumulation in Antarctica presents great problems. It is almost impossible to measure directly the precipitation in the windy conditions prevailing over much of the coast, and accumulation measurements, if they can be made, therefore provide a valuable source of meteorological data in addition to being necessary for the glaciological mass budget. There are two principal methods for measuring accumulation, namely stake measurements and pit sections, but there have been serious drawbacks to the application of both these methods in Antarctica in the past. In order to gauge accumulation over a tract of country directly on measured stakes placed in the snow, it is necessary to make at least two journeys along the line of stakes by surface transport. It is not always easy to reconcile this requirement with the logistics of a complex expedition. The method of measuring net accumulation from the stratigraphy revealed in the walls of pits dug in the snow was previously difficult to apply in Antarctica as the banding is neither simple nor easy to identify. However, the continuity of Australian expeditions has permitted limited stake measurements to be made and research by Schytt (1958) has developed methods which sometimes permit the identification of annual layers in pit sections, so that we are now in a position to make reasonable accumulation estimates.

It is believed that the major part of the Antarctic precipitation falls on the peripheral regions of the continent and that it results from cyclonic activity around the coasts. Far inland, air temperatures are extremely low and the air can only contain very small quantities of precipitable water. The accumulation of ice on Antarctica can therefore be estimated, if a number of radial accumulation profiles running from the coast to about 800 km inland are obtained for several parts of the continent, together with scattered point measurements for the interior. Such full information is not yet available, but when the results of the various IGY investigations are published there should be more data.

In the sloping outer regions of the ice cap, measurement of surface accumulation either by stakes or from pit sections is complicated by the irregularity of deposition brought about by strong winds. During the 3½ months which were spent on the plateau by the writer in the summer of 1957-58, precipitation was associated with two main types of weather: heavy blizzards during which strong south-easterly winds laid down big snow dunes, and unusually calm whiteout periods when star-shaped snow crystals up to 5mm diameter fell vertically. Days with precipitation were comparatively rare and under the more usual conditions strong katabatic winds



ANARE photo 7164 K. B. Mather

PLATE 5(A)
Snow-bridged crevasses from the air.

ANARE photo 7131B M. Mellor

PLATE 5(B)

Giant crevasses on an ice dome situated on the ice cap in latitude 72°50′S, longitude 60°30′E. The width of the larger ones is estimated to be about 70m. The "haycocks" ranged along the edges of the crevasses are characteristic features on the inland ice domes.

(whose velocities and directions were modified by local contours of the ice cap) eroded snow dunes and carried along previously deposited snow by turbulent suspension and saltation, forming sastrugi oriented parallel with the katabatic wind direction.

Russian meteorologists claim that stable high pressure tongues form above the major northward projections of the ice cap so that cyclone movement is blocked and zones of stationary cyclones result (Krichak 1958). A stationary cyclone zone is said to be centred off Mackenzie Bay (Map 2) and the strong winds experienced at Mawson are attributed to reinforcement of cyclonic winds by gravity winds. Precipitation is believed to be connected mainly with the frequent stationary cyclones. It is hoped that the records of dune and sastrugi orientations compiled by ANARE personnel will be examined for further evidence of precipitation origins.

In 1954, R. Dovers placed a line of measured stakes along a weasel trail from Mawson to the Prince Charles Mountains and during the two following summers of 1955 and 1956 the exposed lengths of the stakes were re-measured by P. Crohn. Crohn grouped the stakes along a number of sections of the trail and used the mean decrease of stake height to give the accumulation over the area covered by each section of trail (Crohn 1958). A large number of field measurements made by the

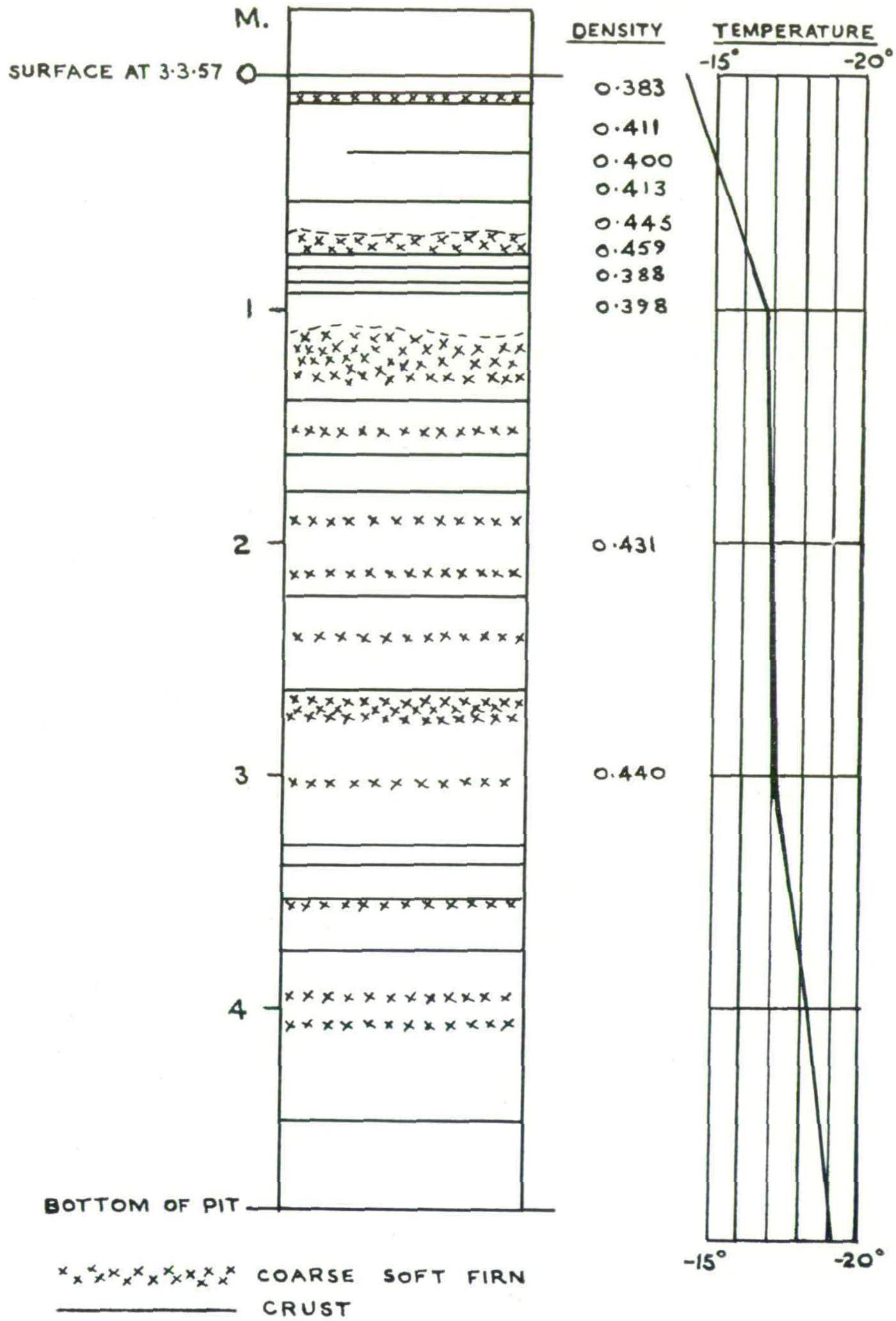


FIG. 3. Firn profile at 900m altitude.

writer suggest that the mean surface firn density adopted by Crohn for his conversion of snow accumulation to equivalent water was too high by about 12% (see also Swithinbank 1959) and if the snow accumulation, averaged over two years, is converted to equivalent water by using a mean density value of $0{\cdot}40 g/cm^3$, instead of Crohn's $0{\cdot}45 g/cm^3$, the following figures are obtained.

TABLE 2.I

Region (km south of Mawson)	Annual net Accumulation (cm of water)
26 to 80	5
80 to 192	10
192 to 256	17
256 to 320	20

In early March 1957 a 5m pit was dug on the plateau 55km inland from Mawson and the stratigraphic profile was studied. A tentative identification of annual layers was made (see Fig. 3) and a 10-year mean accumulation value of 13cm of water resulted. This was very high in comparison with the average stake

ANARE photo 7183 K. B. Mather

PLATE 6

The hard sharp-edged sastrugi of the ice cap are oriented along the direction of the prevailing wind. The irregular surfaces lead to complicated stratigraphy in the buried firn. The tractor in the photograph is towing sledges which carry fuel and equipment, a power drill and a lightweight living caravan. Hard sastrugi punish vehicles and sledges severely.

value of 5cm for the area. However, it is known that considerable variations can occur over a small area and, when the site was re-occupied at the end of July, there had been an accumulation of 7·7cm of water in 4½ months.

During a 6-week period in December 1957 and January 1958, about 4cm of water accumulated between 200 and 400km inland. On the return journey to Mawson in February 1958, tractor and sledge tracks made in November 1957 were intermittently visible, although for much of their length they were buried by snow up to 30cm deep. (Swithinbank (1959) states that summer accumulation on

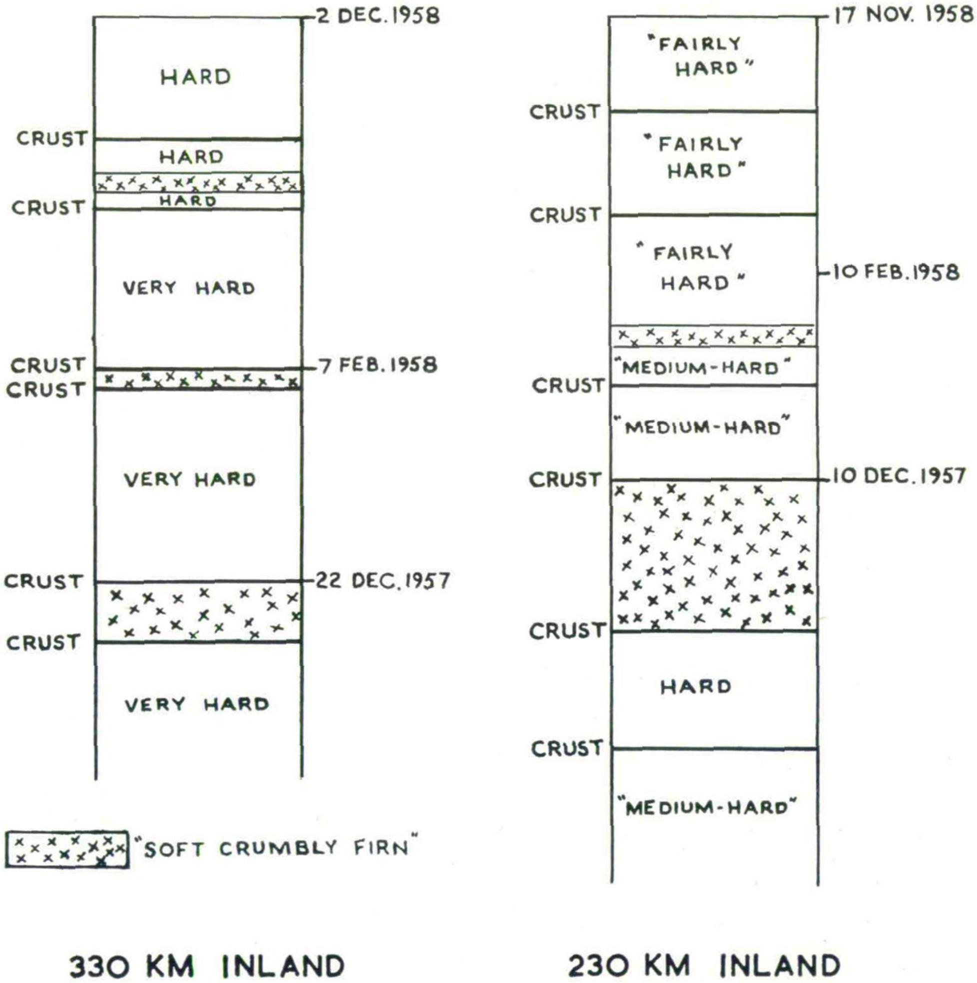


FIG. 4. Development of firn layers.

the plateau slopes of Dronning Maud Land represents 20-40% of the annual accumulation.) There was ample evidence of wind transport during the 100-day southern journey: deep drifts built up around the tractor trains night after night in cloudless weather and, on other occasions, knee-depth deposits of soft snow which fell in calm whiteout periods were quickly dispersed and packed when the wind regained its usual strength. On the highest, relatively wind-free part of the traverse it is possible that accumulation could be as high as 35cm of water, but south of this area the traverse skirted a large bowl-like depression which falls away to the Lambert Glacier, and accumulation in this region 500km inland may be limited by katabatic winds stronger than normal for this distance from the coast.

During the 650km southern journey of 1957-58, a large number of long shallow trenches were bulldozed to a depth of about 1m. Several other trenches were taken to greater depths, two pits, 5·3m and 6·1m deep, were dug, and the accumulation estimates given below were made from these pits and trenches by the writer. Groups of measured stakes and some dye patches were placed at intervals along the 1957-58 trail by the writer and some of these were checked during the following summer by a seismic party under the leadership of I. Adams, O.I.C. Mawson, 1958. Unfortunately, the number of stake groups re-measured was insufficient to provide a check on the 1957 interpretation of firn profiles.

TABLE 2.II

Distance south of coast (km)	Estimated accumulation from pit sections (cm of water)	10 months measured accumulation (Feb. to Dec.) (cm of water)
60	8	—
130	12	—
160 (ice dome belt)	2	—
230	12	11·5 (21·5 from 10-12-57 to 17-11-58)
275	9	0
300	8	15
330	11 or 39	16·5 (26 from 22-12-57 to 2-12-58)
350	11	0
365	15	9
530	14	—
650	9 to 15	—

However, pit sections plotted at points 230 and 330km from the coast by Blake and Jesson of the 1958-59 seismic team give interesting information concerning seasonal effects on the firn layers. At each of these two points the author placed 3 stakes and covered the snow surface with dye powder in December 1957. In February 1958 the stakes were re-measured and a small trench was cut down to the December surface. At both places new snow had accumulated during the summer, but whilst the dye at the 330km point showed clearly in the trench, that at the 230km point had evidently been eroded away before the new snow was laid down. The pit sections plotted in Fig. 4 show clearly that snow distributed over the surface during periods of strong radiation and relatively high temperatures acquires a characteristic texture, with coarse grains and low cohesive strength. Winter snows and deep undisturbed deposits are finer-grained and have a greater resistance to penetration. In recent snow studies, soft layers of coarse-grained firn have generally been regarded as being annual features from which the strata can be dated, but

Fig. 4 shows that two such layers were formed in the summer of 1957-58, the two being separated by deep, and presumably undisturbed, deposits. Thus the dating shown in Fig. 3, and in other profiles, becomes suspect and the complexity of Antarctic snow stratigraphy is underlined.

Table 2.II gives accumulation estimates from the 1957-58 pits and also 10 months accumulation as measured on stakes. In two cases accumulation has been measured over almost a complete year.

The two widely-spaced alternative interpretations at the 330km point resulted from the author's use of two different dating criteria. The low figure comes from a layering established on the basis of crusts and coarse firn layers, whilst the high value is derived from periodic density variations with depth. This pit was dug in what is believed to be a relatively wind-free area at the highest point of the traverse.

2.2. SNOW METAMORPHISM

In order to obtain information on the transformation of firn to ice, several investigations were made in 1957. Variations with depth of grain size, density and hardness were measured in pits by conventional field methods and attempts were made to extend the observations to greater depths by core drilling. The coring barrels supplied with the diesel-hydraulic, flight-auger type drill failed to work in the field, as cuttings jammed at the coring head, thus making penetration impossible. The cutting heads were modified and a completely new coring barrel was made by the author, but the drill was still incapable of pulling cores.

Since the deeper ice could not thus be brought to the surface for density determination, an attempt was made to measure its density in situ. Holes could be drilled to a depth of 38m, and so experiments in the use of radio-isotopes for density determination were conducted. Parallel bore holes were drilled about 2m apart and gamma radiation from a collimated 15 millicurie cobalt—60 source in one hole was transmitted through the firn to a Geiger-Mueller tube at the same depth in the other hole. The count rate was measured on a scaling unit at the surface, using a small diesel generator and a rotary converter as a 240-volt A.C. power source. It was not difficult to drill parallel holes, as the heavy flight-auger was sufficiently flexible to hang vertically after initial plumbing of the drill guides. The radiation counting equipment was assembled from pieces of laboratory apparatus and was unsuited for use in the severe conditions prevailing on the inland ice of Antarctica. The polythene insulation of the co-axial detector cable became brittle and cracked in the low temperatures in spite of careful handling, and several faults, difficult to remedy in the field, developed in the scaling unit. The mounting of the gamma source was such that it was rotationally asymmetrical and the orientation could not be determined after it was lowered into bores on a steel cable.

By repeated measurement it was possible to get an approximate value of $0{\cdot}7 g/cm^3$ for the density at a depth of 28m but detailed density variations with depth could not be obtained.

In 1958, a further attempt to develop radio-isotope bore-logging was made, this time using only a single bore hole. A Harwell gamma bore-logger, type 1417A, was loaned by the Bureau of Mineral Resources, Geology and Geophysics, and, with the assistance of Mr D. Urquhart of the Bureau of Mineral Resources (geophysical laboratory) the equipment was modified to measure ice density by counting back-scattered radiation. A cylindrical lead shield 10·2cm long was fixed to the lower end of the Geiger probe and a 12 millicurie cobalt-60 source was attached to the base of the lead shield. The probe was thus shielded from direct radiation but able to measure back-scatter from the ice surrounding it. Tests of the device were made in Melbourne using drums of water and of kerosene as the media of different density.

The logger was taken to Antarctica by the writer early in 1959 and field tests were made at the Wilkes satellite station (S2), 80km inland from Wilkes Base. The basic equipment was well designed and robust, giving no trouble at low temperature, but again the sensitivity of the apparatus was inadequate for tracing density variations in detail. Tests were made in 6m surface bores and in a previously cored bore hole between 40 and 60m beneath the surface. Density determinations made by the gamma logger had errors up to 17%. The author's experiments in gamma logging have been largely empirical, but a more fundamental consideration of the geometry of back-scattering might lead to a useful probe design. A gamma bore logger could

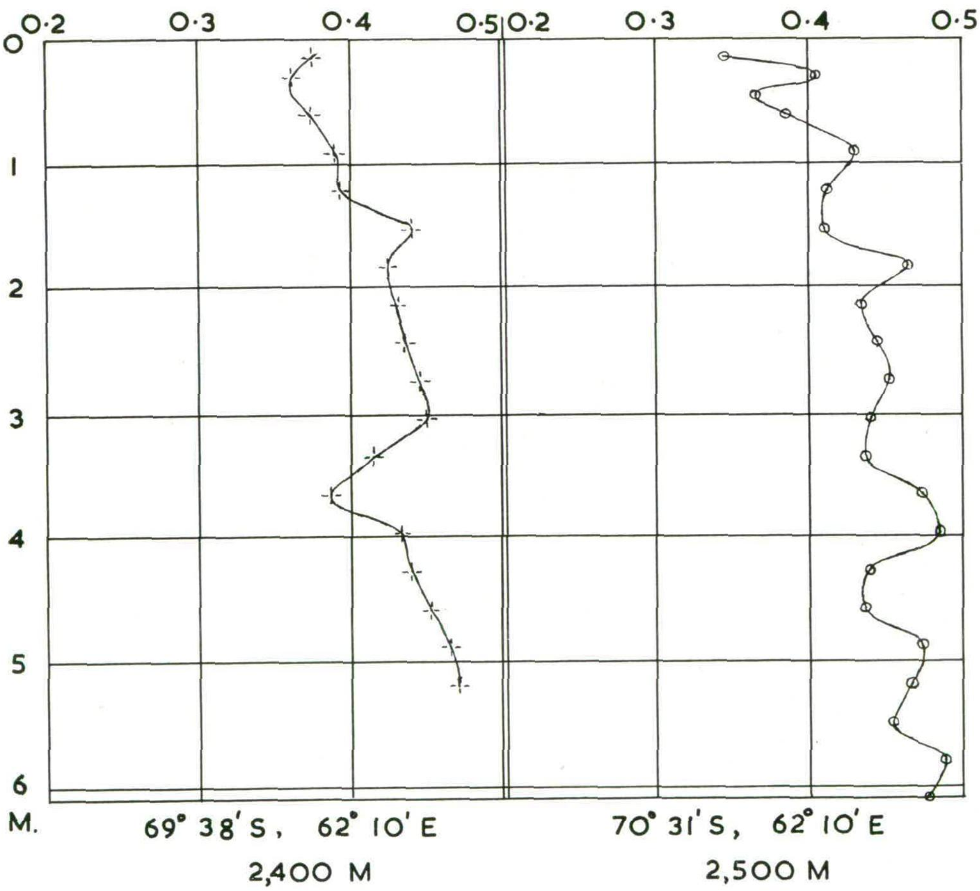


FIG. 5. Firn density in the upper layers.

be a useful tool for recording density changes in deep bores, particularly those produced by thermal drilling.

For investigation of the upper firn layers it is felt that it would be better to use the 3-inch ice-coring auger designed by the U.S. Corps of Engineers' Snow Ice and Permafrost Research Establishment for mobile field investigations (*Journal of Glaciology*, March 1957). This easily portable coring drill works very efficiently to a depth of 10m, it is capable of reaching a depth of about 30m, and a neat density kit for use with the drill is available.

Fig. 5 shows two density profiles for points on the ice cap in Mac.Robertson Land, and in Fig. 6 additional firn data, including grain size, hardness and temperature, are given for one of the points (Camp 15). It will be seen that there is no progressive grain growth below 1m, although crystal sizes from 2 to 7mm diameter (several times larger than the grains in Fig. 6) were measured by the author in thin sections of ice cut from ice cliffs and crevasses in the coastal ablation zone. It has

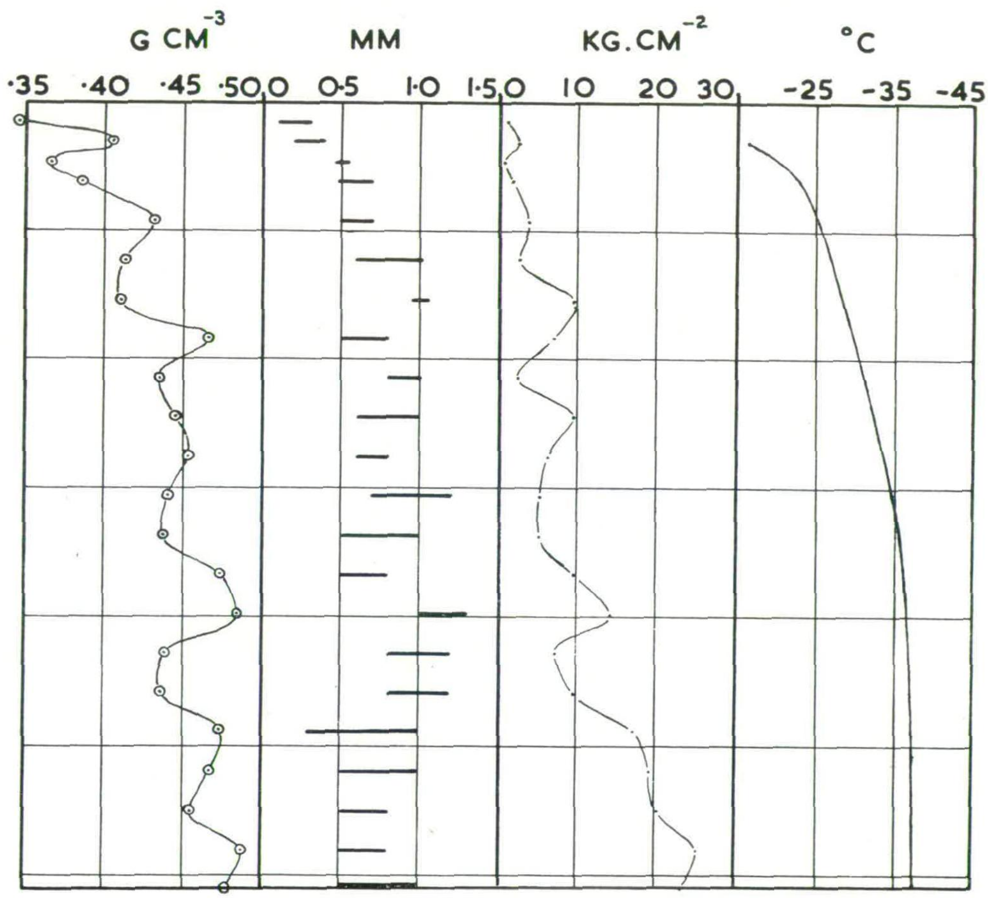


FIG. 6. Firn profile at 70°31′S, 62°10′E (2500m).

been shown by Schytt (1958) and by Stephenson and Lister (1959) that, although there is no grain growth, crystal size increases with depth, the depth—crystal area relationship appearing to be approximately linear (see Fig. 7).

In Fig. 8(A) the smoothed depth—density curve for the inland ice of Mac.-Robertson Land is compared with corresponding curves for other parts of Antarctica, Greenland, Canada and Spitsbergen, and in Fig. 8(B) the densities at greater depths (measured by workers of other nations) are shown. The Mac.-Robertson Land curve is based on measurements at latitudes 69°37′S and 70°31′S where the surface altitudes are 2,400m and 2,500m respectively. A linear extrapolation has been made, using the density measured at 28m depth by the gamma radiation method. The results from Eismitte (3,030m), Byrd Station (1,513m) and Southice (1,350m) are in good agreement with those from Mac.Robertson

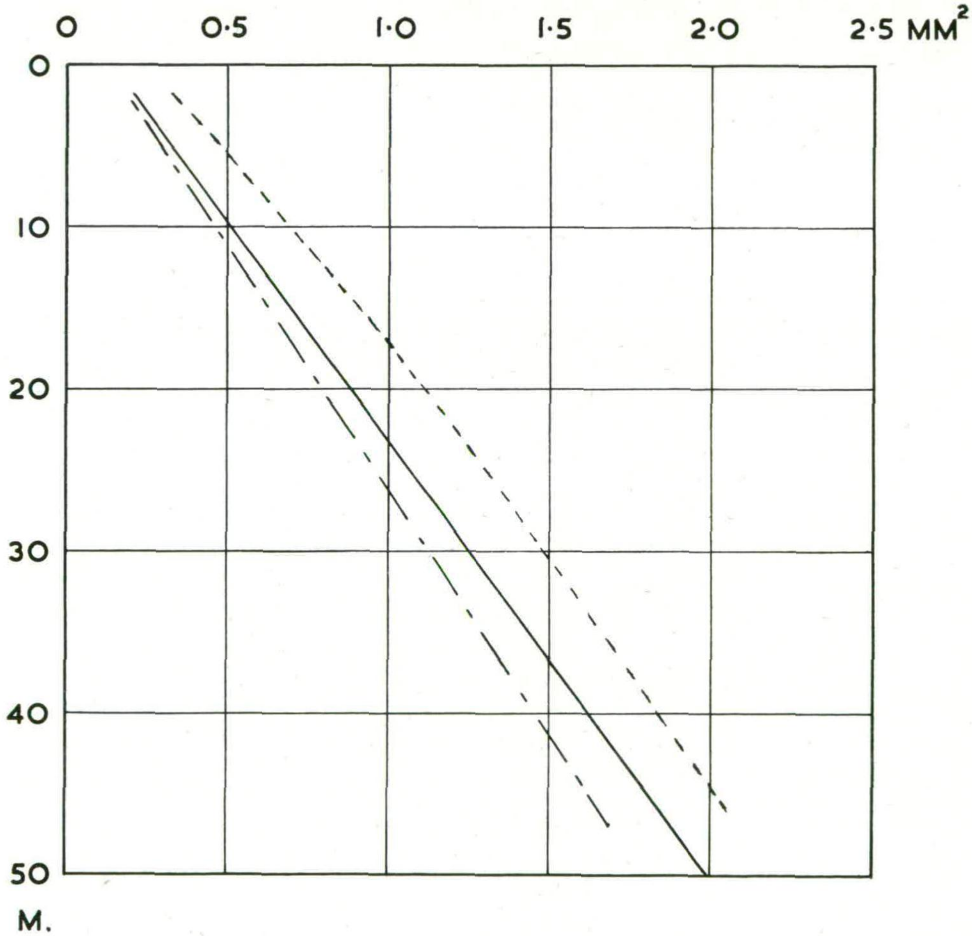


FIG. 7. Crystal size as a function of depth (after Schytt, Stephenson & Lister).
———— Maudheim (Schytt 1958)
— — — hard layers } Southice (Stephenson & Lister 1959)
— — — coarse layers

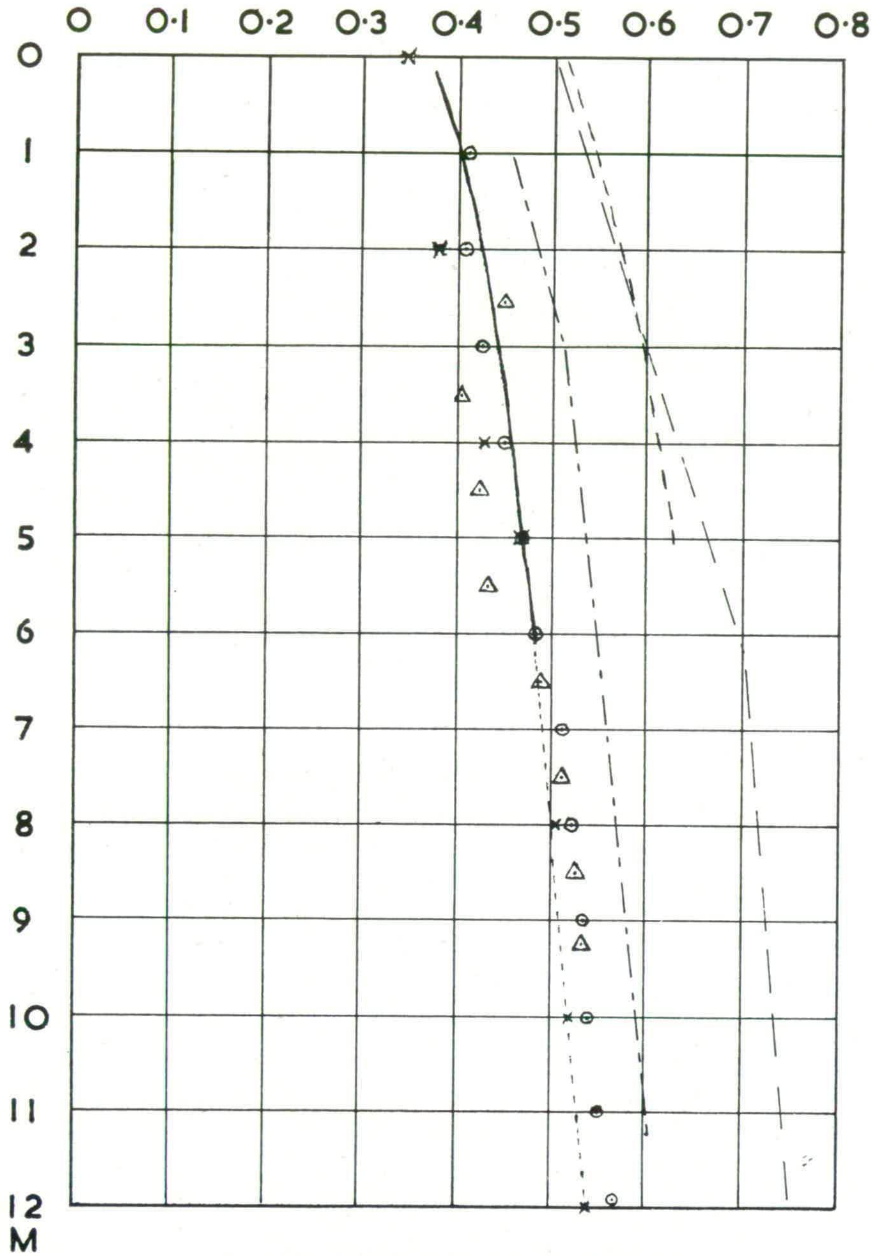


FIG. 8(A). Firn density in the upper layers.

Mac.Robertson Land

,, ,, linear exploration using 28-M density from γ-radiation tests.

Maudheim Ice Shelf (Schytt 1958)

Seward Glacier, Canada (Sharp 1951)

Isachsen Plateau, Spitsbergen (Mellor 1957)

△ Byrd station (Bender 1958)*

⊙ Southice (Stephenson & Lister 1959)

x Eismitte, Greenland (Sorge 1935)

* Data taken from IGY Supplement of Transactions of American Geophysical Union.

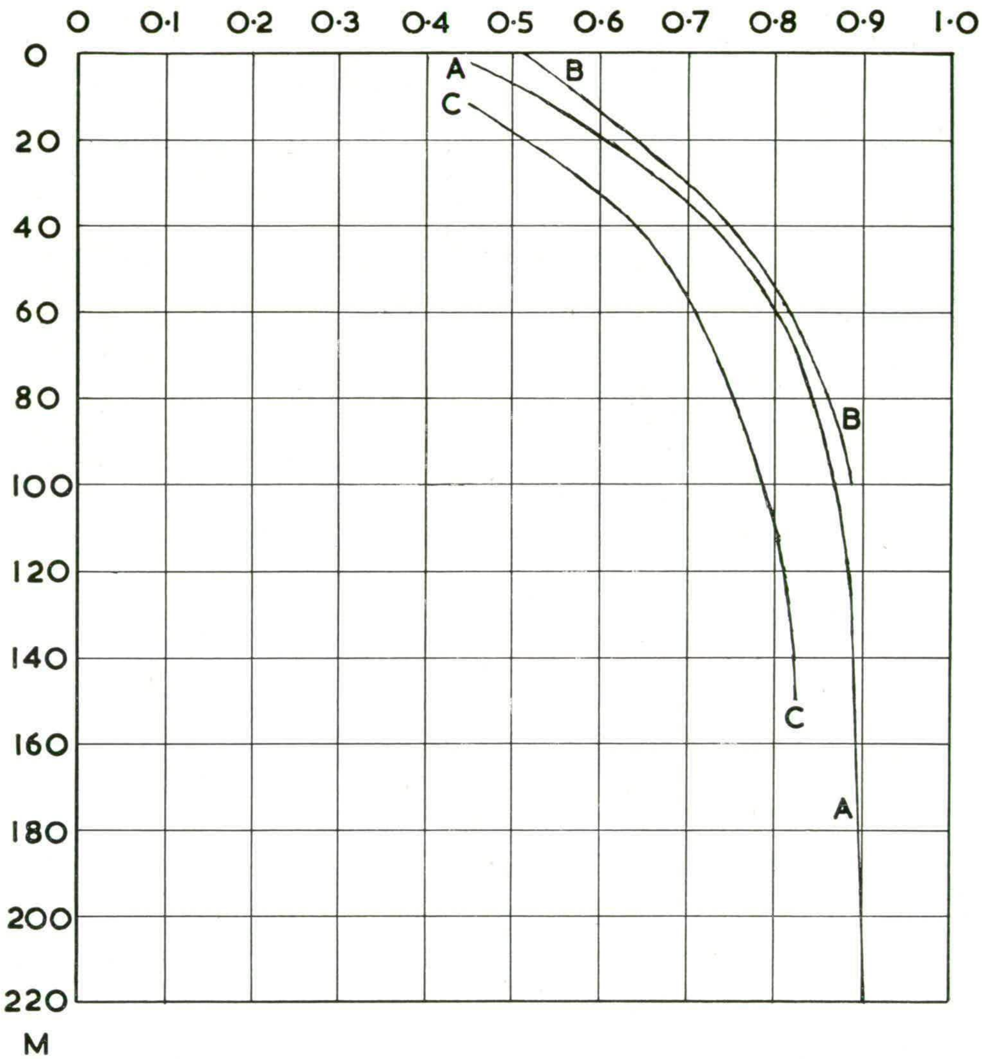


FIG. 8(B). Depth-density measurements from deep bores.
A-A Byrd station
B-B Maudheim
C-C Station Centrale, Greenland.

Land, but densities in the ice shelf at Maudheim are somewhat higher than those at corresponding depths in the inland ice. In Fig. 8(B), where densities at greater depths are shown, the asymptotic approach to the maximum density of normal hexagonal ice, 0·917g/cm^3, can be seen. It is interesting to note that densities of 0·874 and 0·891lg/cm^3, measured by the author at the top and bottom of a 60m ice cliff near Mawson, correspond to the densities at 100 and 160m depth in the Byrd Station borehole.

The rate of increase of density at depth appears to be similar for both ice shelf and inland ice, but the absolute density at a given depth depends on the initial

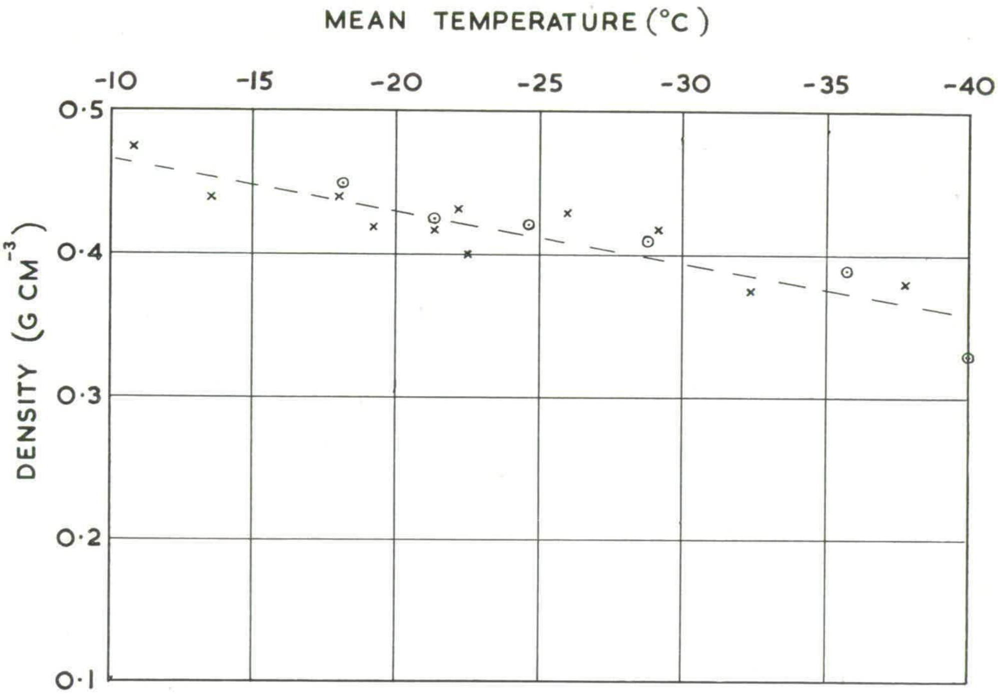


FIG. 9. Surface density plotted against annual mean temperature.
x Mac.Robertson Land
⊙ Dronning Maud Land (Schytt 1958)

surface density. It has been pointed out by Schytt (1958) that surface firn density varies with temperature and wind speed, and Schytt gave a plot of "density of last year's firn" against annual mean temperature. In Fig. 9 a similar plot has been made and Schytt's results are shown alongside those of the present author. It can be seen that there is quite a definite decrease of firn density as the annual mean temperature becomes lower. Wind speeds on the plateau can only be inferred from sastrugi measurements, and so the density values have not been analyzed for wind effects.

The density of hard, bubbly ice taken from crevasses and ice cliffs in the ablation zone was measured and it was established that density increased from 0·874g/cm^3 to 0·891g/cm^3 through a depth of 60m. There should be a simple relation between density and air pressure in the bubbles of the ice, and it was thought that it might be possible to calculate the pressure in the bubbles when they were originally sealed, and hence find the height at which the ice was formed. In trying to follow the work of Bader (1950) an unsuccessful attempt was made to improvise an apparatus for measuring pressures developed in the air bubbles of deep ice, but when a Beaver aircraft flew from Mawson to Mirny, the USSR Antarctic base, in 1957, an elegant apparatus for determining bubble pressures (Shumskiy 1955) was presented to the writer by Professor P. A. Shumskiy, the Russian chief glaciologist. Unfortunately, by the time some minor pressure leaks in the apparatus had been remedied, the writer had to leave Mawson on the major field journey and no usable results were obtained.

2.3. ACCUMULATION ESTIMATES

By interpolation and extrapolation of the accumulation measurements in Mac.Robertson Land, an estimate of the net accumulation in a 1km-wide strip running from the coast to a distance of 850km inland has been made. The estimated amount is $1 \cdot 0 \times 10^{14}$g/km yr. It will be interesting to see how this compares with the accumulation rate in other parts of Antarctica.

A great deal of information was graciously supplied privately to the writer by Professor P. A. Shumskiy of the USSR Antarctic Expedition, and this included an accumulation profile for the ice sheet to the south of Mirny. In that region the deepest accumulations, up to 85cm of water, apparently lie within 60km of the coast; there is between 20 and 30cm of water from 70 to 450km inland, and 850km inland the accumulation is 8cm. This is very different from Mac.Robertson Land, where there is virtually no net accumulation for the first 25km inland, only about 8cm of water 60km inland, and the average quantity from 100 to 600km would not exceed 15cm. An estimate made from Shumskiy's profile gives a value of $1 \cdot 8 \times 10^{14}$ g/km yr for the accumulation to 850km south of Mirny.

There were no accumulation measurements south of Davis, but it is believed that precipitation gauge readings are fairly reliable at this station, which has a mean wind velocity of only 4·7m/sec. Stake measurements on the sea ice during the winter checked with gauge results and the annual precipitation was measured as 6·5cm of water. This is surprisingly low; unfortunately, 1958 figures were not available for comparison at the time of writing (precipitation quantities transmitted in the telegraphic code have proved unreliable).

Schytt (1958) has given mean accumulation rates for the ice shelf, the continental ice slopes, and the margin of the high plateau in Dronning Maud Land. He gives the following values:

Ice shelf	36·5cm	of water
Continental slopes	26	,,
Border of the high plateau	12	,,

By taking suitable distances to represent each condition from a map of the area and extrapolating the results to 850km inland, an accumulation estimate can be derived. This gives an accumulation of about $1 \cdot 5 \times 10^{14}$g/km yr in Dronning Maud Land.

Lister (1959) gives the accumulation on the coasts of the Weddell and Ross Seas as 16cm of water and states that there is a mean of 8cm across the plateau on the route taken by the Commonwealth Trans-Antarctic Expedition. He estimates a mean accumulation of 11cm of water over the whole of Antarctica.

No further meridional accumulation profiles are available at the time of writing, but the magnitudes of values given above are confirmed by certain results from other areas. Loewe (1957) has given the accumulation in Terre Adélie as 20 to 30cm of water between 20 and 50km inland, which is in agreement with Schytt's measurements on the plateau slopes of Dronning Maud Land. Charcot Station, 300km inland in Terre Adélie, has an accumulation of 11cm (Loewe 1960). From banding in the ice cliffs (Plate 7) in Davis Bay, Wilkes Land, the writer estimates an ice shelf accumulation of 35 to 40cm (although coastal downslopes in the same area may have net ablation). Accumulation rates inland from Wilkes base have not yet been published, but there appears to be annual deposition in excess of 20cm of water. Vickers (1958) gives 20cm of water as the accumulation at Little America on the Ross Ice Shelf and suggests 17cm on the Victoria Land plateau. From a popular article by Siple (1958) there seems to be 5 to 6cm of water for 10 months accumulation from February to November at the Pole.

ANARE photo 8160 A. Campbell-Drury

PLATE 7

Well-defined annual layers in the ice cliffs near Lewis Island, Wilkes Land.

3. ICE FLOW

3.1. SHAPE OF THE ICE CAP

If the shape of an ice cap is known, it is possible to get some idea of the flow mechanism by which deformation of the ice occurs. Expressions for ice velocity can be derived and then used to supplement the few measurements of flow rates made in the field. Also, for many problems in glaciology and meteorology, some knowledge of the shape of the Antarctic ice cap is necessary and it is desirable to have a mathematical expression for the surface form of the ice in order to facilitate analyses.

In 1957 a party, of which the author was a member, carried out microbarometer step-heighting to define the profile of the ice cap for a distance of 650km south of Mawson in Mac.Robertson Land. After the first 300km from the coast the profile was affected by local topographic influences (the Prince Charles Mountains and the Lambert Glacier depression) so that the southern half of the profile cannot be used for comparison with idealized profiles. The heighting observations were reduced by Goodspeed (1958) and results have now been incorporated in recent Australian maps of Mac.Robertson Land. There seems reason to suspect, however, that the heights given may be incorrect since the observations were corrected, using air temperatures measured at the surface, i.e., below the temperature inversion. The writer feels that ice cap profiles could be obtained easily and reasonably accurately by use of the radar altimeter in aircraft. Air temperatures can be measured above the inversion, and pressure corrections can be made from meteorological charts. It is also possible for aircraft to record a large number of profiles in a short period.

Some of the most useful heighting has been carried out by Russian expeditions between the main base, Mirny, and the Soviet inland stations. Of particular value are the altitudes measured in the centre of the ice cap; the Pole of Inaccessibility, 82°06′S, 55°00′E, is at a height of 3,710m and Sovietskaya, 78°24′S, 87°35′E, is at 3,700m. It is believed that geodetic levelling has been carried out in the marginal areas of the ice cap and that aircraft on inland flights have made radar altimeter measurements. Radio-sonde data from the inland stations is used for correcting barometric heighting. Treshnikov (1958) gives further data on altitudes in East Antarctica.

In a classical application of plasticity theory to the flow of glaciers and ice sheets, Nye (1951) derived from physical considerations an equation which appears to represent the profile of the Greenland ice cap. Nye considered the forces acting on a vertical prism-shaped element of a circular flat-based ice cap moving by plastic deformation and, by integrating the equilibrium equation for the radial direction, he obtained the parabola $h = \sqrt{2h_o(R - r)}$, where h is the height of the surface at radius r, R is the radius of the ice cap and h_o is a constant depending on the density of the ice and on the bed shear stress (assumed constant).

If Antarctica is assumed to have a similar semi-parabolic profile and a constant bed shear stress, the value of h_o (or τ, the bed shear stress) fitting the Mac.Robertson Land observations leads to unrealistic altitudes for the central regions—some 50% higher than those measured by Russian inland parties. This is hardly surprising when it is remembered that there is very little accumulation in the centre of Antarctica; the central ice must move very much slower than that of the peripheral accumulation zone and smaller bed shear stresses must be expected beneath the central ice. Shumskiy (1958a) states that the ideal form of an ice shield is elliptic, but non-uniform accumulation produces regular deflections from this ideal shape. He considers that, ignoring irregularities due to subglacial relief, East Antarctica is probably a very flat irregular ellipse given by a relation of the form

$$\left(\frac{x}{a}\right)^2 + \left(\frac{y}{b}\right)^n = 1$$

where $n > 2$.

Following the Russian heighting work, Kapitza (1958a) has discussed the shape of ice caps. He initially proposes three possible flow mechanisms, viscous flow, plastic flow, and visco-plastic flow, and from these he arrives at expressions giving the height of the ice cap at any point. The three equations are:

(a) Viscous flow:

$$h = H \sqrt[4]{1 - \left(\frac{r}{R}\right)^2}$$

(b) Plastic flow:

$$h = H \sqrt{1 - \left(\frac{r}{R}\right)}$$

(This is essentially Nye's expression.)

(c) Visco-plastic flow:

$$h = H \sqrt{1 - \left(\frac{r}{R}\right)^2}$$

He compares each of the profiles corresponding to these equations with the observed shapes of Drygalski Island, Bowman Island and East Antarctica and concludes that the assumption of visco-plastic flow gives the best fit with observation.

In Fig. 10 the Australian results have been plotted against the three profiles given by Kapitza and it can be seen that they lie closer to the elliptic visco-plastic arc than to the other two profiles. It therefore seems that the simplest expression giving a reasonable approximation to the shape of the ice cap is of the form

$$h = H \sqrt{1 - \left(\frac{r}{R}\right)^2}$$

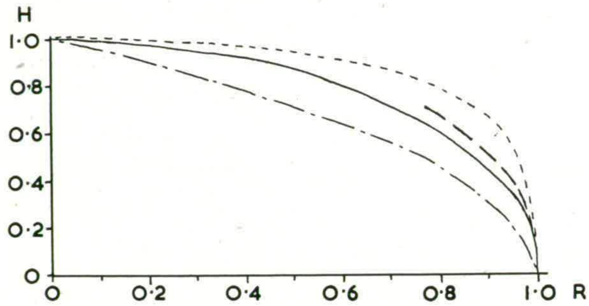


FIG. 10. Computed and observed ice cap profiles (after Kapitza).
— — — — — viscous flow
— — — — plastic flow
———— visco-plastic flow
— — Mac.Robertson Land

For Mac.Robertson Land it is reasonable to assume that $H = 3{,}700$m and $R = 1{,}400$km, so that the expression becomes

$$h = 3{\cdot}7\sqrt{1 - \left(\frac{r}{1{,}400}\right)^2}$$

The support for an assumption of visco-plastic flow is not altogether surprising when the results of creep tests on polycrystalline ice are considered. From laboratory tests Glen (1955) and Steinemann (1954) both derived a flow law of the form

$$\dot{\epsilon} = k\sigma^n,$$

where $\dot{\epsilon}$ is the strain rate, σ the stress, and k and n are constants. If ice behaved as a viscous fluid, the value of the exponent n would be unity; if, on the other hand, ice was perfectly plastic, n would be infinite (Nye 1957). The experiments of Glen and Steinemann on artifically prepared ice gave a value of about 4·2 for n, and the present author's work on polycrystalline specimens of bubbly Antarctic ice led to the same result (Mellor 1959g). Further information on the mechanics of ice deformation, with special reference to Antarctic conditions, is given by Vialov (1958a & b) and Shumskiy (1958b).

Heights measured in Terre Adélie by the French expedition in 1957 have been fitted to an ellipse by Imbert (1959) who gives an expression valid for a distance of about 500km south of Dumont d'Urville:

$$\left(\frac{h}{2{\cdot}68}\right)^2 + \left(\frac{550 - X}{550}\right)^2 = 1,$$

where X is the distance from the coast.

It is interesting to note the implications of this assumed shape if we accept the widely used expression for the bed shear stress (Nye 1951; 1952a & b),

$$\tau = \rho g h \sin \alpha,$$

where α is the surface slope, ρ the ice density, g gravitational acceleration and h the height of the surface.

Since α is small, $\sin\alpha \approx \tan\alpha$ and the expression becomes

$$\tau = \rho g h \frac{dh}{dr};$$

but, as

$$h = H\sqrt{1-\left(\frac{r}{R}\right)^2}$$

and

$$\frac{dh}{dr} = -\frac{Hr}{R^2\sqrt{1-\left(\frac{r}{R}\right)^2}}$$

we find that

$$\tau = -\rho g\left(\frac{H}{R}\right)^2 . r ;$$

that is, for visco-plastic flow the bed stress is a linear function of the radius, the value being a maximum near the coast and zero at the centre of the ice cap. With the dimensions quoted above for the ice cap in Mac.Robertson Land, the bed stress ranges from 0·84 bars near the coast to zero at 1,400km inland. This contrasts with the constant bed stress of 0·88 bars derived by Nye for the Greenland ice cap, under the assumption of plastic flow.

If accumulation can be expressed as a function of the radius (or height) it is possible to obtain a theoretical derivation of the velocity of flow at any cross-section. Such estimates are only of academic interest at present as the theory is not sufficiently well-established for these to be of much value for mass economy considerations, and the estimates of ice flow made in the following pages are based purely on observational data.

3.2. FLOW FEATURES AND TOPOGRAPHIC INFLUENCES

The most important single process in the removal of ice from Antarctica is ice flow, leading to iceberg calving. In Mac.Robertson Land, major features of the subglacial terrain influence the direction and speed of ice flow to a distance of 700km inland and there is an appreciable radial spreading of the ice as strong easterly and westerly components of movement are developed, diverting ice into the Lambert Glacier—Amery Ice Shelf area and to the west of the Enderby Land peninsula (see Map 2).

Where the ice is slowly moving as a sheet into the coastal regions of Mac.-Robertson Land it has a thickness of the order of 2,000m, but within 300km of the coast there is a rapid decrease of depth with the drop in surface elevations and, as the coast is approached, shearing, which is mainly confined to the lowest layers of the thick inland ice, probably becomes appreciable close to the surface, and velocities increase. In a coastal belt about 70km wide, smaller features of the underlying terrain affect the thinner ice and rapid-moving, well defined, ice streams can result (see Plates 9 & 11). These ice streams, which are often referred to simply as "glaciers", may thrust floating tongues a considerable distance out into the sea or they may terminate at the head of a bay. They generally range in size

ANARE photo 7343

PLATE 8

The Lambert Glacier and Patrick Point looking west from the southern extremity of the Mawson Escarpment. Flowing out of the inland ice somewhere south of latitude 74°S, the Lambert Glacier is joined at Patrick Point by major tributaries, the Mellor and Fisher Glaciers, before descending to the Amery Ice Shelf 200km further north. Russian travellers encountered a southern continuation of the glacier trough far inland and it seems almost certain that the Lambert is the world's biggest ice stream.

from about 2km to 15km across. The length of coastline affected by this enhanced flow amounts to about 10% of the total.

Ice flow measurements made by the author in 1957 were aimed at the estimation of typical rates of mass flow for the various types of ice movement. The movement of coastal ice cliffs was measured, sheet flow was measured 12km from the coast, and flow measurements on four ice streams were made. The co-ordinates of a depot 400km inland were accurately determined by astronomical observations so that a re-determination of its position in the future will give some idea of the surface velocity in that area. In 1957 it was not possible to measure the movement of the Amery Ice Shelf and only indirect methods can be used to estimate its calving rate.

3.3. MEASURING TECHNIQUES

In making velocity measurements, two general methods were employed: conventional ground survey and aerial photogrammetry. The first method was used in places which were accessible to survey parties travelling by foot, tractor or aeroplane and the second technique, which was largely experimental, was evolved to

ANARE photo 5566

PLATE 9

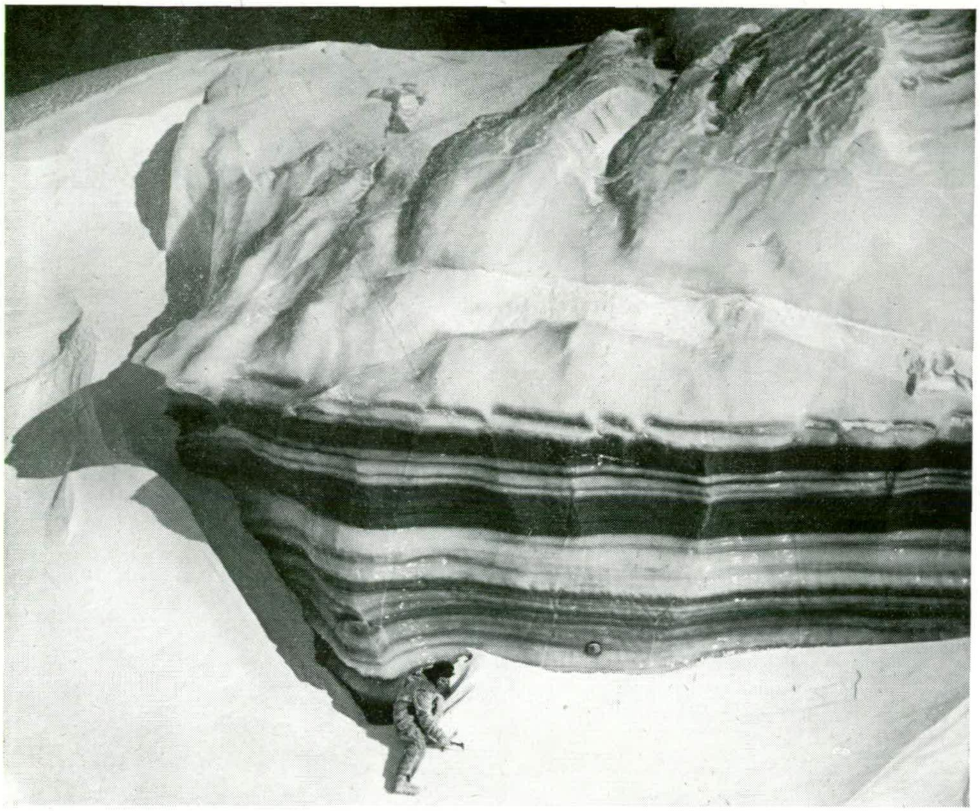
The Robert and Wilma Glaciers are good examples of ice streams. They flow into the King Edward Ice Shelf in Kemp Land.

measure the flow of certain large glaciers inaccessible to ground parties (Mellor 1958b).

The areas where measurements were made are shown on Map 2, and the sections near Mawson are shown in Plate 12. A short description of the work in each locality is given below.

(a) Mawson. Immediately after arrival at Mawson in February 1957, stakes were established along lines runing 1¼km east and west of the rock outcrop on which the base is situated, and approximately 200 metres from the edge of the ice cliffs (see Plate 12 & Fig. 11). Fixed angles were swung from a reference object by a theodolite set up on the Mawson terminal moraine, and offsets from the fixed lines to the stakes were measured directly. The movement was measured three times in ten months. Three stakes were lost by calving.

(b) Henderson—Casey Line. An east-west line of stakes was laid between a nunatak south of Mt Henderson and Casey Range, the distance from the coast being about 12km (Plate 12 & Fig. 12). The line was ranged by theodolite between fixed reference points, and 23 stakes were placed in a distance of 26km. The line



ANARE photo 6965B M. Mellor

PLATE 10

This dark-coloured banding in the basal layers of the continental ice is caused by inclusions of rock fragments. The *C*-axes of the ice crystals in these layers are oriented perpendicularly to the planes of the layers as a result of shear. There are no signs of melting but air bubbles are greatly elongated in the direction of flow.

was run again after 230 days had elapsed and offsets to the stakes were measured directly.

(c) Taylor Glacier. Four stakes were planted on the crevassed tongue of this small glacier along a line defined by theodolite from a neighbouring hill. A triangle was observed on the hills flanking the glacier and one side was measured as a baseline. Angles to the stakes were read from two of the vertices to fix their positions, and the movement of the ice was computed by observing the angular displacements of the stakes from a single point (Fig. 13). Light aircraft were able to land alongside the glacier and repeat measurements could be made in 20 minutes.

(d) Jelbart, Dovers and Hoseason Glaciers. In view of a suspicion that the discharge of ice streams was of great significance to the ice export, it was felt to be highly desirable to secure flow data for some big ice streams and the Jelbart, Dovers and Hoseason Glaciers were singled out for attention. For logistic reasons it was not possible for ground parties to visit these three ice streams, which in any event would have presented great problems in the use of conventional ground survey. The large size of the glaciers, their crevassed surfaces and the lack of fixed control points overlooking the ice all combined to constitute a serious objection to ground survey.

ANARE photo 5572

PLATE 11

The Dovers Glacier, Kemp Land. The fissure running across the glacier is gradually widening and a large berg will eventually be produced. The skerries seen in the picture were used to provide fixed control points for air photogrammetry flow measurements. The centre of the glacier moves about 2½m per day.

Measurements made by triangulation on the small Taylor Glacier tongue (3km wide) suggested that the movement of the big ice streams was sufficiently rapid for measurement by aerial photogrammetry, provided that fixed control points and ground marks on the moving ice could be obtained.

Oblique photographs taken earlier showed numerous rock outcrops and skerries along the coast and it was found that fixed rock control points could be located for each of the glaciers. Moreover, there was photographic evidence that the configuration of crevasses or melt channels could be utilised as ground marks.

Photographic runs, using only the vertical camera of a trimetrogon installation, were planned so that rock islands and outcrops would be included, and the first flights were made in early August 1957. A D.H. Beaver aircraft was used and the runs were made at a true altitude of 3,050m, thus giving a scale of 1:20,000 on photographs taken with a 15·3cm focal length camera. The runs were repeated in December 1957 and the two sets of photographs were brought back to Australia for examination.

The measurements were made by photogrammetrists at the Division of National Mapping, Canberra, after facilities had been made available by Mr B. P. Lambert, Director of the Division. The radial-line method was used to plot fixed control points on rock, and a number of selected common points on the glacier surface were then mapped to the same scale. Comparison of the two positions of corresponding points gave a direct measure of the displacement which had occurred between the two runs. It was found unnecessary to use more than one strip of photos for each glacier, so that only 12 to 16 photos required plotting for each set of measurements.

The period which elapsed between the two sets of surveys was 120 days, and measured displacements at the scale of the prints ranged from 0·13 to 1·73cm, corresponding to velocities of 21 to 290cm per day. Although no independent checks are available, the transverse velocity distributions given by the measurements conform to those which might be expected from established theory.

The method could easily be adapted to measure the movement of the big inland ice streams so long as rock control points and reasonably snow-free ice surfaces are present. As long time lapses are undesirable, the method, as outlined, is probably not suitable for use on glaciers having mid-stream surface velocities of less than about 40cm per day. Flights would be best carried out at the end of one summer and again at the beginning of the next summer, so that the melt-stream patterns would undergo no change. For inland work, use of the radar altimeter would be essential.

In February 1959 the author took oblique air photographs of the Vanderford Glacier in Wilkes Land, defining the principal line with respect to fixed rock features on Browning Peninsula so that it crossed the glacier tongue at right angles to the direction of flow. The 1959 photographs were compared with similar ones taken as part of a trimetrogon survey run in January 1956 to see how the more prominent features of the tongue had changed their position. The general shape of the glacier tongue in 1959 was similar to that of 1956 and, by assuming that no major calving had recently occurred (i.e., the 1959 features were the same ones

U.S. Navy photo

PLATE 12

Coastal ice slopes near Mawson in summer. Black lines on the photograph indicate sections where flow measurements were made. Spray ice can be seen fringing the islands.

seen in 1956, not similar ones newly formed in the interim period), a mean velocity of 180cm/day was deduced. This compares with a maximum surface velocity of 211cm/day measured near Haupt Nunatak on the same glacier by R. Cameron (private communication) in 1957. It therefore seems that the movement of remote glaciers which are flown over only occasionally can easily be checked by taking a few hand-held photographs if suitable rock control is available.

In February 1959 the author also took a run of vertical photographs across the Vanderford Glacier opposite Haupt Nunatak, but on returning to Australia the photographs were found to be very badly overexposed, so that the completion of this project was delayed.

Russian glaciologists in Antarctica used the technique, as did also members of the reconnaissance party of the Expédition Glaciologique Internationale au Groenland (Dolgouchine 1958; Baussart 1958; Hofmann 1958).

It might also be mentioned, in passing, that Russian scientists have estimated the velocity of certain glacier tongues from the deformation produced in fast ice surrounding the glacier snouts (Kapitza 1958b).

3.4. RESULTS

(a) Mawson. The surface movement of the ice is given below; mean flow rates of stakes are plotted in Fig. 11.

TABLE 3.IA

WEST LINE

Period 1957		Stake No. and Distance from Theodolite 1 (335 m)	2 (503 m)	3 (670 m)	4 (837 m)	5 (1105 m)	6 (1170 m)
17/2-11/6	Shift (m)	0	2·41	3·56	3·53	3·02	2·78
(114 days)	Vel. (cm/dy)	0	2·1	3·1	3·1	2·6	2·4
11/6-16/9	Shift (m)	0	1·62	1·77	1·83	1·74	0·94
(97 days)	Vel. (cm/dy)	0	1·7	1·8	1·9	1·8	1·0
16/9-10/12	Shift (m)	0	2·26	3·72	4·48	4·24	4·88
(85 days)	Vel. (cm/dy)	0	2·7	4·4	5·3	5·0	5·7
Total	Shift (m)	0	6·27	9·05	9·84	8·98	8·60
(296 days)	Vel. (cm/dy)	0	2·1	3·1	3·3	3·0	2·9

Three stakes from the east line were lost when the ice cliffs calved on 3rd April 1957. Overall movements of the remaining two stakes are given:

TABLE 3.IB

EAST LINE

Period 1957		Stake No. and Distance from Theodolite 1 (292 m)	2 (655 m)
18/2-10/12	Shift (m)	1·59	13·60
(295 days)	Vel. (cm/dy)	0·5	4·6

The depth of ice along the measured sections is about 120m, a value which would be fairly typical for this length of coast. The higher ice velocities would probably be representative for the coast.

(b) Henderson—Casey. The stakes were placed in May 1957 and re-surveyed in January 1958. The movements and flow rates from May 1957 to January 1958 are given below, and surface velocities are plotted against ice depths, measured by seismic methods, in Fig. 12. A later survey was made in the autumn of 1958 by G. Knuckey and I. McLeod. The last two columns of Table 3.II show the flow rate in the autumn of 1958 and the overall mean velocity from May 1957 to autumn 1958.

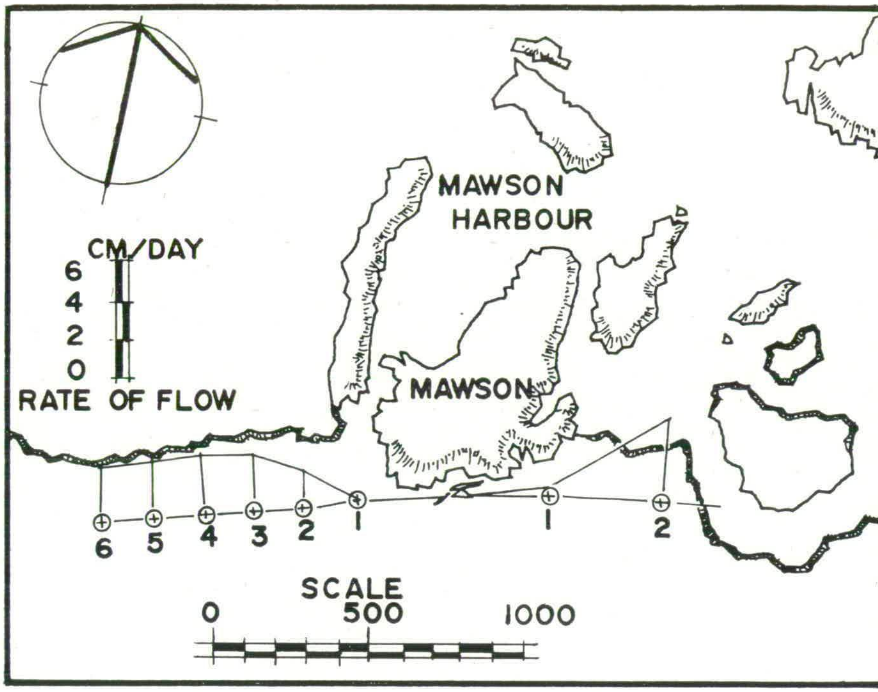


FIG. 11. Movement of the ice adjacent to Mawson.

TABLE 3.II

Stake No.	Movement (m) (May 1957-Jan. 1958)	Rate of flow (cm/day) (May 1957-Jan. 1958)	(Autumn 1958)	(May 1957-Autumn 1958)
0	—	—	—	2·8
1	8·4	3·7	3·9	3·7
2	9·0	3·9	4·5	4·0
3	11·6	5·0	4·1	4·8
4	14·5	6·3	3·1	5·2
5	14·0	6·1	6·1	6·1
6	9·9	4·3	3·9	4·1
7	6·4	2·8	2·6	2·7
8	7·3	3·2	3·0	3·1
9	8·5	3·7	3·7	3·7
10	9·6	4·2	4·5	4·2
11	10·1	4·4	3·7	4·1
12	10·5	4·6	3·7	4·2
13	10·7	4·7	6·9	5·3
14	13·7	6·0	5·5	5·8
15	15·6	6·7	7·0	6·8
16	17·1	7·4	6·4	7·0
17	20·0	8·7	9·1	8·9
18	25·9	11·3	7·6	10·0
19	23·8	10·4	6·7	9·1
20	22·7	9·9	9·1	9·6
21	—	—	—	10·5
22	—	—	—	11·7

Crevasse hazards prevented re-survey of certain stakes in January 1958.

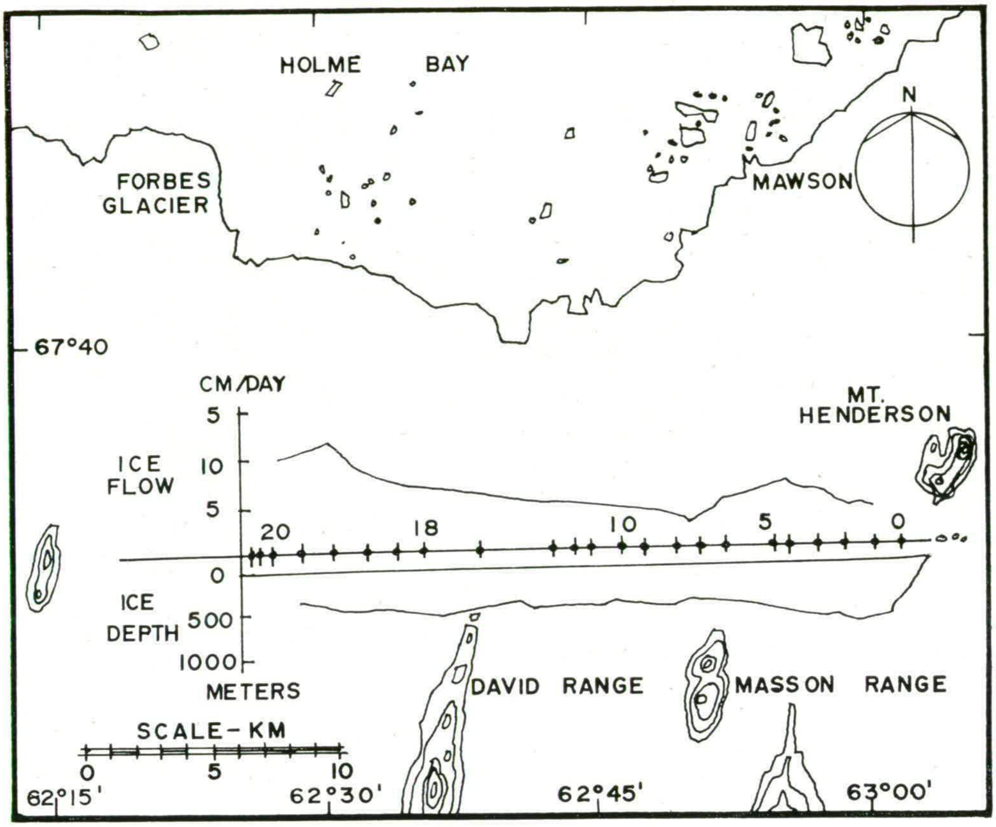


FIG. 12. Ice flow between Mount Henderson and Casey Range, Mac. Robertson Land. The ice depths along the stake line are shown diagrammatically.

(c) Taylor Glacier. The placing of stakes on this glacier tongue was limited by adverse weather conditions and crevasses. Variation in the rates of movement are shown in Fig. 13.

TABLE 3.III

Period		Stake No. 1	2	3	4
6/7/57-30/8/57	Movement (m)	17·2	17·8	18·4	19·0
(55 days)	Rate (cm/day)	31·3	32·4	33·4	34·6
30/8/57-27/10/57	Movement (m)	15·6	15·7	19·4	19·6
(58 days)	Rate (cm/day)	26·9	27·1	33·5	33·8
27/10/57-30/11/57	Movement (m)	13·3	14·6	11·8	11·9
(34 days)	Rate (cm/day)	39·2	43·0	34·8	35·1
Total	Movement (m)	46·1	48·1	49·6	50·5
(147 days)	Rate (cm/day)	31·4	32·8	33·8	34·4

(d) Jelbart Glacier. The readings given below are of limited value without a photographic mosaic or detailed map, consequently the estimated mean surface velocities for each of the streams that were measured photogrammetrically are given.

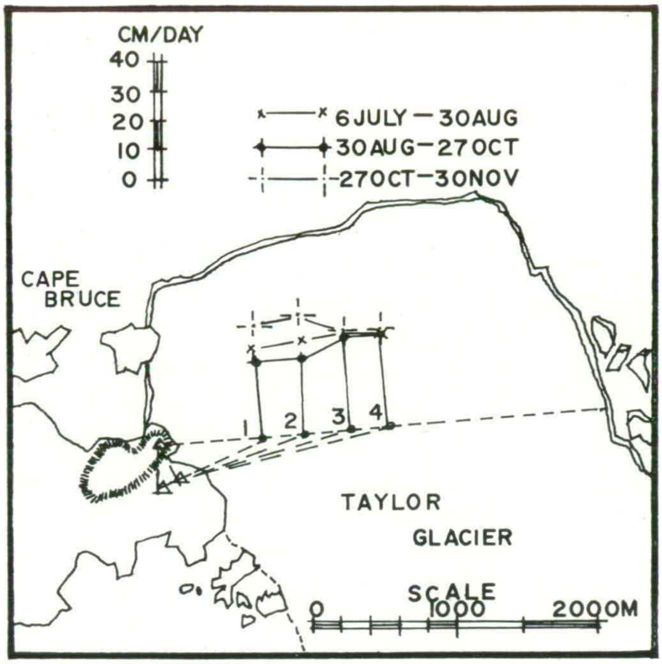


FIG. 13. Movement of Taylor Glacier.

TABLE 3.IV.A

EAST TONGUE (PERIOD OF 118 DAYS)

Point	Movement (m)	Rate of Flow (cm/day)
1	97	82
2	107	91
3	112	95
4	101	86
5	96	81

Mean surface velocity 88cm/day.

TABLE 3.IV.B

WEST TONGUE (PERIOD OF 118 DAYS)

Point	Movement (m)	Rate of Flow (cm/day)
1	16	14
2	33	28
3	25	21

Mean surface velocity 23cm/day.

(e) Dovers Glacier. The points lie on a line crossing a portion of the glacier which is due to produce a large berg. The mean surface velocity is corrected for the widening of the crack seen in Plate 11.

TABLE 3.V

PERIOD—121 DAYS

Point	Movement (m)	Rate of Flow (cm/day)
1	26	21
2	94	78
3	172	142
4	214	177
5	229	190
6	261	216
7	308	255
8	323	267
9	355	294
10	104	86

The mean velocity for the central section of the stream was 215cm/day and the mean surface velocity across the full width of the stream 200cm/day.

(f) Hoseason Glacier. This glacier is quite a small one and the measurements were made across a land-based section of the ice stream.

TABLE 3.VI

PERIOD—125 DAYS

Point	Movement (m)	Rate of Flow (cm/day)
1	104	83
2	109	87
3	130	104
4	130	104

Mean surface velocity 95cm/day.

3.5. DISCUSSION

Some idea of the rate of movement of the coastal ice cliffs can be gained from the Mawson measurements. The higher velocities are more likely to be representative, as the lower speeds were measured close to exposed rock.

There is a temptation to see indications of seasonal fluctuation in the flow rates given for the Mawson west line. Such a variation in the surface velocity is possible, but the evidence is too scant to permit more than speculation.

The movement of the Henderson—Casey line, 12km inland, gives a mean surface velocity of 6cm/day, a figure which is surprisingly high compared with the movement of the coastal ice cliffs. The west end of the line extended into the Forbes Glacier, however, and measurements made on this length raise the mean value for the line by about 10 per cent. If a mean velocity of 5cm/day is taken for the general sheet flow, the ice will suffer a thinning of about 200m on its passage down to the coast from ablation at the prevailing rate. The mean ice thickness along the line is 550m; if the full thickness of the ice is in motion there should be a depth of 350m at the coast, according to the measured values of velocity and ablation. The actual depth at the coast does not exceed 200m, so that, if continuity is to be satisfied, we must either postulate the existence of basal depressions containing ice which is virtually stagnant or assume a progressive change in the vertical distribution of velocity. The data are again too slight for anything beyond speculation.

The measurements discussed above suggest that the average movement of the ice cliffs along this coast is approximately 4cm/day.

The Taylor Glacier stakes show that even a small coastal glacier moves much faster than the ice sheet generally, the velocities being almost an order of magnitude higher. The irregular way in which the stakes moved indicates that the rock constraints at the tongue tend to produce intermittent motion.

The measurements on the Jelbart, Dovers and Hoseason Glaciers bring out the significance of the work done by ice streams in removing ice from the continent. The Jelbart and Hoseason Glaciers have velocities over 20 times as great as the general ice sheet; the Dovers Glacier moves at 50 times the speed of the ice cliffs to east and west. Although ice streams take up no more than about 10 per cent of the total length of coastline, it is evident that they remove at least as much ice and probably considerably more, than the sheet flow over the remaining length of coast.

Loewe (1956) quotes the measurements of Perroud and Mawson, which give flow rates for the coastal ice of Terre Adélie of a few metres a year and 30m a year respectively. Loewe doubts the validity of these figures as characteristic velocities for the ice sheet, but measurements in the Mawson region give similar low speeds of ice flow. The mean velocities which Loewe quotes as necessary to maintain constant ice volume are an order of magnitude higher than the measured ones, and are based on the mass flow necessary to remove an assumed accumulation. The accumulation figure of 10cm of water taken by Loewe seems reasonable when compared with measurements in Mac.Robertson Land and elsewhere, but the effect of ice streams is not considered (a surface velocity of 82cm/day is quoted for the Glacier de Zélée in Loewe (1956)—page 105. If we assume that sheet flow takes place over 90 per cent of the coastline at a speed of 4cm/day, and that "glacier" or ice stream flow at 100cm/day takes up 10 per cent of the coastline, then we can see that ice streams are about three times as effective as sheet flow in removing ice from the continent. Whilst this does not raise the berg formation rate to anything like that considered necessary for equilibrium by Loewe, it does improve the agreement somewhat.

Wright and Priestley (1922) quote a number of measurements and estimates of ice velocities, chiefly in the McMurdo Sound area. These range from 0·3 to 85cm/day, but it is difficult to assess the significance of the figures without further information. Measurements of Drygalski at Gaussberg are also quoted by Wright and Priestley, the movement of the continental ice there being from 33 to 44cm/day. The continental ice immediately to east and west of Gaussberg, however, is of the stream type (Mawson 1942) and these velocities are similar to those measured on the Taylor Glacier in Mac.Robertson Land.

More recently, US glaciologists of Wilkes base made measurements on the Vanderford Glacier (longitude 111°E) during 1957 and established a velocity of 211cm/day (private communication, R. Cameron). The Vanderford is a good example of a medium-sized ice stream and is moving at roughly the same speed as the Dovers Glacier in Kemp Land. This confirmation of high velocities indicates that 100cm/day is not likely to be an over-estimate for the average velocity of ice streams overall.

3.6. CONCLUSIONS

The measurements show that continental ice reaches the sea or ice shelf by two types of movement—sheet flow and stream flow. The rate of sheet flow must vary to some extent with ice depth and the nature of the subglacial topography, but a figure of 4cm/day as the mean velocity is likely to be of the right order. Ice stream velocities are determined mainly by their dimensions, but most streams would have mean velocities lying in the range 30 to 300cm/day. An overall average figure of 100cm/day would be of the right order and would probably not be high.

Vertical velocity profiles are not considered here, but the export estimates have been roughly compensated by accepting rather low surface velocities.

The calving rate of ice shelves is difficult to estimate from pre-IGY data, but a rough calculation can be made. Shackleton (Wright and Priestley 1922) measured the movement of the Ross Ice Shelf as 450m/yr, and Gould (1957) gives an estimate of 530m/yr for the same shelf. Swithinbank (1955) gives a figure of 300 m/yr for the movement of the ice shelf at Maudheim, and also proves an equilibrium thickness of about 190m for the shelf ice. 400m/yr might be used as a working estimate of movement.

Rough export estimates for the whole of Antarctica are given below, assuming 140m as the coastal thickness of the continental ice, $0{\cdot}87$ g/cm^3 as its density, and $0{\cdot}84$g/cm^3 as the mean density of shelf ice:

Ice shelf (7,500km)	$0{\cdot}48 \times 10^{18}$g/yr
Sheet flow (11,000km)	$0{\cdot}02 \times 10^{18}$g/yr
Stream flow (1,500km)	$0{\cdot}07 \times 10^{18}$g/yr

These figures bring out the great importance of ice shelves in iceberg production. The export quantities for sheet flow and stream flow may be taken as reasonable estimates, but it is apparent that the calving rate for ice shelves is calculated from very slight data. Nevertheless, these quantities can be used until further glaciological and cartographic information becomes available.

4. ABLATION

4.1. CAUSES OF ABLATION AND ABLATION ZONES

Under the heading of ablation are included ice losses resulting from direct evaporation from snow and ice surfaces (sometimes called sublimation), evaporation of meltwater, and run-off of meltwater to the sea.

Evaporation can occur so long as the air above the surface is not saturated at the prevailing temperature; air flowing down the continental ice slopes under gravity warms as it nears the coast and its relative humidity falls, making evaporation possible if no snow is being carried by the wind. Melting can take place when solar radiation is absorbed by the ice and this often occurs with air temperatures below freezing. Convective heat plays little part in producing melting. Clean snow surfaces reflect a high proportion of the incident radiation and snow is capable of holding free melt water in its pores. Bare, hard ice, on the other hand, has a lower albedo than snow and so absorbs more radiation; thus more intense melting results and the melt water flows downhill over impermeable ice until it reaches the sea or drains into crevasses. For this reason sloping marginal areas which are swept clear of snow by strong winds are liable to considerable ablation losses.

Around Mawson the ablation zone is a coastal strip having a width of some 20km (see Plate 12), but the width of the ablation zone and the height of the firn limit vary from place to place along the coast between longitude 45°E and 80°E. At some points there is a permanent snow cover and the firn limit is at sea level. Examples of this are the small domed ice caps of Fold Island, Broka Island, Law Promontory and Casey Bay, and the Amery and King Edward Ice Shelves. On the other hand, ablation areas may stretch for long distances inland and the firn limit can be as high as 950-1,000m. Plate 15 shows the snow-free ice surface and part of a big melt channel near Clemence Massif, some 400km from the sea. Net ablation areas extend well inland in parts of Enderby Land, and on the Lambert Glacier in Mac.Robertson Land melt channels cut into bare hard ice 450km from the sea and at an elevation of 950m. This same glacier feeds north into the Amery Ice Shelf, where the firn limit is virtually at sea level. The so-called Antarctic "oases" are examples of extended net ablation zones. The Vestfold Hills, a typical "oasis" receive a winter snow cover which is removed by rapid melting during the summer, when the big areas of dark rock absorb a large percentage of the solar radiation. Snow melting is accelerated by grains of wind-blown sand lying on the surface. Summer melting also extends to the continental ice south of the Hills. To the south of the Bunger Hills, also, there is intense summer melting and great melt lakes form on the ice.

ANARE photo 7420

PLATE 13

Permanently snow-free ice near Clemence Massif, Mac.Robertson Land. This area is south of the Amery Ice Shelf, but strong summer melting occurs. Surface drainage leads to the formation of large river systems and a melt-river can be seen on the right of this photograph.

There is ample evidence that this alternation of wide ablation zones (indicated by expanses of bare ice and glacier-free hills) and coastal accumulation areas (such as ice shelves and ice-capped islands) is maintained along the whole length of the coastline of Australian Antarctic Territory. It seems likely that the width of the ablation zone on the coast of Mac.Robertson Land is determined mainly by the speed and relative direction of the prevailing wind. This view was formed as a result of careful observations from the air, using sastrugi as indicators of katabatic winds, obtaining directions of cyclonic winds from offshore snowdrifts, and comparing conditions on different sides of bays and promontories. Wind speeds at Taylor Glacier, where the firn limit is several hundred metres lower than in the Mawson area, were significantly lower than those at Mawson. Swithinbank (1957) has shown that surface slopes lead to irregularities of accumulation, snow generally being removed from downslopes and being deposited on flat areas or in depressions. Similar effects were noticed earlier by Wright and Priestley (1922). The inland ablation areas associated with belts of ice domes are almost certainly formed by wind-scouring of locally disturbed ice. Alternate erosion and deposition can be seen all along the Mac.Robertson Land coast and ground observations, together with stereo examination of trimetrogon photographs, confirm that the humps and down-slopes suffer erosion whilst deposition occurs in hollows and on flat areas.

In December and January the melting is intense below 200 metres altitude and deep melt channels are cut. The ice surface fractures to a depth of about 5cm, the granules themselves being polygonal in cross-section, 0·5 to 2·5cm in diameter, and 2 to 6cm long. Thin sections were made from some of these granules and a rather hurried inspection under the polariser indicated that they were single crystals with their C-axes vertical. Similar ablation surfaces are common on temperate and Arctic glaciers and experiments have shown that preferential melting may occur as a result of a concentration of impurities at the crystal boundaries (Renaud 1949). An attempt to reproduce the effect by exposing ice slabs to infra-red radiation in sub-zero air temperatures was unsuccessful.

Stones or sand lying on the ice absorb radiation and sink into the ice during the summer, forming cryoconite holes. These cryoconite holes are between 60cm and 90cm deep and have the usual air tubes radiating from a central pipe. Dirty snow commonly gives rise to micropenitent formation during the early stages of the thaw.

4.2. MEASUREMENT OF ABLATION AND ABLATION CHARACTERISTICS

For the measurement of ablation, stakes were used, although there has been some criticism of the stake method due to the falling of untended stakes in summer when solar radiation is intense. Several methods were tried at Mawson but ordinary bamboo stakes were found most satisfactory. Tapered bamboo stakes, 2·5m long, were painted white to reflect radiation and were set deep in 2·5cm-diameter holes so that no more than 30cm of the thin end of the stake projected vertically above the surface in spring. Separate marker flags were required, but the measuring stakes themselves did not become displaced.

Evaporation proceeds at an appreciable rate throughout the year in Mac.Robertson Land and Princess Elizabeth Land. When mean winter ablation rates of 0·7mm of water per day were first reported by R. Dovers (Loewe 1956a) from Mawson they were thought to be exceptional, but subsequent observations by P. Crohn and by the present writer have confirmed the early measurements. At 425m above sea level and 16km inland from Mawson midwinter ablation rates are still as high as 0·6mm of water per day and it is believed that a mean rate of 0·5mm per day could well apply between 20 and 50km inland.

In 1958, stakes were placed alongside depots established by aircraft and were later re-measured by field parties (McLeod 1959). A bare ice area 1km north of the Leckie Range, 60km from the coast in Kemp Land, lost 12cm of ice between late September and the end of December; this probably represents about 1mm of water per day in that period. Immediately to the north of the nunataks situated at approximately 67°30′S, 52°45′E, there was a loss of 1·9cm of ice between late September and early November (about 0·4mm of water per day) and a further loss of 10cm of ice between early November and mid-December (about 2·3mm of water per day). Both these places are at an altitude of about 1,200m. As the measurements were made on hard ice, wind erosion can be discounted but a certain amount of melt water produced in December would re-freeze in another place. Nevertheless, these observations support the belief that appreciable evaporation may still take place 50km inland. Considerable ablation is reported from Beaver Lake (70°47′S, 68°20′E) but the rather complicated nature of this place (Mellor & McKinnon 1959) makes it difficult to assess the significance of the figures.

As early as October, absorbed radiation can produce surface melting on the ice slopes with negative air temperatures, and evaporation rates rise. Melting becomes more frequent in November, and in December and January the bare ice of the ablation zone melts regularly. Snow surfaces, with their higher albedos, suffer less melting but blue ice bands are formed on the ice-capped islands and big coastal drifts. Summer melting is intense on the bare ice below the 200m level, and between heights of 200 and 500m there is still sufficient water to form melt streams. The steep slopes near Mawson drain melt water efficiently and there is no significant formation of superimposed ice, but in other regions, e.g., Wilkes Land, there is widespread occurrence of superimposed ice. The determination of ice losses in a superimposed ice area will be rather difficult as there is a continuous transfer of material, making it necessary to gauge melt water actually passing into the sea.

Summer melting on the lower slopes brings the annual net ablation there to well over 500mm of water, but it seems that summer conditions do not greatly enhance the mean ablation rate at 500m altitude, possibly because the smaller slopes produce only a slow migration of melt water.

The mean wind velocity of Davis is less than half that of Mawson and the air has a higher humidity. Measurements by R. Dingle and B. Stinear in the Vestfold Hills show that winter ablation is less intense than in the Mawson area, a typical rate being 0·3mm per day against 0·7mm per day at Mawson. In spring, the evaporation rate rises to a value which is comparable with the Mawson one: 1·3mm against 1·5mm per day. The mean air temperatures at Vestfold Hills in December and January are —0·2°C and —0·3°C respectively and heavy melting, followed by a certain amount of runoff to the sea, occurs. There are no ablation figures available for the summer months but it seems unlikely that the evaporation rate could be lower than 2mm of water per day, since 2·3mm per day has been measured in late November.

Both gauge and stake readings gave a result of about 5·5cm for the precipitation from the end of March to the end of November 1957, and evaporation in the same period was close to balancing precipitation during the winter. Although summer precipitation is only about 1cm of water, lakes in the Vestfold Hills receive a considerable inflow from melting snow. The lakes have been examined by a geologist, Bruce Stinear, and it is believed that their levels are fairly constant, in spite of the summer inflow and the absence of surface outlets or subsurface drainage. This indicates a high summer evaporation over the limited area of the lakes.

On the evidence collected so far there seems no reason to believe that ablation losses in Princess Elizabeth Land are substantially less than those in Mac.Robertson Land.

4.3. RESULTS

A summary of ablation data for the Mawson area is given in this section. The tables are based mainly on the writer's observations in 1957, but comparisons have been made wherever possible with the measurements made by R. Dovers, 1954, P. Crohn, 1955 and 1956, and I. McLeod, 1958. More weight has been given to the 1957 measurements, since they generally represent mean readings from several stakes, whereas in other years only single stakes have been used.

TABLE 4.I

TOTAL NET ABLATION (22 FEB. 1957 TO 26 FEB. 1958)

Altitude (metres a.s.l.)	Net Ablation (mm of ice)	Net Ablation (mm of water)	
60	625	535	
150	575	495	
180	570	490	
305	300	260	
335	270	230	
365	255	220	
425	275	235	(single measurement)

All the above measurements were made on bare ice.

The figures given in this table permit an estimate of the annual loss of ice to be made. Working from a contour map of the area and extrapolating the results to the firn limit, we arrive at a figure of $5 \cdot 3 \times 10^{12}$gm of water as the annual loss in a strip 1km wide, running south from Mawson.

TABLE 4.II

MEAN ABLATION RATES—SUMMER AND WINTER

Altitude (metres a.s.l.)	Ablation Rate (mm of water per day) Summer (Nov.-Feb.)	Winter (May-Jun.)
60	2·92	0·74
150	2·73	0·63
180	2·64	0·66
305	1·23	0·65
335	0·91	0·65
365	0·87	0·64
425	0·90	0·63

This illustrates the fall in ablation with increase of altitude, over the summer months, and the very small effect of changes of altitude during the winter. The winter ablation rate of 0·6 to 0·7mm of water is also maintained at all measured levels during the months of July and August. In 1958 the mean evaporation rate between 110 and 180m elevation in May and June was 0·65mm of water per day, which is in agreement with the 1957 values. The 1958 measurements also show that there was no decrease of ablation with increasing altitude between late February and mid-November.

TABLE 4.III

NET ABLATION ON MAWSON SNOWDRIFTS

Period (1957)	Mean Lowering of Snow Surface (mm)	Mean Ablation Rate (mm of water per day)
12 July-29 July	29·2	0·72
29 July-9 Aug.	6·3	0·24
9 Aug.-23 Aug.	15·9	0·48
23 Aug.-2 Sept.	12·7	0·53
2 Sept.-27 Sept.	31·7	0·53
27 Sept.-26 Oct.	66·5	0·96
26 Oct.-7 Nov.	34·8	1·22
7 Nov.-28 Nov.	63·5	1·29
28 Nov.-10 Dec.	38·1	1·40

The accumulation over the measured surfaces of the Mawson snowdrifts was negligible after the end of June, and the measurements thus give reliable figures for gross snow ablation.

TABLE 4.IV

ANNUAL NET ABLATION AT 60 METRES ABOVE SEA LEVEL, 1955-1958

Period	Net Ablation (mm of ice)	(mm of water)
22 Feb. 1955-16 Mar. 1956	915	790
1 Jan. 1956- 1 Jan. 1957	850	730
16 Mar. 1957-26 Feb. 1958	625	535

In 1958, measurements did not cover a complete year. The net ablation from mid-March 1958 to the beginning of February 1959 amounted to 340mm of ice at 60m elevation and to 490mm of ice at the 90m level.

The figures indicate that the annual net ablation in 1957 was considerably lower than in the two previous years. The 1957 result was a mean of 8 readings

TABLE 4.V

MONTHLY MEAN ABLATION RATES, 1954-1957 (60 METRES A.S.L.)

Month	Net Ablation Rates (mm of ice per day) 1954	1955	1956	1957
Jan.	—	—	11·4	—
Feb.	—	—	3·7	—
Mar.	—	{1·1	1·5	1·3
Apr.	—	1·1	1·0	1·2
May	{0·8	{0·8	0·8	0·9
Jun	0·8	0·8	0·8	{0·9
Jul.	0·8	0·6	0·5	0·9
Aug.	{1·1	0·8	0·6	0·6
Sep.	1·1	0·6	0·8	0·8
Oct.	{1·5	1·4	0·9	1·4
Nov.	1·5	1·5	1·9	1·7
Dec.	—	5·7	8·5	(4·8 mid-Dec. to late Feb.)

Bracketed figures indicate mean values over 2 or more months.

made along a strip 2km long, whilst the 1955 and 1956 quantities are derived from a single measuring point. It is unfortunate that a measurement was not made in 1957 at the site of the earlier ones, as this would have given a better comparison. The maximum deviations from the mean value for the 60-metre stakes in 1957 were —18% and +30%.

Measurements made in 1958 by McLeod are hardly comparable, since he mentions that the stake at the 60m level was placed in a sheltered position and his readings show considerably more ablation at 90m elevation.

The monthly ablation rates for the 4 years tabulated show a good agreement and may, therefore, be accepted with some degree of confidence. There is a depressing lack of data for the summer months, due to the departure of observers on inland journeys and the lack of continuity through the relief operations. Ablation rates as high as 27·5mm of ice per day were measured in early January 1956, and the mean ablation rate for January 1958, was probably (unreliable measurements) about 10mm per day. All the measurements were made at a distance from the sea of less than 1km.

TABLE 4.V.A

	Month	Net Ablation Rate (mm ice/day) 60m	90m
1958	Jan.	—	—
	Feb.	—	—
	Mar.	0·7	0·9
	Apr.	0·2	0·9
	May	0·6	0·8
	Jun.	0·4	0·8
	Jul.	0·2	0·7
	Aug.	0·4	0·4
	Sep.	0·2	1·1
	Oct.	1·6	1·8
	Nov.	1·8	4·1
	Dec.	2·9	2·7
1959	Jan	3·3	3·3
Early	Feb.		

Table 4.V.A gives daily rates of ice loss from the middle of one month to the middle of the next.

Ablation measurements were made on three lakes in the Vestfold Hills during 1957 by W. R. J. Dingle, Officer-in-Charge of Davis Station. The lakes all received a snow cover periodically, but this was removed, redistributed, or compacted by

TABLE 4.VI

ABLATION ON LOOKOUT LAKE (15 METRES ABOVE SEA LEVEL)

Period (1957)	Ablation Rate (mm of ice per day)
14 June-20 July	0·30
22 August-4 October	0·15
4 October-16 October	0·79
16 October-30 October	0·79
30 October-6 November	0·80
6 November-13 November	1·36
13 November-20 November	1·36

wind, so that the measurements do not give ablation rates for a snow surface. Tables 4.VI to VIII give ablation rates on the three lakes.

There was a net loss of ice of 54mm on Lookout Lake between 14 June and 20 November. A snow cover over the measured surface persisted from 20 July to 22 August, and some fusion between the lying snow and the lake surface appeared to take place. By 4 December the lake was 90% free of ice.

TABLE 4.VII
ABLATION ON STATION LAKE (13·5 METRES ABOVE SEA LEVEL)

Period (1957)	Ablation Rate (mm of ice per day)
18 September-9 October	0·30
9 October-23 October	0·79
23 October-6 November	0·57
6 November-20 November	0·80
22 November-27 November	2·50

The lowering of the ice surface on Station Lake amounted to 37mm between 13 June and 20 November. Snow lay over the measuring area from 21 June to 20 July, and in that period a snow layer 220mm deep was dissipated. Snow again drifted on the surface and persisted from 30 July to 18 September. The ablation rates listed are for bare ice. The lake was 80% free of ice by 4 December.

TABLE 4.VIII
ABLATION ON CAMP LAKE (18·5 METRES ABOVE SEA LEVEL)

Period (1957)	Ablation Rate (mm of ice per day)
28 June-14 July	0·21
15 September-11 October	0·37
11 October-24 October	0·98

Camp Lake suffered a net ice loss of 29mm between 28 June and 24 October. The figures quoted are for the ablation of bare ice, the measuring area being snow-covered during the intervening periods. The ice on the lake was breaking up on 5 December, but a 90% cover remained.

It may be of interest to give here some ablation figures for sea ice at Mawson and at Cape Bruce (67°27′S, 60°50′E).

TABLE 4.IX
SEA ICE ABLATION AT MAWSON

Period (1957)	Net Ablation Rate (mm of ice per day)
31 May-19 June	2·00 (some snow)
19 June-12 July	0·70
12 July-29 July	0·56
29 July-9 August	0 (snow)
9 August-23 August	0·69
23 August-2 September	0·32
2 September-27 September	0·25
27 September-26 October	1·32
26 October-7 November	1·07
7 November-28 November	4·55

The sea ice at Mawson was generally kept clear of snow by the wind, and most of the values listed are for the ablation of bare ice. The rates are similar to those measured on the Vestfold lakes, although the ablation of both sea ice and lake ice is rather less than that occurring on the plateau ice. The sea ice ablation at Mawson was intensified in the early summer by the presence of a layer of rock dust and coal dust downwind of the camp area.

TABLE 4.X

SEA ICE ABLATION AT CAPE BRUCE

Period (1957)	Net Ablation Rate (mm of ice per day)
7 July-23 October	0·23
23 October-30 November	3·00

The net ablation of sea ice at Cape Bruce was less than that at Mawson. Between 7 July and 30 November Mawson lost 185mm of ice and Cape Bruce lost 114mm; the difference is believed to be due to lower winds and more frequent snow cover at Cape Bruce.

The stake measurements made on the sea ice at Davis are not included here, since frequent snowfalls make the derivation of ablation rates difficult.

4.4. DISCUSSION

The coasts of Terre Adélie are somewhat similar to those of Mac.Robertson Land, and Loewe (1956) records a number of ablation measurements made by himself and others on the coast of Terre Adélie. In this region the firn limit is at an altitude of 500 metres, some 300 metres lower than in the Mawson area. There are, however, places within 80km of Mawson where the firn limit is lower than 500 metres above sea level. There is permanent bare ice along the Terre Adélie coast, but melt streams are not formed during the summer. The annual mean temperature at the firn limit in Terre Adélie is about —18°C; south of Mawson the annual mean temperature is also about —18°C at the firn limit, though the altitude is greater.

The table below compares the ablation values of Loewe with corresponding ones for the Mawson area (close estimates).

TABLE 4.XI

Period	Terre Adélie (40 metres a.s.l.) (mms of ice)	Mawson (60 metres a.s.l.) (mms of ice)		
		1955	1956	1957
mid February to mid March	60 (névé & ice)	—	60	40
mid February to end of April	50 (1951)	80	110	70
mid December to mid July	400 (1951-52)	—	720	480
end of April to end of October	30 (1951)	210	130	170
end of October to 1 December	60 (névé & ice)	50	60	55
end of February to end of November	130 (1951)	270	260	270

It is difficult to draw any firm conclusions from such a comparison, but certain impressions may be formed. It does seem that winter ablation is more pronounced at Mawson than in Terre Adélie, in spite of the lower mean wind velocities of the former. The greater frequency of drifting snow in Terre Adélie, however, must produce more humid conditions, which probably limit the winter ablation. The spring and autumn ablation quantities appear to be similar for both regions, and the entry "mid-December to mid-July" suggests a fairly intense summer ablation for Terre Adélie. The absence of melt channels on the coastal ice of Terre Adélie makes this inference rather questionable.

Both Swithinbank (1957) and Wade (1945) conclude from their studies that ablation losses on the ice shelves at Maudheim and Little America are negligible. In McMurdo Sound, however, David and Priestley (1914) estimate that 180mm of ice are lost by evaporation each year. Wright and Priestley (1914) suggest annual ablation losses from 150mm to 300mm for other places in the same locality. Whilst these figures are lower than corresponding ones for the Mawson area, they still represent an appreciable loss.

4.5. CONCLUSIONS

Summing up, we may say that ablation proceeds continuously throughout the year on the coast of Mac.Robertson Land and constitutes an important item of ice wastage. A loss of $5 \cdot 3 \times 10^{12}$g/km/yr between the coast and the firn limit has been estimated, and it was also suggested that a mean evaporation rate of $0 \cdot 5$mm per day might apply between 20 and 50km inland, giving an additional loss of $4 \cdot 5 \times 10^{12}$g/km/yr.

The relative importance of the various ablative processes changes from season to season. During the summer, ice is removed chiefly by melting followed by run-off and evaporation. Radiation is very much more important than convection, which has little effect 2km inland. Melting does not occur during the winter, but a substantial proportion of the annual ice loss is effected by direct evaporation. The rate of evaporation is influenced by wind speeds, temperature, and relative humidity of the air. In the spring and autumn, direct evaporation proceeds, but the absorption of solar radiation can produce surface melting and evaporation at air temperatures well below the freezing point.

If mean temperatures and pressures are assumed for the heart of Antarctica and for the fringe regions, then a comparison of saturation mixing ratios for the air of both zones can be made. If $-45°$C and 700mb are assumed for the inland conditions and $-20°$C and 1000mb for the temperature and pressure of the coastal regions, the saturation mixing ratios are $0 \cdot 064$ and $0 \cdot 646$g/kg respectively. Thus the air flowing north is theoretically capable of taking up more moisture on descending to the warmer coastal areas, and so evaporation may occur in all the steeply sloping fringe regions.

Details of the ablation processes in the Mawson region could be derived from a study of ablation measurements in conjunction with simultaneous meteorological observations and it is hoped that this will be done at a later date. However, on the data available at the time of writing, even conservative estimates of total ablation loss in Antarctica lead to a figure which is worthy of inclusion in the mass balance of the continent.

5. WIND-BLOWN SNOW

5.1. GAUGING DRIFTING SNOW

5.1.1. *Introduction*

Thousands of kilometres of the coastline of Antarctica are swept by fierce winds which blow down the slopes of the high ice plateau and carry large quantities of snow out to sea or on to ice shelves. In order to assess the significance of this snow transport it is necessary to estimate the amount of snow being moved across the coast annually but, to do this with any accuracy the frequency, intensity and vertical distribution of the drift must be measured at numerous points around the continent.

Probably the most detailed consideration of the problem prior to the IGY arose from the work of Loewe in Terre Adélie during 1951, although several previous attempts to measure the quantities of drifting snow have been made.

The first recorded measurements were made in Spitsbergen during the First Polar Year, 1882-83, by Andrée (1883), who passed snow-filled air through a bag which trapped the snow whilst permitting the passage of air (something like a vacuum cleaner). In 1908 Wegener (1911) made some measurements in north-east Greenland, and in 1912 a drift collector was installed at Cape Denison, the base of the Australasian Antarctic Expedition (Mawson 1915). This last drift trap consisted of a large wooden box fitted with a metal cone, the apex of which pointed into the wind. Drifting snow entered through a small hole at the apex of the cone, snow particles settled out in the voluminous box, and snow-free air passed out through a gravity-closing trapdoor on the downwind side of the box.

A replica of the Cape Denison box was built by the French party wintering in Terre Adélie in 1950 (Loewe 1956). In 1951 this trap continued to be used by Loewe (1956), who also constructed a streamlined gauge and another which trapped a sample of snow-filled air by simultaneous closure of inlet and exit ports. In the same year at Port Martin, Barré (1954) made blizzard stratification measurements by mounting cylindrical cans at various heights on a 2-metre mast and allowing snow to drive into their open ends for a measured time.

It was anticipated that drift gauging would be required at Mawson during the IGY, but circumstances prevented any preparatory work in instruments before the expedition sailed in December 1956. During the winter of 1957, new types of drift gauges were developed at Mawson and, soon after midwinter, measurements of the density and vertical distribution of drift snow were begun. Two types of drift gauge were brought back to Australia in March 1958 for testing, and a detailed analysis of the 1957 results was initiated. As a result of the tests and analyses, the drift gauges were further developed and a completely new installation was set up at Wilkes base for operation during 1959.

5.1.2. *Development of new gauges*

After seeing something of Antarctic drift conditions it was soon realized that a single fixed measuring box, giving a running total of transported snow at a given level, was of very limited value, even if the snow were collected efficiently and accurately. Since the amount of snow passing a particular section is determined largely by the drift density distribution, it seemed desirable to develop a battery of small gauges capable of defining the vertical drift profile. Mean density values for the main types of drift could then be used in conjunction with drift reports ("present weather" codes 36, 37, 38, 39) and wind velocities, to arrive at a figure for the mass of snow transported.

The new gauges were designed to remove snow from an air stream by expanding the stream's cross-section and thus reducing its velocity. They were small enough to be detached from their mountings for weighing, so that the mass of snow collected during a measured period could be determined in that way. As impedance to air flow through the gauge had to be kept to a minimum, baffles, gauzes or trapdoors were not acceptable in the absence of calibration facilities. The gauges had to hold steadily into wind and a certain amount of external streamlining was necessary. Another less obvious consideration entered into the design: the gauges had to be built from available oddments by the designer, who had little skill as a sheet metal worker.

A model was improvised from a transparent perspex sugar canister and two polythene photographic funnels. The ends were cut from the canister and its lid, and the funnels were cemented over the holes, giving tapered inlet and outlet to the canister. After tracing the flow pattern by blowing tobacco smoke through the model, the exit diameter was increased and tests were then made in heavy drift. Snow settled out inside the collector quite satisfactorily and examination of the deposit under the microscope showed it to have a grading similar to that of the free drift snow.

The first collector was built to the form shown in Fig. 14A. It consisted of a cylindrical body, or deposition section, fitted with tinplate cones at each end to give tapered intake and exit, and an inlet tube reaching out ahead of the intake cone in order to sample undisturbed air. The collector pivoted on a rod passing through its centre section and into a fibre mounting bush, and vertical fins attached to the exit cone held the nose into wind. The tail cone was soldered to a press-on lid which could be removed from the centre section to give access to the interior. A large inlet tube was provided on this first experimental gauge so that a range of smaller nose tubes could be tested, using a pierced rubber bung as a coupling.

After an exposure of one hour in moderate drift (wind about 25m/sec), using the full inlet diameter of 2·54cm, there was very little snow inside the collector, which was as expected from prior comparisons of calculated settling time and transit time in the collector. With a 1·75cm-diameter nose tube fitted, snow was deposited in a form similar to that of a longitudinal dune. Whilst the bulk of the deposit lay close to the end of the inlet cone, the "dune" was continuous to the exit cone and it was apparent that snow was being carried out in the exhaust stream. A smaller nose tube brought the tail of the "dune" inside the centre section and a

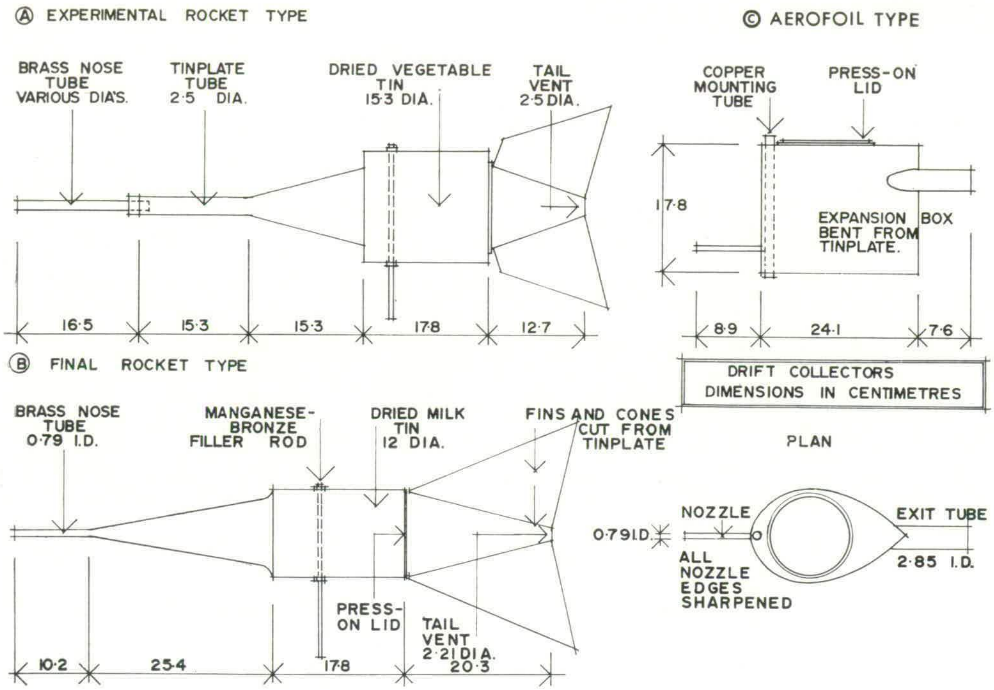


FIG. 14. Drift collectors (A, B and C).

further reduction to 1·11cm diameter led to a tailing-off of the deposit about half-way down the centre-section.

Flow velocities in the gauge are proportional to cross-sectional areas and the principal velocity ratios in the experimental gauge fitted with the 1·11cm-diameter nose tube were as follows:

nose to centre section	188 : 1;
nose to exit	5·1 : 1.

Calculations indicated that this collector could not be expected to effect complete deposition in winds exceeding 35m/sec but, as there was no time for further experiment, the basic form of the gauge was accepted.

The final "rocket"-type collector was built to the dimensions shown in Fig. 14B. This gauge, or precipitator, had cleaner lines than the experimental one, effected a greater velocity reduction, and held the wind very steadily. Its principal velocity ratios were:

nose to centre section	232 : 1;
nose to exit	7·8 : 1.

On theoretical grounds, settlement of the major particle range should be complete for winds up to 40m/sec and examinations of deposits inside the precipitator tend to confirm this prediction. In heavy drift at high wind speeds the precipitator can only be exposed for about half an hour at a height of one metre; if it is left for longer than this the deposited snow will begin to affect its efficiency.

A different type of experimental collector was built in order to measure the rate of flow of drift close to the surface. It consisted of an expansion box with an aerofoil profile and inlet and exit tubes were attached directly to the leading and trailing edges respectively. A press-on lid gave access to the interior and a bearing tube passed vertically through the forward part of the expansion box. A precipitator of this type is detailed in Fig. 14C. The expansion ratios of this type were approximately:

nose to maximum expansion	1 : 390;
nose to exit	1 : 13.

Whilst these at first appear good, the sudden expansions and contractions were unsatisfactory and the length was rather short. It now seems almost certain that the precipitating efficiency of these boxes was inferior to that of the rockets. They did not hold the wind steadily, but tended to yaw. This fault was remedied to some extent by fastening a length of cord to the trailing edge and allowing it to stream in the wind. Used close to the ground the boxes yielded useful data and, although not completely satisfactory themselves, they probably provide the basis for a gauge to give closely spaced measurements near the surface.

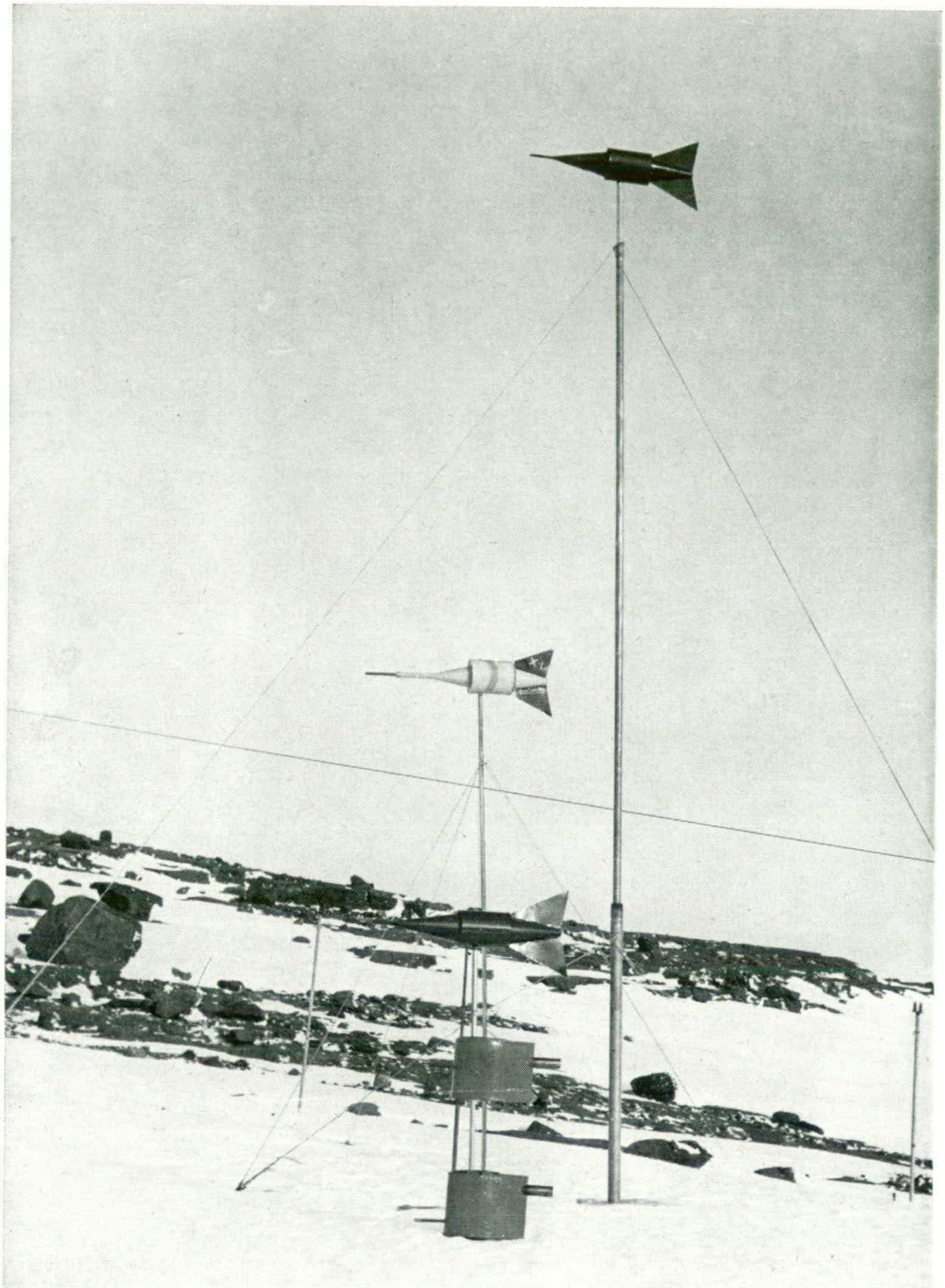
5.1.3. *Using the gauges*

Both the rocket- and aerofoil-type gauges were put into service and mounted so that readings could be made at heights of 4cm, $\frac{1}{2}$m, 2m, and 4m (see Plate 14). The installation was situated on a smooth and gently sloping snow surface in the Mawson camp area within easy reach of the operator's sleeping hut.

When measurements were required the precipitators were weighed in the cold-porch of the hut, taken out to their mountings and left for a suitable period. The exposure time generally varied from 20 minutes to one hour, depending on the density of the drift. At the end of the period the gauges were carefully carried back to the hut and re-weighed. After removing most of the deposit and weighing again they could be replaced on their mountings for a further run (making sure that the gauges were substantially below the freezing temperature before exposing them to drift).

A running record of drift intensity at the 2-metre level was sometimes kept, measuring half-hourly totals by the alternate exposure of two identical rockets. This was done in conjunction with measurements of electrostatic charge on a horizontal antenna at the same level.

On dividing the weight of snow deposited in a gauge by the intake area and the exposure time, the mass flow of snow per unit area and per unit time is obtained. The calibration factor of the gauge is used to adjust this result. However, in order to extrapolate the profiles, it is necessary to know not the rate of flow of the drift but its density at various levels; and to work from the gauge results to a density profile the vertical wind profile must be known. It was possible to process the preliminary results using wind profiles and surface roughness parameters deter-



ANARE photo 7109 M. Mellor

PLATE 14

The drift-gauging installation at Mawson in 1957. It was difficult to handle the 4-metre collector in very strong winds.

mined by other workers, but this is an unsatisfactory treatment and new drift-gauging installations ought to include a small anemometer mast to define the wind conditions of the location.

5.1.4. *Tests on the 1957 precipitators*

A rocket and an aerofoil box were brought back to Australia in March 1958 and wind tunnel tests were carried out at the Commonwealth Aeronautical Research Laboratories in Melbourne by Mr T. N. Pound (1958).

Three static pressure tappings were arranged in the nose tube of each collector, the collector was mounted in the working section of the tunnel and inlet static pressure was recorded against free stream static pressure for a range of wind velocities from 12 to 45m/sec. A hose was later attached to the exit of the collector and air was drawn through at rates regulated to reproduce the previously measured static pressure differences. The flows through the collector were measured by a gas meter and the ratio of inlet velocity to free stream velocity, V_i/V_o, was plotted against V_o to give a flow calibration.

For the rocket type precipitator, V_i/V_o varied from 0·69 at 12m/sec to 0·78 at 45m/sec, and for the aerofoil box V_i/V_o ranged from 1·01 at 12m/sec to 0·95 at 40m/sec with zero yaw angle. As the aerofoil box tended to yaw, tests were run with two fixed yaw angles $\beta = -8\frac{1}{2}°$ and $\beta = -15°$. These gave a less regular variation of V_i/V_o, the range of values being:

$$\beta = -8\tfrac{1}{2}°: \quad V_i/V_o\,(\text{max}) = 1{\cdot}03, \quad V_i/V_o\,(\text{min}) = 0{\cdot}92;$$
$$\beta = -15°: \quad V_i/V_o\,(\text{max}) = 1{\cdot}03, \quad V_i/V_o\,(\text{min}) = 0{\cdot}91.$$

Neither type of precipitator caused much disturbance of the air being sampled when working at zero yaw angle.

These tests gave a flow calibration for the gauges but did not, of course, determine what percentage of the snow passing through the gauges was actually deposited. Rough tests of precipitation efficiency made at Mawson by series running internal inspection indicated that, in heavy drift, air leaving the rockets was almost completely snow-free but deposition in the aerofoil boxes was incomplete.

5.1.5. *The Wilkes installation*

The drift readings made in 1957, though few in number, led to most interesting results which threw light not only on the mass transport of snow but also on the physics of blowing snow. It was therefore decided that improved gauges, together with anemometers, should be installed at Wilkes base after the station had been placed in the custody of Australia by the United States.

Six rocket-type precipitators were built to the dimensions shown in Fig. 15A. They differ from the 1957 rockets only in dimensions and detail, but the changes should result in an improved efficiency. The principal velocity ratios are:

Nose to centre section	314 : 1;
nose to exit	7·5 : 1.

They are pivoted at the ends of cross-arms on a mast which also carries three anemometers to define the wind profile. The sampling heights are 15cm, ¼m, ½m, 1m, 2m, and 4m.

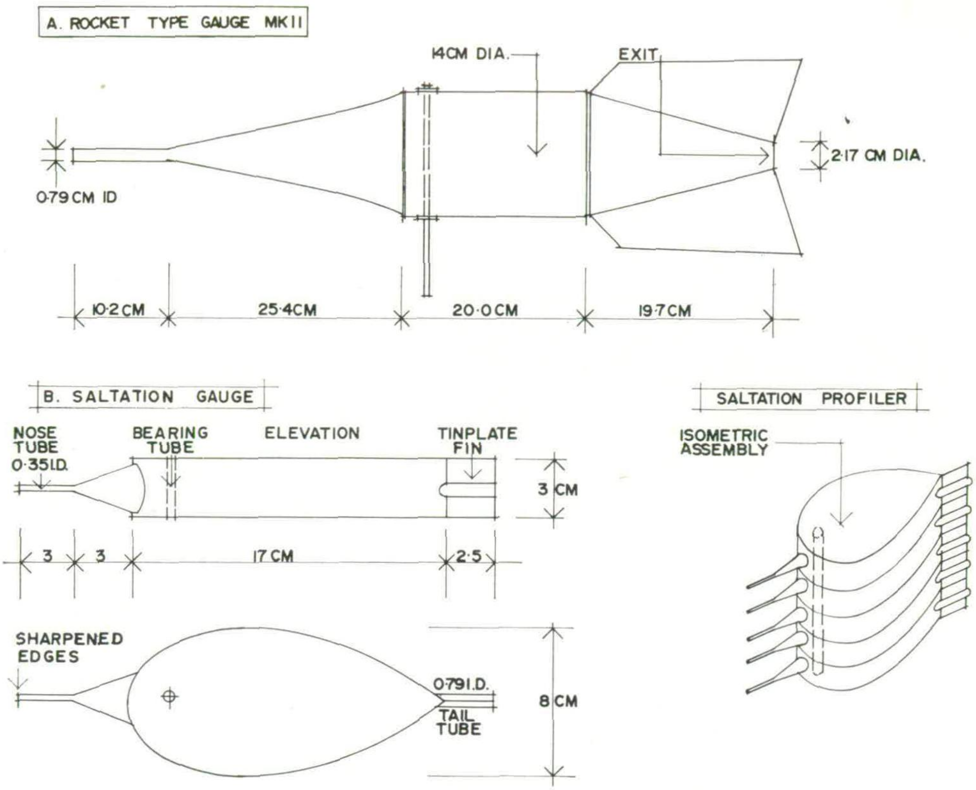


FIG. 15. Drift gauges (A and B).

The 1957 results show two types of snow movement: the blowing of snow suspended by turbulent transfer, and the bouncing of particles along the surface. The turbulent motion was described by Loewe (1956) but the special surface motion has not previously been described for snow. However, this motion by successive impact with, and rebound from, the surface was studied in great detail for desert sand by Bagnold (1941) who named it "saltation". The necessity for density measurements in the saltation layer has now been recognized and in November 1958 a bank of gauges for profiling in this layer was designed and built by the writer.

The saltation gauges are a development of the aerofoil box and the form and dimensions are shown in Fig. 15B. The principal velocity ratios are:

nose to maximum expansion	250 : 1;
nose to exit	5 : 1.

These should suffice, as the wind velocity decreases close to the surface. The five gauges nest together on a single vertical axis (see assembly sketch, Fig. 15) so that readings are made at heights of 1·5, 4·5, 7·5, 10·5 and 13·5cm above the surface. The gauges will probably be given only a short exposure, something like 5 minutes, and after the final weighing the deposit will be cleared by melting it out. A carrying case will be provided to protect the gauges during movement to and from the profiling site.

A flow calibration for the new rocket-type gauges was determined by Mr T. N. Pound, using a method similar to that described above. The ratio V_i/V_o was found to vary smoothly from 0·72 at 12m/sec to a maximum of 0·78 at 37m/sec and down to 0·77 at 46m/sec.

It was not difficult to see how the gauges could be improved, although it was doubtful whether the expense of producing elaborate streamlined traps is justified at this stage. An extension of the vertical profile to 8m would be helpful, but otherwise the present arrangement seems to be satisfactory. It is hoped that drift gauging will now go ahead along with further investigation of wind conditions between the surface and 500m.

Several other expeditions attempted drift gauging during the IGY but no details were available to the author at the time of writing. It is known from private conversations that the gauges built for one expedition were fitted with fabric filters and it is hard to see how accurate results could be obtained with such gauges. The initial impedance would be very high and the variation of the impedance with the amount of snow collected would make it hard to produce an accurate flow calibration.

5.2. SOME PROPERTIES OF DRIFTING SNOW

5.2.1. *Introduction*

The 1957 drift-gauging results were brought back to Melbourne by the author and were analysed with the help of Dr Uwe Radok, Meteorology Department, University of Melbourne. In the following pages the theory of drift snow transport is given, results are discussed and further developments of snow transport theory emerging from the analyses are given.

5.2.2. *Theory of turbulent drift snow transport*

In a detailed study of snow transport, Loewe* (1956) considered stationary conditions in which turbulence transports upwards as much snow as settles down under the influence of gravity. This case is governed by the relation:

$$-wn - \frac{A_z}{\rho} \cdot \frac{\partial n}{\partial z} = 0 \qquad (5.1)$$

where z = height above snow surface (cm)

w = free fall velocity of the snow particles (cm/sec)

n = weight of snow per unit weight of air (so that $n\rho$ = drift snow density)

ρ = density of the air (g/cm³)

A_z = coefficient of exchange (so that A_z/ρ is the "eddy diffusivity" K).

The exchange coefficient (or K) is a function of height and for a given thermal stratification depends only on the vertical wind profile. During blizzards the strati-

* The basic equations governing the turbulent transport of snow particles appear to have been derived independently by Loewe (Études de glaciologie en Terre Adélie 1951-52, Expéditions Polaires Françaises, Paris 1956) and by Shiotani and Arai (A short note on the snow storm. Proceedings of the 2nd Japan National Congress for Applied Mechanics 1952, Science Council of Japan, Tokyo 1953).

fication may be assumed to be neutral, although the admixture of drift snow may create near the ground a layer of heavier "air", quite apart from the cooling effects of radiation and evaporation. However, it was shown by Loewe that even at low levels the drift snow density remains small compared with that of the air (of the order of $5 g/m^3$ as against $1 \cdot 33 \times 10^3 g/m^3$ at the 1m level in heavy drift). The latter can be assumed as constant throughout the boundary layer considered here.

A first approximation to the vertical wind profile under neutral conditions is a power law of the form

$$v_z = v_1 \, z^{\frac{1}{p}} \tag{5.2}$$

from which it follows that the vertical profile of the exchange coefficient A_z is given by the relation

$$A_z = A_1 \; z^{\frac{(p-1)}{}} \tag{5.3}$$

Equations (5.2) and (5.3) are known as Schmidt's "conjugate power law" (Lettau 1939). A rather better approximation to the neutral wind profile is, however, the logarithmic law,

$$v_z = \frac{u_*}{k} \ln \frac{(z+z_0)}{z_0} \tag{5.4}$$

where $u_* =$ "friction velocity" defined by $\rho u^{*2} = \tau$, the shearing stress (constant in the boundary layer)

$z_o =$ roughness parameter

$k =$ v. Karman's constant, $0 \cdot 4$.

For the logarithmic profile, (5.4), the exchange coefficient A_z is a linear function of height (Lettau 1939), viz.,

$$A_z = k\rho u_*(z+z_0) \tag{5.5}$$

where u_* is a function of v at a given height by virtue of (5.4).

The solution of equation (5.1) takes different forms according to whether (5.3) or (5.5) is chosen to represent A_z. Taking first (5.3) with $p = 7$, following Loewe (1956), the drift snow density at height z, $n_z\rho$, becomes

$$n_z\rho = n_0\rho e^{-\frac{7w\rho}{A^1} \cdot z^{\frac{1}{7}}} \tag{5.6}$$

where A_1 represents the exchange coefficient 1cm above the surface and n_o the drift snow content immediately next to the ground. n_o is unknown but can be eliminated, following Loewe, by applying (5.6) to two levels at which n_z is measured. Then

$$\frac{n_a}{n_b} = e^{-\left(\frac{7\rho w}{A_1}\right)\left(z_a^{\frac{1}{7}} - z_b^{\frac{1}{7}}\right)} \tag{5.7}$$

and the unknown parameters w and A_1 are obtained in the form

$$\frac{w}{A_1} = \frac{\ln(n_a/n_b)}{7\rho\left(z_a^{\frac{1}{7}} - z_b^{\frac{1}{7}}\right)} \tag{5.8}$$

Since density observations at two levels fully determine the vertical profile, additional observations can be used to test it. Loewe had at his disposal Barré's (1954) observations made 5cm, 100cm and 150cm above the surface. The first two of these yielded Loewe's profile,

$$\frac{n_z}{n_0} = e^{-2 \cdot 3 z} \tag{5.9}$$

Hence the ratio of the densities at 100 and 150cm would be expected to have the value 1·29. The observed densities were 6·6 and 4·0g/m³, giving a ratio of 1·65 instead, i.e., a more rapid decrease in drift snow density between 1 and 1·5 metres than predicted by theory. However, Barré's drift observations were made by a very primitive method and the discrepancy probably arises from this. The more precise Mawson measurements will be seen to indicate the existence of two separate drift regimes of quite different character, near the surface and higher up respectively.

Equation (5.8) can be applied directly to the Mawson data, since w and A_1 are not required separately for the calculation of density ratio (although A_1 is needed when the calculations are extended to higher layers, cf. Loewe (1956)). In the present case the observed drift densities at 400 and 100cm will be used to compute the exponent 7 w/A_1 while the observations at 200cm (made with the same type of trap as those at 400 and 100cm and at 4cm) will serve as a check on the assumed velocity and exchange coefficient profiles. Although the logarithmic wind profile (5.4) is much superior to the power law, (5.2), the values of the roughness parameter z and of the friction velocity u_* must be known before it can be applied. A great deal of information on these parameters as functions of wind speed under blizzard conditions was published by Liljequist (1957), a member of the Norwegian-British-Swedish Antarctic Expedition. Conditions at Mawson are different from those at Maudheim in that the former lies on rocks at the foot of steep ice slopes while Maudheim was situated on a flat ice shelf. However, the Mawson measurements were made at a time when local snow drifts had built up to a streamlined equilibrium level, and it therefore seemed legitimate to try Liljequist's data in the absence of comparable information for Mawson itself.

With the relation (5.5) for the exchange coefficient, the solution of equation (5.1) takes the form

$$\frac{n_z}{n_{400}} = \left[\frac{(z+z_0)}{(400+z_0)}\right]^{-\frac{w}{ku_*}} \tag{5.10}$$

It should be noted that in this case w is the only unknown parameter; if this is assumed (Loewe's figure of 30cm/sec was used in the design of the drift collectors), all densities are determined in terms of the density at a single level. However, it was considered preferable, in the first place, to compute again the value of the exponent from the ratio of two observed densities for the 400 and 100cm levels. A comparison will then be made, as before, of the observed and computed drift densities for the 200 and 4cm levels. It should be noted that the "observed" densities will all be somewhat different from those resulting for the exponential profile, as the logarithmic wind profile leads to somewhat different wind velocities at collector levels.

5.2.3. *Observations*

Table 5.I gives five sets of observed drift snow amounts and the corresponding densities for the exponential and logarithmic wind profiles which are also included. The basic wind velocity at anemometer level in each case represents a mean value for approximately 30 minutes, obtained by combining the Dines anemometer reading

TABLE 5.I
DRIFT SNOW DENSITIES AND AMOUNTS
Logarithmic wind profiles after Liljequist (1957)

Height cm	Drift amount g/cm²/sec × 10³	Wind speed		Drift density (gm/m³)			
		Exponential m/sec	Logarithmic m/sec	Exponential		Logarithmic	
				Observed	Computed	Observed	Computed
CASE 1: 10 m wind 23 m/sec; Log profile: $z_0 = 0{\cdot}08$ cm, $u_* = 97$ cm/sec							
400	3·03	20·2	20·7	1·50		1·46	
200	3·40	18·3	19·0	1·86	2·10	1·79	1·99
100	4·75	16·6	17·3	2·86		2·74	
50	3·75	15·0	15·6	3·58*		3·44*	
4	14·93	10·5	9·5	17·05*	8·43	18·84*	11·70
CASE 2: 10 m wind 24 m/sec; Log profile: $z_0 = 0{\cdot}11$ cm, $u_* = 104$ cm/sec							
400	7·20	21·0	21·6	3·43		3·33	
200	7·95	19·1	19·8	4·16	4·57	4·02	4·35
100	10·28	17·3	18·0	5·93		5·70	
50	8·00	15·7	18·2	7·30*		7·07*	
4	48·4	10·9	9·7	53·1*	14·8	59·7*	19·6
CASE 3: 10 m wind 25 m/sec; Log profile: $z_0 = 0{\cdot}12$ cm, $u_* = 111$ cm/sec							
400	18·93	21·9	22·5	8·61		8·39	
200	25·0	18·9	20·6	12·55	11·55	12·10	11·13
100	27·4	18·0	18·7	15·20		14·60	
50	22·5	16·3	16·7	19·75*		19·20*	
4	129·8	11·4	9·8	136·6*	39·1	159·0*	51·8
CASE 4: 10 m wind 33 m/sec; Log profile: $z_0 = 0{\cdot}65$ cm, $u_* = 178$ cm/sec							
400	33·3	28·9	28·7	11·50		11·60	
200	35·7	26·3	25·6	13·55	13·55	13·90	13·90
100	37·9	23·8	22·6	15·90		16·70	
50	28·3	21·6	19·5	18·7*		20·4*	
4	159·7	15·2	8·9	126·0*	27·4	215·0*	37·2
CASE 5: 10 m wind 36 m/sec; Log profile: $z_0 = 1{\cdot}1$ cm, $u_* = 210$ cm/sec							
400	26·0	31·6	31·7	8·22		8·20	
200	31·9	28·6	28·1	11·12	10·67	11·32	10·85
100	35·0	25·9	24·4	13·50		14·33	
50	28·7	23·5	20·8	17·4*		19·60*	
4	121·0	16·4	8·5	88·8*	30·7	171·0*	47·5

* 50 cm and 4 cm densities adjusted for collection efficiency with factors 1·43 and 1·20 respectively

(usually too low under blizzard conditions owing to blockages) with those of the station cup anemometer. The observed drift snow amounts in column 2 have been reduced by means of the winds at the collector levels, given in columns 3 and 4, to densities shown in columns 5 and 7, while the computed densities for the 200 and 4cm levels are shown in columns 6 and 8. The observed densities for the 50cm and 4cm levels had to be adjusted, owing to the fact that a different type of trap with lower collection efficiency was used at these levels. The adjustment was based on the fact that the 50cm densities, on the average, were only 70% of their theoretical expectation (individual values ranging from 64 to 72%). The observed 50cm values were therefore raised by a factor of $1/{\cdot}7 = 1{\cdot}43$ and those for the 4cm level (where the wind velocity is roughly half that at 50cm) by a factor of 1·2. In this way the information at the 4cm level was preserved for comparison with theoretical estimates.

The most striking feature of Table 5.I is that, whereas the 200cm values are in reasonable agreement with theory for both wind profiles (which differ little from one another above 100cm) the 4cm values are considerably larger than expected. This suggests that Loewe's calculations may have been vitiated by coupling the 100cm and 5cm levels where in fact different drift mechanisms appear to operate. Visual observations suggest that, near the ground, wind-blown snow grains move on flat trajectories, impacting with and rebounding from the surface. This process was first observed for sand by Bagnold (1936) who described it in detail and gave the name of "saltation" to this mode of progression. Its existence was recognized by Loewe (1956) who for a first estimate assumed that "whereas sand is hopping snow is flying". The present observations, however, raise the possibility that a substantial part of the drift snow may also move by saltation. This makes it necessary to consider in the following, separately, what will be termed the "diffusion range", represented by the observations between 100 and 400cm where the snow transport is by turbulence, and the "saltation range" for which only the 4cm observations are available.

Table 5.I shows also that the drift density is not determined uniquely by the wind velocity. It is in fact not uncommon for drift to cease completely while the wind continues without any appreciable change in velocity. Fig. 16 shows an example of changes in drift density at a single level together with the wind velocities during a blizzard observed at Mawson on 11 and 12 September 1957. It is thus

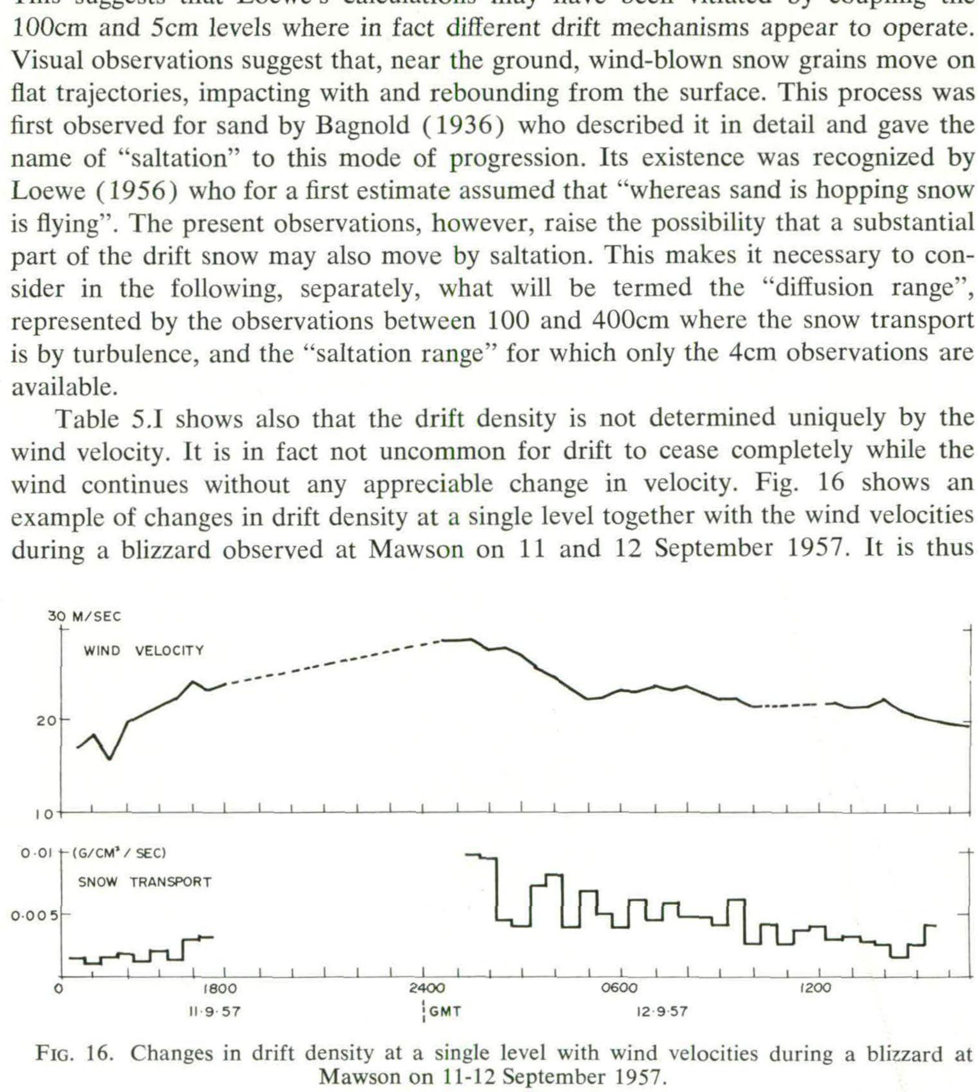


FIG. 16. Changes in drift density at a single level with wind velocities during a blizzard at Mawson on 11-12 September 1957.

clear that observed density profiles may be affected by changes in the degree of "saturation" (Liljequist 1957).

The observed drift snow densities at the 400 and 100cm levels, so far used only to determine the theoretical density profiles, can also serve for a check on the validity in the present case of Liljequist's (1957) friction velocities, or alternatively for estimates of the free fall velocity w of the snow particles. Neglecting the roughness parameter z_o, equation 5.10 for the two levels reads

$$\frac{n_{100}}{n_{400}} \approx \left(\frac{1}{4}\right)^{-\frac{w}{ku_*}} \tag{5.11}$$

Hence the ratio of the friction and free fall velocities is

$$\frac{u_*}{w} = k^{-1}\frac{\log 4}{\log\left(\frac{n_{100}}{n_{400}}\right)} = \frac{1 \cdot 51}{\log\left(\frac{n_{100}}{n_{400}}\right)} \tag{5.12}$$

Assuming, as before, $w = 30$cm/sec, the values of u_* in the five cases of Table 5.I are 164, 188, 183, 322 and 209cm/sec. These differ systematically from the corresponding values of Liljequist used for the wind profiles, 97, 104, 111, 178, and 210cm/sec. An alternative interpretation would be that w varied from case to case. When the Liljequist values of u_* are substituted in (5.12), they yield for w the value 18cm/sec in the first 4 cases and 30cm/sec in the last. The first value seems rather low† and hence there remains some suggestion that the recomputed friction velocity, or in other words the density ratio n_{100}/n_{400}, did not increase with wind velocity as required by the theory. This amounts to further evidence to the effect that the saturated drift conditions implied by equation (5.1) may often not be realized.

5.3. DRIFT SNOW TRANSPORT IN THE DIFFUSION RANGE

For the purpose of a revised estimate of the total drift snow transport by turbulence, the average of the five observed density ratios n_{100}/n_{400}, will be used to compute an average value for the exponent $-w/ku_*$, viz., 0·381. With $w = 30$cm/sec this corresponds to a friction velocity of 197cm/sec. On the other hand, for the mean wind associated with heavy drift at Mawson, 28m/sec (which is also not far from the average of the anemometer winds in the five cases of Table 5.I) Liljequist (1957) gives $u = 129$cm/sec only. To bring the exponent in line with this value of u_* the free fall velocity has to be reduced to 20cm/sec.

For the transport estimate a logarithmic wind profile with Liljequist's values for 27·5m/sec will be assumed, viz., $z_o = 0 \cdot 2$cm, $u^* = 129$cm/sec. This leads to a value of 820 for the exchange coefficient at 120 metres. Above that level a linear decrease in A_z will be assumed, viz.,

$$A_z = 820 - 0 \cdot 03\ (z - 12{,}000) \tag{5.13}$$

By comparison, Loewe (1956) placed his maximum exchange coefficient (600, for a wind of 35m/sec at anemometer level) at 150 metres and assumes a slower rate of decrease higher up (0·02). In both cases the exchange coefficients become

† Later measurements at Wilkes, in February 1959, have shown free fall velocities under summer conditions ranging from 30 cm/sec to 100 cm/sec.

negligible just above 400 metres. The level of the maximum A_z and all details of the decrease higher up are conjectural, since the available information of this kind (cf. e.g., Rossby and Montgomery (1935)) seems inapplicable to the present case of a pronounced gravity wind in a shallow surface layer. Observation suggests, however, that dense snow drift can extend to 300 metres so that Loewe's assumptions (on which the present ones have been modelled) are very plausible.

The detailed snow transport calculations based on the present data are given later. It will suffice here to compare the estimated densities and snow transports with Loewe's pioneer estimates. Fig. 17 shows the two density profiles. It should be

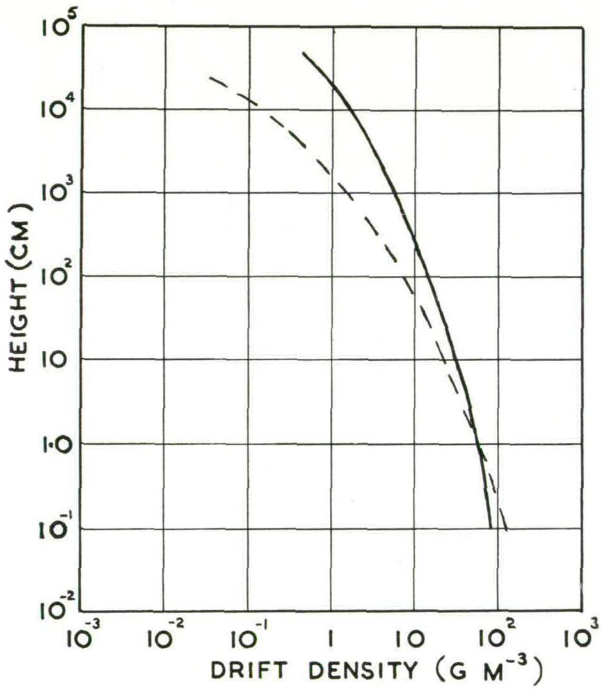


FIG. 17. Average drift snow density as a function of height.
——————— present estimate
— — — — Loewe 1956

noted that Loewe's values refer to a stronger wind at anemometer level. Nevertheless, except below 1cm, the present estimates are throughout higher, the difference becoming more accentuated with increasing height. The transport estimates are set out in Table 5.II where especially the total snow contents for the different layers should be noted. The total turbulent snow transport estimate based on the present data is 153g/cm/sec, as compared with Loewe's figure of 28g/cm/sec for a stronger wind.

The implications of the revised estimate can only be touched on briefly here; a full discussion will be given later. Loewe's estimate implied, for the extra-ordinary conditions characterizing the coast of Adélie Land, an annual drift snow transport

TABLE 5.II

SNOW TRANSPORT ESTIMATES FOR DIFFUSION RANGE

Height range *m*	Loewe (1956)			Mellor and Radok		
	Total drift content $g/m^3 \times 10^3$	Mean wind m/sec	Transport g/cm sec g	Total drift content $g/m^3 \times 10^3$	Mean wind m/sec	Transport g/cm sec
0– 10	3·5	35	12	7·5	25	19
10–150 10–120	3·5	38	13·5	27·8	35	92
150–300 120–300	0·8	30	2·5	19·4	20·5	40
300–430				1·7	13	2
Total	7·8		28	56·4		153

of at least 18×10^9kg, probably 35 to 40×10^9kg through each kilometre of coast line (assumed at right angles to the katabatic wind). The revised estimate gives a figure of about 15×10^9kg a year for the cases of code 38 and 39 drift alone at Mawson and hence presumably at most other points of the fringe region close to the ice cap. This figure has been corrected for the inclination of the katabatic wind to the coast line (assumed as 45°) and moreover does not yet include the drift snow transported by saltation. Its magnitude throws a completely new light on the problem of the Antarctic mass balance and clearly is in urgent need of further confirmation. This was expected to come from a much more extensive series of drift measurements at Wilkes during 1959 with traps of the type described previously and anemometers at severals levels. Meanwhile, however, it remains to complete the present transport estimate by considering the processes in the saltation layer.

5.4. PROCESSES AND MASS TRANSPORT IN THE SALTATION LAYER

In the layer of heaviest drift near the surface processes and conditions must differ radically from those of the diffusion range. No study comparable to Bagnold's (1941) for sand appears to have been made of snow drifting in this layer, and the difficulties facing such a study would be even greater than those Bagnold had to overcome. Yet the contribution made by saltation to the total mass transport in drifting snow cannot be discounted a priori, since experience shows that saltation is active in some form already at wind velocity below those required for appreciable turbulent drift (estimated by Liljequist (1957) as at least 13m/sec at the 10-metre level).

The question thus arises as to whether the results available for sand saltation could be modified so as to provide at least a rough estimate of the total mass involved in snow saltation at a given wind speed; for an answer we consider now briefly Bagnold's approach to the problem.

From observations both in the desert and in a wind tunnel, Bagnold found that sand grains advance in series of leaps on trajectories which, by and large, rise vertically into the air and return to the surface under a very shallow angle. Thus in first approximation the grain begins its leap without forward momentum and returns to the surface with momentum qu (q = mass of grain, u = final horizontal velocity component) which has been extracted from the air mainly near the top of the path where most time is spent. The momentum loss is equivalent to additional friction and appears in the vertical wind profile for drift conditions as an increase in both the roughness parameter z_o and the shear velocity u_* (Bagnold's terminology is slightly different). The total momentum per unit length in a lane of unit width extracted from the air by all the sand grains saltating in the lane equals the additional surface stress over and above that due to turbulent friction, so that

$$\tau' - \tau = \Sigma \frac{qu}{l} \tag{5.14}$$

where l is the length of the jump made by the individual grain and the prime throughout refers to drift conditions. Equation (5.14) assumes that the entire forward momentum of each grain is lost on impact; the grain thus either ejects another grain with an initial vertical velocity w or else rebounds itself vertically from the surface for another leap. It should be noted that this assumption leads to minimum estimates for the mass transport for a given surface stress increase $\tau' - \tau$. By definition

$$\tau' = \rho u_*'^2, \qquad \tau = \rho u_*^2$$

Bagnold examined a great number of individual particle trajectories and found a relation, valid for a wide range of grain sizes and winds, between the ratio u/l in (5.14) and the initial upward velocity w, viz.,

$$\frac{u}{l} = \frac{g}{w} \tag{5.15}$$

where g is the acceleration of gravity. With this, the additional surface stress due to sand drift becomes

$$\tau' - \tau = \Sigma \frac{qg}{w} \tag{5.16}$$

Now the crucial step which makes possible an estimate of the total mass of saltating sand is the replacement of the sum in (5.16) by the product of the total mass Q and an average vertical velocity W valid for the "characteristic" path of the sand grains for the wind profile in question. Bagnold was able to do this because two features of such a characteristic path showed up clearly in his experimental results: the estimated characteristic path length L closely agreed with the wave-length of observed sand ripples, and a fixed fraction of the maximum height reached by the characteristic trajectory was found to be marked by a distortion of the logarithmic wind profile (suggesting maximum momentum transfer at that level). These observations confirmed that the characteristic path had physical reality and made it possible to write (5.16) as

$$\tau' - \tau = \frac{Qg}{W} \tag{5.17}$$

For the characteristic ejection velocity W, Bagnold postulated on somewhat arbitrary grounds a relation of the form

$$W = Bu'_* \tag{5.18}$$

where u'_* is the friction velocity observed under drift conditions and B a constant "impact coefficient", with the value 0·8 for the sand most closely studied. Substituting from (5.18) in (5.17) and neglecting τ against τ', so that $\tau' - \tau = \rho u'^2_*$ finally yielded for the total mass Q the relation

$$Q = \left(\frac{B\rho}{g}\right) u'^3_* \tag{5.19}$$

Bagnold's observations showed that, in place of the shear velocity u'_*, the difference between the wind velocity at a standard level and a "threshold" velocity v_t could be used†, and that a further factor $(d/D)^{\frac{1}{2}}$ was needed to allow for grain diameters d different from his standard diameter $D = 0{\cdot}25$cm, apart from an empirical factor C which made the computed quantities Q equal to those observed. C was found to vary from 1·9 for saltation over deep sand (where as a rule the impact of a grain ejects another grain) to 4·4 for saltation over a hard bare surface (where rebounding predominates). The final expression for the weight of the sand passing a fixed point in a lane of unit width in unit time becomes thus

$$Q = 1{\cdot}5 \times 10^{-6} \sqrt{\frac{d}{D}}\,(v - v_t)^3 \text{g/cm/sec} \qquad \text{(deep sand)}$$

$$(5.20)$$

$$Q = 2{\cdot}9 \times 10^{-6} \sqrt{\frac{d}{D}}\,(v - v_t)^3 \text{g/cm/sec} \qquad \text{(hard surface)}$$

where v = wind velocity at 1m.

From the brief survey just given it can be seen that, in order to apply Bagnold's concepts to snow drift, a relation similar to (5.15) between the saltation path characteristics and the initial velocity w of snow grains would have to be established. Furthermore, the impact coefficient for snow or some equivalent relation between w and the wind profile parameters must be found.

A relation corresponding to (5.15) was obtained from two snow grain trajectories constructed by means of the elaborate graphical technique described by Bagnold (1936), using Antarctic conditions as regards air density and viscosity. Evidently, a far greater number of trajectories would be needed to determine the relation with any sort of accuracy; these are being prepared by numerical integration of the equations for the motion of a snow particle under the influence of gravity and turbulent friction, viz. (cf. Fig. 18),

$$\dot{u}' = (u' - U')\, v\, \frac{\rho A}{2g}\, c(v) \tag{5.21}$$

$$\dot{w}' = -g + w'\, v\, \frac{\rho A}{2g}\, c(v)$$

$$v = [(u' - U')^2 + w'^2]^{\frac{1}{2}}$$

† It can be shown on purely geometrical grounds that, in Bagnold's case, the additional shearing stress due to drift is strictly proportional to $(V_z - V_t)^2$, so that there is no real need to disregard τ.

where u' and U' are the horizontal particle and wind velocities respectively and $c = c(v)$ is the resistance coefficient for a sphere equivalent to the snow grain (cross section area A, mass q). The calculations at the time of writing were being made on the electronic computer CSIRAC at the University of Melbourne†. For the

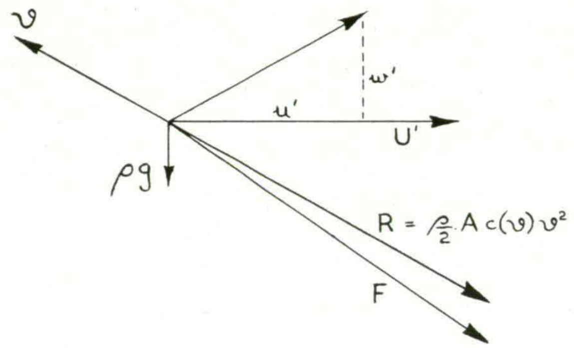


FIG. 18. Forces acting on a snow particle.

present purpose, however, it must suffice that the two graphical trajectories suggest for the relation between u/l on the one hand and w on the other, in the case of typical blizzard snow grains,

$$\frac{u}{l} = 0\cdot65\ \frac{g}{w} \tag{5.22}$$

The question of the relation between the initial velocity w and the friction velocity u'_* is more difficult and seemed at first intractable without observations equivalent to Bagnold's own. However, a physical interpretation of the impact coefficient B may provide a solution. The relation (5.18) viz., $w = Bu'_*$, evidently has the form of the logarithmic wind profile, if B is given the (constant) value $5\cdot75 \log (z + z_o/z_o)$ and w then represents the wind at height $z = z_o(10B/5\cdot75 - 1)$. Thus Bagnold's value of $B = 0\cdot8$ implies that w equals the horizontal wind at height $0\cdot377\ z_o$.

While this is not directly useful, it suggests a modified definition for the impact coefficient in terms of the final impact velocity u. Writing $w = bu = Bu'_*$ makes the factor $B/b = 5\cdot75 \log (z_1 + z_o/z_o)$ a measure of the height z_1 at which the

TABLE 5.III
WIND PROFILE AND TRAJECTORY PARAMETERS FOR SAND
(after Bagnold (1937))

Friction velocity u'_* cm/sec	25	40·4	50·5	62·5	88
Initial velocity $W = 0\cdot8\, u'$	20	32·3	40·3	50	70
Final velocity U	147	164	194	227	335
W/U	0·136	0·197	0·208	0·220	0·210

† This work was to be carried out by Mr D. Jenssen of the Meteorology Department, Melbourne University.

horizontal wind velocity equals that of the final impact. This line of argument is confirmed by the fact that, for moderate and strong winds, Bagnold's W values can be expressed as a quasi-constant fraction of the final velocity U equally well as by (5.18). This is demonstrated in Table 5.III:

We still require information regarding the impact efficiencies of sand and snow, In the absence of such information, it can be argued that using the same ratio W/U for sand as for snow (say 0·21) leads to very low maximum heights for the characteristic snow trajectories, in contrast with experience. It seems then, more likely that the ejection or rebounding of snow particles is more efficient than the same process in the case of the harder quartz. Such a trend is in fact apparent from the relative impact efficiences of different materials (cf., e.g., Hütte 1955, p. 733). It will therefore be assumed for the following that the initial velocities during snow saltation are of the order of 50% of the final impact velocities, or

$$B = 0{\cdot}5 \frac{U}{u'_*} \approx l \quad \text{or} \quad W = u'_* \tag{5.23}$$

with the values of U and u_* (88cm/sec) for the two snow trajectories mentioned earlier†.

The snow transport by saltation can now be estimated, using directly the values of the surface stress due to snow drift given by Liljequist (1957). The mass passing a fixed point in a lane one centimetre wide in unit time is

$$Q = (\tau' - \tau) \frac{l}{u} \tag{5.24}$$

and with (22) and (23) this becomes, since $u'^2_* = (\tau' - \tau)/\rho$,

$$Q = \frac{(\tau' - \tau)^{\frac{3}{2}}}{0{\cdot}65\, g\rho^{\frac{1}{2}}} \tag{5.25}$$

Finally, introducing Bagnold's proportionality factor C for saltation over bare ground, 4·4, and multiplication by g to obtain weights instead of masses,

$$Q = 0{\cdot}18\, (\tau' - \tau)^{\frac{3}{2}} \text{g/cm/sec}$$
$$\text{with } (\tau' - \tau) \text{ in dynes/cm}^2.$$

Table 5.IV shows the saltation transport figures as a function of wind velocity at the 10m level. The weight of drift snow saltating through each centimetre at right angles to the katabatic wind approximately doubles for each 2·5metre/sec increase in wind velocity above 20m/sec.

Since the additional stress, due to drift, increases with the 5th power of the wind velocity (Liljequist 1957), the weight of the snow moved by saltation increases with $(v - 13)^{7 \cdot 5}$.

† Later observations (cf. footnote, p. 70) suggest that (5.23) should read $W = cu'_*$ when c may be as large as 3 or 4. In a 6 m/sec wind the observed maximum saltation height was 13 cm.

TABLE 5.IV

WEIGHT OF DRIFT SNOW TRANSPORTED BY SALTATION
(based on winds and stresses given by Liljequist (1957))

Velocity at 10m m/sec	15	17·5	20·	22·5	25	27·5	30	32·5
$\tau' - \tau$ dynes/cm^2	0·3	0·9	2·1	3·7	6·5	10·1	16·0	24·0
Q g/cm sec	0·03	0·12	0·55	1·28	3·0	5·8	11·6	21·1

This rate is in marked contrast with Bagnold's result for sand, viz. $(v - v_t)^3$, and it seems highly desirable to have some experimental check on the figures of Table 5.IV. Pending measurements at Wilkes with the "saltation profiler" during 1959, the drift snow amounts collected at the 4cm-level at Mawson and listed in Table 5.I can be used with due caution. After deducting the expected snow transport by turbulence at the 4cm-level amounts of the order of 0·05 to 0·1g/cm/sec remain to be accounted for by saltation. The sampler orifice had a diameter of 0·79cm, so that approximately 0·1 to 0·2g/cm/sec would have passed through a square with 1cm sides with its centre at the 4cm-level.

Little is known about the vertical density profile during saltation since the readily observed features only depend on the total mass involved. However, Bagnold (1936) has given a typical distribution curve for sand which shows the broad features of the density profile, especially the rapid decrease in density with altitude. If this is combined with the logarithmic wind profile, a bottom layer of fairly uniform transport results; above this the weight of drift snow per centimetre layer decreases very rapidly as the vertical rate of increase in wind velocity becomes small.

For snow it must be expected that the same would apply with a reduction in the vertical scale. For a mean wind at the 10-metre level of 21m/sec the "characteristic" snow trajectory reaches a height of 0·5cm only, as compared with 2·0cm for sand. Bagnold's example corresponds to a mean height of just under 1cm which may be roughly applicable to the cases in Table 5.I. Now, a graphical integration of Bagnold's distribution curve shows that a collector placed at the 4cm level would have yielded approximately 5% of the total sand transport; for snow the corresponding figure would be less, say 2% only. Since the total has been estimated for a 30m/sec wind as around 10g/cm/sec, a 4cm transport of 0·1 to 0·2g/cm/sec appears of the right order of magnitude.

However, the different powers of the velocity entering into the transport equations for sand and snow saltation respectively require confirmation. In the present treatment the 5th power law governing the increase of the additional shearing stress has been ascribed entirely to an increase in the mass transported. It cannot be ruled out, however, that the relation (5.22) would undergo a change for greater velocities than that for which it was established (21m/sec at the 10-metre level: this was the greatest value for which Bagnold has given comparative figures). It is hoped that the trajectory calculations referred to earlier will clear up this point.

5.5. CONCLUSION

The last section has provided an estimate for the saltation transport which, if anything, is less certain than that for the turbulent transport. Nevertheless, the order of magnitude can perhaps be trusted. It is then seen that, except for very strong winds (and even then, only provided the $(v - 13)^{7 \cdot 5}$ law is correct), the transport by turbulence, as established by the Mawson measurements, is an order of magnitude larger than that by saltation, and the total transport figure is not substantially increased by including the saltation contribution. For all that, it retains considerable interest since it defines a layer in which turbulent transport theory may not be applied. Furthermore, it seems possible that saltation may have something to do with the peculiar electrical phenomena described by Barré (1954).

5.6. ESTIMATION OF MASS TRANSPORT

In the previous sections details of wind transport mechanisms have been given and results have been interpreted. Those results are now used to estimate the mass of snow carried across the coastline by the wind each year.

For the turbulent diffusion range, vertical distributions of the exchange coefficient A_z and the drift density n_ρ will be estimated. These will be compared with corresponding values used by Loewe (1956) and then the drift densities will be integrated to give the masses of snow in various layers. Three main layers will be treated: the surface boundary layer, 0-10m, a layer of increasing exchange coefficient, from 10-120m, and a layer of decreasing exchange coefficient, from 120-390m. The snow quantity in each layer will be multiplied by a suitable mean wind velocity for the layer to give a rate of mass transport. Finally, typical rates of flow will be used with drift reports to give an estimate of the annual export.

Transport by saltation will be estimated from figures given in the previous section together with drift reports.

The mean wind speed at the 10m level assumed for the transport estimates is 28m/sec. For a wind of 27·5m/sec, Liljequist gives the parameters

$$z_o = 0 \cdot 2\text{cm}; \qquad u_* = 129\text{cm/sec}.$$

Assuming a logarithmic wind profile, the exchange coefficient up to the height of the maximum value is given by equation (5.5),

$$A_z = k \rho u_* \ (z + z_o),$$

which becomes, after substitution of constants and parameters,

$$A_z = 6 \cdot 86 \times 10^{-2} \ (z + 0 \cdot 2).$$

In the absence of data on katabatic winls, the height at which the exchange coefficient is a maximum was taken as 120m, and above that level the exchange coefficient was assumed to decrease linearly according to equation (5.13):

$$A_z = A_{12,000} - 0 \cdot 03 \ (z - 12,000)$$

or, $$A_z = 1,180 - 0 \cdot 03z.$$

For the range in which the logarithmic wind profile holds, the drift density can be determined from equation (5.10):

$$\frac{n_z}{n_{400}} = \left[\frac{(z + z_o)}{(400 + z_o)} \right]^{-\frac{w}{ku_*}}$$

For this relation a mean value of the exponent has been determined from observations to be 0·381 and ρn_{400}, the drift density at 400cm, was found to have a mean value of $6 \cdot 6 \times 10^{-6}$g/cm³. Therefore,

$$\rho n_z = 6 \cdot 6 \times 10^{-6} \left[\frac{400 \cdot 2}{z + 0 \cdot 2}\right]^{0.381} .$$

For the levels above 120m the solution of equation (5.1) becomes

$$n_z = n_{12,000} \left[1 + \frac{(z - 12,000)}{A_{12,000}} \frac{dA}{dz}\right]^{-\frac{\omega\rho}{dA/dz}} ;$$

so that, after substitution,

$$\rho n_z = 1 \cdot 8 \times 10^{-6} \left[\frac{460 + 0 \cdot 03\, z}{820}\right]^{-1.33} .$$

Table 5.V compares the vertical distributions of exchange coefficient and drift density found by the method described above with corresponding figures given by Loewe.

TABLE 5.V

Height z (cm)	Mellor		Loewe	
	Exchange coefficient A_z	Drift density $n\rho$ (gm/m³)	Exchange coefficient A_z	Drift density $n\rho$ (gm/m³)
1	0·08	60	0·12	—
10	0·7	27	0·86	—
50	3·4	15	3·4	—
100	7	11	7	6·6
200	14	8·6	11	—
400	27	6·6	—	—
500	34	6·1	25	—
1,000	69	4·6	45	1·2
2,000	140	3·6	85	0·64
5,000	340	2·5	205	0·26
10,000	690	1·9	400	0·13
12,000	820	1·8	—	—
15,000	730	1·6	600	0·09
20,000	580	1·3	—	0·056
30,000	280	0·9	300	0·022
40,000	—	—	100	—

We now consider the snow contained in a vertical column of unit cross-section. The amount of snow in the column is calculated by integration of the drift snow densities given in Table 5.V. Surface boundary layer, 0-10m. The quantity of snow contained in a vertical column of unit cross-section is given by

$$\begin{aligned} N_{0-1000} &= \int_0^{1000} \rho n_z \, dz \\ &= \frac{6 \cdot 6 \times 10^{-6} \times 9 \cdot 77}{1 - 0 \cdot 381} \left[(z + 0 \cdot 2)^{1-0.381} \right]^{1000} \\ &= 7 \cdot 5 \times 10^{-3}\text{g/cm}^2 \end{aligned}$$

Layer from 10-120m. The mass of snow in the column is

$$N_{1000-12000} = \int_{1000}^{12000} \rho n_z \, dz$$

$$= 1{\cdot}04 \times 10^{-4} \left[z^{0.619} \right]_{1000}^{12000}$$

$$= 27{\cdot}8 \times 10^{-3} \text{g/cm}^2 \quad .$$

Layer from 120-390m. This time the expression for n_z is of a different form.

$$N_{12000-39000} = \int_{12000}^{39000} \rho n_z \, dz$$

$$= \frac{n_{12000} A_{12000}}{dA/dz - w\rho} \left[\left(1 + \frac{(z - 12000)}{A_{12000}} \frac{dA}{dz} \right)^{1 - \frac{w\rho}{dA/dz}} \right]_{12000}^{39000}$$

$$= 21{\cdot}1 \times 10^{-3} \text{g/cm}^2 \quad .$$

If the snow quantities for these layers are now multiplied by an effective mean wind velocity for each layer, an estimate of the rate of mass flow results. A more satisfactory method of making the estimate would be to integrate the product $\rho n_z \, u_z$, but with the present lack of wind data this more rigorous treatment is hardly justified. The mass flow through each layer is given below.

Layer from 0-10m.

$$\begin{aligned} N_z &= 7{\cdot}5 \times 10^{-3} \text{g/cm}^2 \\ u_m &= 2{,}500 \text{cm/sec} \\ \text{mass flow, } M &= N_z u_m = 19 \text{g/cm/sec} \end{aligned}$$

Layer from 10-120m.

$$\begin{aligned} N_z &= 27{\cdot}8 \times 10^{-3} \text{g/cm}^2 \\ u_m &= 3{,}300 \text{cm/sec} \\ \text{mass flow, } M &= N_z u_m = 92 \text{g/cm/sec} \end{aligned}$$

Layer from 120-390m.

$$\begin{aligned} N_z &= 21{\cdot}1 \times 10^{-3} \text{g/cm}^{-2} \\ u_m &= 2{,}000 \text{cm/sec} \\ \text{mass flow, } M &= N_z u_m = 42 \text{g/cm/sec} \end{aligned}$$

This gives a total of 153g/cm/sec along the wind direction for a mean wind of 28m/sec at 10m height, which may be taken as representative of drift, which is moderate to heavy above eye level. The table below gives a 4-year average of drift frequency at Mawson. Drift reports may vary considerably with the observer, but averaged over several years and a number of independent observers they should give a reasonably balanced picture.

Weather code	36	37	38	39
No. of hours	481	58	246	321

The drift weather code represents:

36—light to moderate below eye level;
37—heavy below eye level;
38—light to moderate above eye level;
39—heavy above eye level.

To obtain the mass transport, we multiply the rate of mass flow by the total hours in code 39 plus one-quarter of the total hours in code 38, since this last group covers light to moderate drift.

Saltation rates were estimated in the previous section and rates matching the mean winds of codes 36, 37, 38, and 39 have been multiplied by the total drift hours of those groups to give the annual mass transport by saltation.

The final estimates of annual mass transport along the wind direction are:

Turbulent transport	$0{\cdot}21 \times 10^{14}$g/km yr
Saltation	$0{\cdot}012$,, ,,
Total transport in wind direction	$0{\cdot}22 \times 10^{14}$g/km yr

On the open snowfields south of Mawson at latitude 68°S, the blizzard dunes have a bearing of 135° true and the main sastrugi direction is 150° true. We have no information on changes of wind direction with altitude and, for the purpose of this estimate, the mass transport in the wind direction can be divided by $\sqrt{2}$ to give the meridional transport (i.e., we assume a mean wind direction of 135°). In this way the annual meridional transport becomes $0{\cdot}16 \times 10^{14}$g/km/yr, which is close to Loewe's minimum estimate for Terre Adélie.

Mawson has an average wind velocity of 10·5m/sec and drift of one kind or another prevails for about 14% of the time. At Port Martin in Terre Adélie the mean wind speed is 19m/sec and the frequency of all drift is about 50%. Figures presented by Loewe (1957) show Mawson to have a lower drift frequency than several other Antarctic stations, although the mean winds of these stations are lower than that of Mawson. Though information is still lacking, it seems that drift transport at Mawson is not exceptionally high, so that the figure given above may be applicable to a considerable length of the Antarctic coastline.

6. OCEANIC MELTING

With a scarcity of data on bottom-melting beneath ice shelves, there is consequently a diversity of opinion on its importance to the mass balance of Antarctica. In a survey of Antarctic problems, Wexler (1958) assigned an enormous significance to oceanic melting, assuming that four times as much ice is lost by this means as by all other loss processes. Wright (1925) has also stated that there is considerable melting beneath the Ross Ice Shelf but, on the other hand, Debenham (1949) believes that some ice shelves *gain* mass by bottom accretion and Sir George Simpson, during the discussion of Professor Debenham's paper, supported the bottom accretion theory on different grounds. However, core-drilling at Little America V indicates that there is no basal accretion beneath the Ross Ice Shelf at that point ("Antarctic", March 1959). In a recent report, Swithinbank (1958) examines the problem in the light of various studies made at Maudheim and, although his two initial estimates vary by a factor of 5, it seems that the annual loss may be about 20cm of water at the seaward extremity of the ice shelf. Swithinbank also points out that the melting will decrease with distance from the ice front. From consideration of the chemistry of the process it seems that it is possible for considerable bottom-melting to take place, but so far there have been few observational data to support the various theories. The present writer feels that large balancing estimates of bottom-melting have been made too glibly in certain mass balance presentations for, as Loewe (1960) points out, melting beneath ice shelves can only remove ice which actually passes into the shelves, and this is certainly less than half of the total. If equilibrium of the ice shelf is to be maintained, assumptions of very rapid oceanic melting lead to a necessity for rates of surface accumulation and inflow of continental ice which seem unrealistic.

Wexler (1958) used the heat transport calculations of Sverdrup (1953) to deduce a rate of bottom-melting for the whole of Antarctica. By means of a questionable extrapolation of Sverdrup's results, he arrives at a value of $1{\cdot}30 \times 10^{18}$g/yr for the total ice loss. In making this estimate Wexler assumes an ice front of 14,000km length which is almost twice the frontage of ice shelf appearing on modern maps of Antarctica.

There is surprisingly little information available on this important problem, although it seems likely that the intensive investigations made on the Ross Ice Shelf by US expeditions in recent years will have shed some new light on the matter. The most recent estimate to become available is that of Lister (1959) who suggests that the total ice loss by oceanic melting is about $0{\cdot}7 \times 10^{18}$g/yr.

For the purpose of a working estimate, we might take an annual loss of 20cm of water, after Swithinbank, and suppose that melting at this rate occurs for an effective distance of 30km from the ice front. This gives a loss of $0{\cdot}06 \times 10^{14}$ g/km/yr and, if a total ice front length of 8,000km is accepted, the annual loss for the whole of Antarctica becomes only $0{\cdot}05 \times 10^{18}$g/yr. The wide discrepancy between this and the above estimates shows the urgent need for more data.

7. VARIATIONS OF THE ICE MARGINS

7.1. INTRODUCTION

In the preceding sections, measurements and estimates of the rates of gain and loss of ice have been given. The rates of gain and loss will be compared in order to form an opinion on whether the ice cap is suffering a change of total ice mass or whether it is in equilibrium. However, to gain additional direct information on past changes in the size of the ice cap, mountains and smaller rock exposures were examined for evidence of submergence by ice. In this way it was hoped to determine whether the ice cap has been growing or shrinking in the past 50 to 100 years and also to get some idea of how long-past climatic changes affected the ice cap.

Since the last century, glaciers over most of the earth have been shrinking and, during the past century, the rate of recession has generally increased. A mass of evidence of this trend has been collected by Ahlmann in a comprehensive survey of glacier variations (Ahlmann 1953), while other writers, too numerous to mention, have brought forward proof of recent glacier recessions in Europe, North and South America, Africa, Asia, New Zealand, and the Arctic regions. Since about 1940-50, however, there are signs that the recession has become less general in the northern Hemisphere.

Glacial recession has been general throughout the Arctic and sub-Arctic, and numerous examples have been reported from Alaska, Canada, Iceland, Jan Mayen, Spitsbergen, Franz Joseph Land, Novaya Zemlya, Norway, and the outflow glaciers of the Greenland ice cap. It would seem reasonable, then, to consider the possibility of some corresponding process in Antarctica and, indeed, some reports (Warner 1945; Knowles 1945) give the impression that ice recession was occurring in West Antarctica in 1940. Further reports of recent glacier retreat at Heard Island have been made by several members of the ANARE.

Against this, however, Schytt (1953) was able to state that no appreciable thinning of the ice had occurred in Dronning Maud Land for many years prior to 1950.

In these circumstances the writer approached the question of ice margin variations in East Antarctica with an open mind. The studies in this field were divided into two broad categories: an attempt to show whether or not there had been fluctuations of the ice margins during the past century or so, and a search for evidence of changes on a large time scale, i.e., post-Pleistocene variations. These will be considered in turn.

7.2. RECENT FLUCTUATIONS

7.2.1. *Coastal ice cliffs*

In the summer of 1936-7 the coast of Mac.Robertson Land was mapped in detail by the Lars Christensen Expedition (Christensen 1939). The charts resulting from this aerial survey are believed to show the position of the ice front accurately, since the numerous islands, skerries, and rock outcrops dotted along the coast greatly facilitate photo control. In 1957, special flights were made by ANARE for glaciological purposes, and the snouts of three glaciers were surveyed twice during the year, using the vertical camera of a trimetrogon battery. These surveys were intended to serve two purposes: to check the positions of the ice fronts against their 1936 positions, and to obtain a measure of the rate of flow of the glaciers. It was immediately apparent from the aircraft that no drastic changes had taken place between 1936 and 1957, and this impression was confirmed when the photographs were examined. Prints of the 1936 oblique photographs were made available to the author by the Norsk Polarinstitutt and these were compared with similar aerial oblique photographs taken in 1956. The detailed results follow.

The Dovers Glacier calves into Stefansson Bay and does not exhibit the projecting tongue which generally characterizes fast-moving ice streams along this coast. The many rock exposures at the place where the glacier debouches make it easy to plot the ice front. In December 1957 the ice cliffs were 3½ kilometres in advance of their 1936 positions. The breakaway of a large part of the glacier end is imminent, and if this occurs in 1958 the ice front will once more be in the 1936 position.

The tongue of the Hoseason Glacier projects almost 10 kilometres seawards but is stabilized by rock exposures on each flank. In August 1957 the glacier terminated about 1½ kilometres farther back than the 1936 position.

The tongue of the Jelbart Glacier is in the form of two prongs, and rock projects above the surface of the sea in the bay between them. In 1954 it was observed from the air that much of the west prong shown on the 1936 map had broken off and disappeared. A small part of the east prong calved in 1956 and in 1957 the end of this prong was 1 kilometre short of its 1936 extent. In 1957 the west prong was 4 kilometres behind the 1936 position and flow measurements indicate that it will take about 40 years to regain that former position.

Photographs show that the Taylor Glacier was in a more advanced position in 1956 than in 1936, the difference being only a few hundred metres. The 1936 photographs show that the glacier was about to calve, a split having occurred along the hinge line. Both the 1936 and the 1956 photographs show the hinge line and a longitudinal trough in the same general positions.

Occasional small rocks emerge from the ice cliffs along the coast and some of these can be scrutinised on both 1936 and 1956 photographs. In all the cases examined there were no noticeable changes, so that any fluctuations in the position of the ice front at these points could not have exceeded 100 metres.

Perhaps the most interesting photographs are those showing the Fold Island ice cap. This is a small domed ice cap (25sq km in area) which has formed over a group of small islands lying 7km offshore of the continental ice sheet. It receives a snow accumulation, suffers ablation, and loses ice by calving, so that it might be considered to be a rather sensitive miniature ice cap. The spacing of the easily recognizable annual bands suggests that the residence time for a snow particle in the icecap would be of the order of only 50 years.

Study of the photographs failed to reveal any change in the areal extent or thickness of the ice during the past 20 years.

Before leaving the subject of photographic evidence, it might be well to discourage the immediate consideration of climatic implications in the reported advance of the Sandefjord Bay glacier tongues (Roscoe 1956). The fantastic confusion of icebergs, broken glacier tongues and bay ice in this corner of Prydz Bay apparently persists for many years, but it is quite possible that exceptional gales could clear the ice back to the 1936 position. If this did not occur an eastward extension of the Amery Ice Shelf might result.

It appears, therefore, that there have been no significant changes in the positions of these glacier tongues between 1936 and 1958. The calving period varies from glacier to glacier, and for medium-sized glaciers the cycle of extension to instability, followed by calving, is of at least several years duration.

It might be added that many of the smaller inland valley glaciers of the Mawson Escarpment, Princess Elizabeth Land, and of the Scott Mountains in Enderby Land exhibit elevated lateral moraines which have their crests some 50 metres above the level of the present glacier surface. Numerous cases were examined from the air, but no ground observations proved possible, so that the significance of these features remains a subject for speculation. Péwé (1958a and b) suggests that similar moraines adjacent to alpine glaciers in the McMurdo Sound region may represent minute glacial advances of the last few centuries and Swithinbank (1959) also feels that oscillations of ice level may have occurred.

Finally, a somewhat different set of data might be mentioned. In February 1955 a series of cairns was laid down along the ice-rock contact at Mawson. When the cairns were checked in January 1956, the ice had receded, the average distance being 2·34 metres and the maximum, 6·70 metres (Crohn 1958). In February 1958 the cairns were again checked and this time the average distance from cairns to ice was 1·83 metres with a maximum of 7·30 metres. The plateau ice merges smoothly into the rock and moraine at Mawson, so that the ice along the surface contact is a thin wedge only. This thin ice is likely to be sensitive to normal year to year variations of ablation, and hence these measurements suggest no significant changes.

7.2.2. *Botanical evidence*

On Mount Henderson, 15 kilometres inland from Mawson, the lichen *Gasparinnia harrissonii* was found growing within two metres of the ice-rock contact, which is at an altitude of about 500 metres above sea level. The rate of local migration of this species is not known, but it might be expected to be slow. Some patches of lichen were growing close to the ice on bare granitic gneiss in positions which receive little direct sunlight even in summer, and the rock is very dry for eight months of the year. Beschel (1956), in his writings on the dating of lichens, shows that lichen growth is influenced directly by the amount of available light and water; this being so, conditions on Mount Henderson cannot be particularly favourable to lichen migration. Similar evidence of lichens growing close to the ice-rock contact led Schytt (1953) and Swithinbank (1959) to the conclusion that no glacial retreat has taken place in Dronning Maud Land during the past century. If a method of dating these Antarctic rock lichens should be developed, fresh samples could easily be collected from Mount Henderson as long as Mawson station is occupied.

When the first landing on the coast of Oates Land was made in February 1959 the writer had the opportunity of studying the ice margins in that area. Although a number of small glaciers were no longer filling the valleys and cwms in which they were situated, mosses and lichens were found growing close to the ice-rock contact, which was taken as evidence that there had been no rapid retreat in recent decades. In one case, dense firn was dug away from the rock contact and moss-bearing rock was found buried beneath the icy firn.

7.2.3. *Ice recession on Heard Island*

The foregoing has shown that there is no definite evidence for a recession of the ice margins of the Antarctic continent at the present time. When we go further north, however, the picture becomes very different.

Heard Island is situated in latitude 53°05′S, longitude 73°30′E, about 1750 kilometres north of Mac.Robertson Land and approximately 150 kilometres south of the Antarctic convergence. The island is roughly circular in plan and about 25 kilometres across. It is dominated by the 2,750 metre Mawson Peak, which is an active volcano, and glaciers flow down the mountain flanks to the precipitous coast. The firn limit is at 300 metres above sea level, and the island is cloud-covered for most of the year.

ANARE photo 271 A. Campbell-Drury

PLATE 15

Big Ben and the Baudissin Glacier, Heard Island. The trim line, indicating recent ice thinning, runs round the moraine terrace which is just discernible on the right of the glacier.

Four scientific expeditions called at the island between 1873 and 1930, and in December 1947 a scientific station was established there by the ANARE (Law and Béchervaise 1957). This station was manned until 1955, and since that time ANARE relief ships returning from Antarctica have often called in the autumn.

Evidence for glacial retreat at Heard Island was advanced by Lambeth in 1951, but no dating of the ice recession resulted from his work. It is the opinion of casual observers that the glaciers have receded further during the period 1947-1958, but the photographic evidence is inconclusive. Likewise, photographs taken in 1902 (Drygalski 1908) and 1929 (BANZARE photographic files) do not afford a good means of comparison.

There is one item of evidence, however, which can be regarded as a reliable indicator of recent thinning of the ice (i.e., during the past 100 years or so). On the west bank of the Baudissin Glacier a lateral moraine forms a terrace along the slopes of Mount Drygalski (Plate 15), which is itself covered in old morainal material. Mosses and *Azorella* grow on Mount Drygalski, but the moraine below the terrace is bare of vegetation. Thus there is a clearly defined trim line very similar in appearance to those seen in Iceland. The elevation of the trim line above the present glacier surface was estimated at 30 metres.

The wind directions at Heard Island and over the surrounding ocean have changed during the past century. A hundred years ago there were far fewer south-westerly winds and far more northerly winds than there are at the present time (Loewe, Radok and Grant 1952).

It is not really surprising to find that climatic amelioriation has produced a decrease in the volume of ice on Heard Island without bringing about any noticeable changes in Antarctica. At Heard Island convection and rainfall are more important agents of ablation than radiation, and a rise in the mean temperature will intensify their effectiveness. Changes in the seasonal distribution of precipitation could also affect the ice balance. In Antarctica, however, calving is of greater importance than ablation, which in any case is brought about chiefly by radiation. A minor rise in the mean temperature would not be likely to influence the rate of calving or ablation there to any great extent, and the reaction of the continental ice sheet to climatic fluctuations must be expected to be slow.

7.3. MAJOR VOLUME CHANGES OF THE ANTARCTIC ICE

When flying over the mountains of Enderby Land, Kemp Land, Mac.Robertson Land, and Princess Elizabeth Land, it is immediately apparent that the ice sheet was once of a much greater thickness. In the Prince Charles Mountains and in the Scott and Tula Mountains, high mountains have moraine-covered flanks, and smaller mountains may be completely moraine-covered. In these same mountains deserted cwms and glacial valleys may be seen (Plate 16), while along the coast there are numerous islands exhibiting dry valleys, erratics, and roches moutonnées profiles. The height of the moraine deposits on the higher mountains suggests that the ice sheet formerly had a thickness at least 600 metres greater than its present one, at a distance of 100 to 200 kilometres from the present coastline. Swithinbank (1959) states that the present ice level in Dronning Maud Land is at least 800m lower

ANARE photo

PLATE 16

An ice-free glacial valley in the Mawson Escarpment. This valley was formerly occupied by a tributary glacier bringing ice into the main stream of the Lambert Glacier, but the ice has disappeared during the post-Pleistocene retreat.

than at some previous glacial maximum, and Péwé (1958a) gives the maximum height of moraine deposits around McMurdo Sound as 610m. During a survey of a mountain range 650 kilometres south of Mawson, the author found the lower hills and nunataks to be completely moraine-covered, and estimated the height of the moraine on the main peak, Mount Menzies, as 450 metres above the present ice level. The elevation of the surface of the ice sheet in this region is 1,900 metres above sea level. In the Framnes Mountains, close to Mawson, erratics have been observed 180 metres above ice level, and dry glacial valleys can be seen up to 120 metres above the present ice surface.

Terraces on the rock slopes at Welch Island and near Byrd Head were tentatively identified as raised beaches. The elevation of the terraces was between 20 and 40m above present sea level. On the coast of Oates Land, sea shells were found on the rocky hillsides up to 45m above sea level, but it is possible that these have been dropped by birds. Russian workers found terraces from 2 to 120m above sea level and on terraces from 5 to 15m above sea level numerous shells and calcareous remnants of sea creatures were discovered (Voronov 1958).

The coast of East Antarctica is fringed by numerous submarine banks upon which large numbers of icebergs become grounded. These banks, which lie 80 to 150 kilometres north of the present coastline have been described as old moraines (Mawson 1935) which indicate a former extent of the ice sheet. If this assumption is correct, it might answer the question posed by Swithinbank (1959), who asks where the eroded rock of Antarctica has been deposited. Conflicting theories are held about the origin of these banks, but it is by no means impossible that they are, in fact, ancient terminal moraines.

There is ample evidence, then, that the ice sheet has been much thicker in the past. It is not so easy to decide how long ago this major recession occurred. Some idea of the time scale involved can be formed by estimating the volume of ice lost since the previous glacial maximum and comparing it with rates of loss by ice flow and ablation based on present day measurements. If it is assumed that nourishment of the ice cap ceased during the glacial retreat, a minimum estimate of the time required to achieve the necessary draw-down of ice level can be made. Using values based on the figures give in Section 8 it is found that the major recession could not have been initiated less than about 10,000 years ago.

In the Prince Charles Mountains the author found evidence suggesting that the major ice recession terminated some considerable time ago. Two large melt-water lakes, Radok Lake and Beaver Lake,* are connected by a winding, water-cut gorge, the lakes themselves occupying rock bowls at the foot of a massif and some 500-600m below the highest level of the moraines (see Plate 17). The gorge connecting the two lakes is about 6km long, 60m deep with sideslopes of approximately 45°, and has a total fall of perhaps 60m. The rock is a soft sandstone and water flows out of Radok Lake for only a few weeks each year, the pattern of refrozen melt-water on the surface of Beaver Lake suggesting that the discharge from the channel is not excessive. It is difficult to estimate the length of time necessary to cut such

* The somewhat complex character of Beaver Lake is described elsewhere (Mellor and McKinnon 1959).

ANARE photo 7209 M. Mellor

PLATE 17

A winding water-cut gorge joins Radok Lake (top) and Beaver Lake (bottom). The dark patch around the end of the gorge is relatively bubble-free ice formed by re-freezing of melt water.

a channel, since erosion rates for established water courses are not applicable and calculations cannot be made without stream gaugings. Water leaving lakes generally carries little sediment and is only mildly erosive (Wagner 1950) and this would indicate a slow cutting process in the case under consideration, particularly as water flows for only a short period each year. On the basis of experience gained whilst engaged in river control engineering, the writer estimated that the gorge must have taken at least several centuries to become fully formed. The channel now seems to have a stable profile, and so the main erosion may have been complete for some time. Taking all these things into consideration it, can confidently be stated that in the Lambert Glacier area the ice has not been much above its present level during the past few hundred years.

7.4. CONCLUSIONS

The studies of recent ice fluctuations show that, in Mac.Robertson Land and Kemp Land, the ice has suffered neither thinning nor retreat to any detectable extent during the past 21 years. The botanical evidence suggests that the volume of ice may have been constant for a longer period. Photographic evidence from McMurdo Sound (Péwé 1958a and b) shows that only negligible changes have occurred in that region in 46 years. Lister (1959) reports that the ice margins around the

Shackleton Mountains are static or in slow retreat, and Shumskiy (1957) has given evidence that there has been no appreciable retreat around the Bunger Hills in recent decades. Markov (reported in Lister 1959) found that the ice sheet had not receded from Gaussberg since the visit of Drygalski in 1902 (Drygalski 1908), and Russian work in Bunger Hills (Avsyuk, Markov and Shumskiy 1956) showed that the ice edge there was either stationary or in slow retreat.

One theory holds that climatic amelioration might lead to increased precipitation in Antarctica, with a resultant expansion of the ice sheet. Loewe (1956, 1960) has suggested that the level of the inland ice may be rising, but considers that centuries may elapse before the effects appear in the fringe regions. No such tendency has been exhibited in the Fold Island ice cap during the past 20 years. Air temperatures in the Arctic have risen since the beginning of the present century but a decrease in the Greenland precipitation during the last 30 years has been shown (Diamond 1958). On the other hand, certain accumulation measurements (Mellor 1957) have indicated an increasing precipitation trend in Spitsbergen, where observations show the recent temperature rise to have culminated about 1940 (Hesselberg and Johannessen 1957).

There is some evidence that recent changes in atmospheric circulation have affected conditions over the Southern Ocean, but so far no observable changes in the volume of Antarctic ice have resulted. This could be because the changes have not affected the overall regime of the ice sheet, or merely because effects of a response are slow in reaching the marginal regions.

The Antarctic ice cap was once of a much greater volume, the ice level being up to 600m higher at some previous glacial maximum in the regions studied. It is believed that the major recession which eventually reduced the ice cap to its present size was initiated not less than 10,000 years ago. Shumskiy (Avsyuk, Markov and Shumskiy 1956) states that the minimum time necessary to free the once-covered Bunger Oasis from ice must have been considerably greater than 4,500 years and he further mentions that the oasis must have been ice-free for a long period. Shumskiy (1957) also gives the opinion that the great ice retreat took place more than 10,000 years ago. The Russian and Australian estimates thus seem to be in fair agreement and they suggest that the decline of the Pleistocene glaciation and the onset of the Climatic Optimum in Antarctica were contemporaneous with these events in other parts of the world.

During the last 2,000 years there have probably been minor re-advances and retreats of the ice, but the present observations give no means of dating these fluctuations.

8. THE MASS BUDGET

In the previous sections the import and export of ice and its redistribution within Antarctica have been discussed, and estimates of the rates at which the various mechanisms proceed have been given. In the last Section it was shown that the recent changes in atmospheric circulation which have produced drastic reductions in the size of glaciers in some parts of the world have had no noticeable effects on the margins of the Antarctic ice sheet so far, and it now remains to consider quantitatively the gains and losses of ice.

In drawing up a mass balance for Antarctica it is important to distinguish between a meteorological water balance, in which total water vapour import is estimated, and a glaciological mass balance, in which the gain of ice mass is assessed from the net accumulation of snow. In a meteorological budget *total* ice losses must be balanced against water vapour import, but for a glaciological balance it is only necessary to estimate *net* ice losses and the *net* accumulation of snow on the continent. Thus, contrary to the practice followed by several previous writers, wind-blown snow and ablation losses from above the firn limit do not enter into a balance in which mass gains are determined from measurements of snow accumulation. They remain important estimates, however, as they may be added to the measured net accumulation to give some idea of the precipitation together with formation of hoar frost and rime over Antarctica.

8.1. PRECIPITATION OVER THE ICE SHEET

So far it has not been possible to measure precipitation directly in Antarctica, but it has sometimes been assumed that snow accumulation gives a measure of the precipitation. This implies that snow transport and evaporation produce negligible losses. Whilst neglect of the latter may be justified, the Mawson drift-gauging suggests that export of blown snow may be considerable and so mean precipitation will be estimated here from net accumulation plus the mass of ice lost by snow transport and evaporation.

In the sector between longitude 45°E and 80°E the mass of water substance precipitated between the coast and a boundary about 850km inland would be obtained by summing the following quantities:

Net accumulation	12×10^{16}g/yr
Blown snow	$2{\cdot}6 \times 10^{16}$g/yr
Evaporation (other than that in net ablation zones)	$0{\cdot}9 \times 10^{16}$g/yr
Total	$15{\cdot}5 \times 10^{16}$g/yr

This falls over an area of approximately 1,100,000km^2, so that the mean precipitation is 14cm of water. The zone lies roughly between the latitudes of 66°S and 76°S, and the precipitation estimate of Meinardus (1938) for the belt of Antarctica between latitudes 70°S and 75°S is 11cm.

The precipitation over the whole of the ice sheet can be estimated in the same way:

Net accumulation	$1{\cdot}6 \times 10^{18}$g/yr
Blown snow	$0{\cdot}19 \times 10^{18}$g/yr
Evaporation	$0{\cdot}06 \times 10^{18}$g/yr
Total	$1{\cdot}85 \times 10^{18}$g/yr

Over the 14,000,000km^2 of Antarctica this represents a mean precipitation of 13cm of water. Meinardus (1938) estimates a mean precipitation of 7cm between latitudes 70°S and 90°S, and Kosack (1954) gives a figure of 20cm. Using accumulation and precipitation as synonymous terms, Lister (1959) estimates 11cm.

Mass budget for the sector from long. 45°E to long. 80°E

The rates of loss and gain are here multiplied by appropriate effective lengths for each process to give a tentative mass budget for the sector. Most of the estimates made for this sector by the writer have already been compared with the results of other workers and they are used here without modification.

Ice flow leads to the export of the following quantities of ice:

Ice shelf (230km)	$1{\cdot}5 \times 10^{16}$g/yr
Ice streams (270km)	$1{\cdot}3 \times 10^{16}$g/yr
Sheet flow (1,600km)	$0{\cdot}3 \times 10^{16}$g/yr
Total	$3{\cdot}1 \times 10^{16}$g/yr

Net ablation is comparatively important in this sector and, if it is assumed to proceed at the Mawson rate over a length of 2,100km, the loss of ice is $1{\cdot}1 \times 10^{16}$g/yr.

Evaporation above the firn limit is probably responsible for an ice loss (over a length of 2,100km) of about $0{\cdot}95 \times 10^{16}$g/yr.

Drifting snow, if it is effective at the Mawson rate over a length of 1,600km, will remove $2{\cdot}6 \times 10^{16}$g/yr.

Oceanic melting is only effective over a length of about 250km and, at the rate suggested earlier, will remove $0{\cdot}15 \times 10^{16}$g/yr.

Net accumulation at the strip rate quoted in section II is probably effective over a length of about 1,200km, so that the annual gain of ice will be 12×10^{16}g/yr.

Thus the net losses amount to only $4{\cdot}4 \times 10^{16}$g/yr whilst the net gain of ice is 12×10^{16}. This implies that the ice cap is growing at the present time, but before accepting such a radical idea we must reconsider the estimates and investigate the basic data for gross errors.

From the existing measurements in Mac.Robertson Land there seems to be little likelihood that the accumulation is overestimated for the first 600km from the coast, although there could be an abrupt decrease further inland. Against the results of other expeditions, however, the estimated accumulation for Mac.Robertson Land seems to be on the low side and no valid reason for reducing the estimate can be found at present. This leads us to suspect the loss estimates. The most important single loss process is iceberg calving and it must be admitted that ice flow quantities may well be under-estimated. The thickness assumed for the edge of the continental ice is rather small and the contributions of ice streams could be larger than is indicated here. It is also possible that the movement of the Amery Ice Shelf is more rapid than was assumed. If all these criticisms are allowed then the export of ice by calving might be raised by a factor of 2. The estimate of net ablation is the best that could be made on the available data and, as the figure given is already very high in comparison with corresponding values for other parts of Antarctica, any attempt to raise it still further would probably meet with adverse criticism. This leaves only the estimate of oceanic melting to be commented on. Since the length of ice shelf along the coast of the sector under discussion is only 250km, raising the rate of melting by an order of magnitude only elevates this process to the same level of importance as net ablation, so that it can hardly be underestimation of oceanic melting which is producing the discrepancy between losses and gains in the budget given above.

If, therefore, the accumulation estimate for this sector is allowed to stand, and the loss estimates are increased to what are at present thought to be maximum values, there still remains an excess of accumulation, the figures being 12×10^{16}g/yr against about 9×10^{16}g/yr. Thus, whilst feeling the need for more data, we are at the same time forced to admit the possibility of an increasing ice volume in this sector. Before this theory is discussed further it will be of interest to draw up a mass budget for the whole of the Antarctic ice sheet to see what trends are indicated for the entire continent.

8.2. MASS BUDGET FOR THE ANTARCTIC ICE SHEET

In drawing up a budget for the whole of Antarctica it is necessary to consider the results of workers in other parts of the continent and to decide on rates of loss and gain which are most nearly typical. It must then be decided over what lengths of territory the various processes are active at the mean rates to deduce the final entries for the mass balance.

Ice flow export from the whole of Antarctica was estimated in section III and the figures given were:

Ice shelf (7,500km)	$0{\cdot}48 \times 10^{18}$g/yr
Ice streams (1,500km)	$0{\cdot}07 \times 10^{18}$g/yr
Sheet flow (11,000km)	$0{\cdot}02 \times 10^{18}$g/yr
Total	$0{\cdot}57 \times 10^{18}$g/yr.

These estimates take account of most of the pre-IGY measurements of ice flow and it was mentioned previously that the total export by calving was estimated by Loewe at $0 \cdot 67 \times 10^{18}$g/yr and by Lister at $0 \cdot 14 \times 10^{18}$g/yr.

Net ablation is widely believed to be a negligible factor in the Antarctic mass economy. Whilst Loewe (1960) may be correct in his view that melt losses are of only local importance, it is hard to ignore the feeling that coastal runoff and evaporation are significant after seeing strong summer ablation taking place along the whole length of Australian Antarctic Territory. There is a need for more widespread observations, but it seems possible for ablation to account for as much as 5% of the ice loss. If the Mawson rate is applied over a length of 14,000km the overall loss becomes $0 \cdot 074 \times 10^{18}$g/yr.

Evaporation above the firn limit is generally believed to be balanced by direct crystallization on the surface, but there is a possibility that an evaporation loss uncompensated by crystallization occurs in the fringe regions. Applying the rate estimated for Mawson over a length of 14,000km the result is: $0 \cdot 063 \times 10^{18}$g/yr. Lister (1959) gives a figure for evaporation loss which is equivalent to $0 \cdot 014 \times 10^{18}$g/yr, or about one quarter of the present writer's value.

Drifting snow estimates vary considerably, but if the rate accepted for Mawson is effective over a circumferential length of 12,000km the total loss is $0 \cdot 19 \times 10^{18}$g/yr. This is an order of magnitude higher than the $0 \cdot 02 \times 10^{18}$g/yr claimed by Loewe (1959) after considering the mean rate of snow transport for the whole coast to be one thirtieth of the value for Terre Adélie. Lister (1959) gives $0 \cdot 014 \times 10^{18}$g/yr. However, some justification of the present high estimate has been given in the section dealing with blowing snow.

Oceanic melting cannot really be estimated on currently available data. The rate of melting suggested in Section 6, if applied to a total ice front length of 8,000km, gives a loss of only: $0 \cdot 05 \times 10^{18}$g/yr.

Lister assumes a value fourteen times larger than this, whilst Wexler's figure is twenty-six times bigger.

Accumulation measurements by various workers in different parts of the ice sheet have led to much more consistent values than have the measurements of loss processes. Three meridional profiles through the main accumulation zone were mentioned in Section 2. These were:

Shumskiy (USSR Expedition)	$1 \cdot 8 \times 10^{14}$g/km/yr
Schytt (Norwegian-British-Swedish-Expedition)	$1 \cdot 5 \times 10^{14}$g/km/yr
ANARE	$1 \cdot 0 \times 10^{14}$g/km/yr.

A mean value of $1 \cdot 4 \times 10^{14}$g/km/yr might be taken over an effective length of 11,000km. This gives the accumulation in the peripheral zone, to which a further 15% might be added to allow for an average accumulation of 5cm of water over the heart of the continent. The resulting figure for the accumulation over the whole of Antarctica would then become $1 \cdot 7 \times 10^{18}$g/yr.

This has to be compared with the estimates of Loewe (1960) and Lister (1959), who give values of $1 \cdot 58 \times 10^{18}$g/yr and $1 \cdot 54 \times 10^{18}$g/yr respectively. Taking these into consideration, a good mean value for the overall net accumulation would be $1 \cdot 6 \times 10^{18}$g/yr.

Totalling up the net gains and net losses we find that there is again a surplus of accumulation. Whilst the net gain is $1 \cdot 6 \times 10^{18}$g/yr, the losses (calving, oceanic melting and net ablation) amount to no more than $0 \cdot 7 \times 10^{18}$g/yr.

It has already been shown that the various accumulation estimates are in fair agreement and so the figure given can be accepted for the time being. As before, the ice exported by calving may have been underestimated; the mean thickness of continental ice assumed seems rather small for the whole of Antarctica and, as mentioned before, the role of ice streams may be somewhat more important than has been suggested. Of ice shelf movement we still know very little. The export estimate given here is four times bigger than Lister's figure and fourteen times bigger than that given by Wexler (1958). Loewe (1960) suggests a slightly larger value than the one given here. The estimates of Lister and Wexler seem to be too small and it is even felt that the present estimate is on the low side, so that a 23% increase to $0 \cdot 7 \times 10^{18}$g/yr might be conceded. In view of the widespread opinion that ablation is responsible for only negligible ice removal there is no justification for raising that estimate. Having reached the maximum admissible values for calving and ablation it remains to consider what rates of melting beneath ice shelves would be necessary to preserve a balance and whether, in fact, the ice sheet is in equilibrium.

With the enhanced ice flow value we find that losses amount to about half the annual net gain, so that there remain $0 \cdot 8 \times 10^{18}$g/yr to be accounted for. Taking an ice shelf length of 8,000km, it would require basal melting at a mean rate of 200cm of water per year over the outermost 50km of the shelves. This is at least an order of magnitude higher than the value suggested by Swithinbank (1958) and, as Loewe (1960) has already pointed out, acceptance of this figure necessitates what seem to be unrealistic assumptions regarding surface accumulation and the inflow velocity of continental ice.

Thus it seems improbable that the surplus of accumulation over loss can be completely explained by invoking theories of strong basal melting beneath ice shelves and, in spite of the flimsy structure of the estimates, it is necessary to consider Loewe's (1956; 1960) suggestions that the level of the inland ice is rising.

The writer's observations in the coastal regions of Australian Antarctic Territory indicate that the edges of the ice sheet are neither advancing nor retreating to any appreciable extent at the present time, which is in broad agreement with the findings of Schytt (1953; 1954) and Swithinbank (1959) in Dronning Maud Land, Péwé (1958 a & b) in the Ross Dependency, and Lister (1959) in the Shackleton Mountains. Russian glaciologists ("Antarctic", 1958) claim that the ice margins are in retreat, although the rate of recession is very much smaller than has been common for Northern Hemisphere glaciers in the last few decades. However, Loewe (1960) believes that it will be some hundreds of years before the effects of increased accumulation are felt in the coastal regions, even taking into account the more rapid rate of travel of "waves" produced by increased accumulation (Nye 1958).

The observed contemporary rise in mean sea level has often been adduced as evidence for the melting of the two great ice caps of the earth. Loewe (1960) points out that, with Antarctica gaining substance at the rate of $0{\cdot}25 \times 10^{18}$g/yr, a drop in mean sea level of 5·5mm per decade might be expected, but Gutenberg (1941) finds a contrary effect, i.e., a rise of 10mm per decade. Loewe then goes on to show that rises in sea level need not be caused by glacier melting but may be due to water expansion following temperature rises and, furthermore, the temperature increases required to produce the observed rise in sea level are not unreasonable when compared with increases in global air temperatures and in surface temperatures of the waters of north-west Europe.

It is difficult to carry the argument further without access to stratigraphic results from deep bores and results of margin variation studies employing modern radioisotope techniques. Further useful investigations might be made on the nunataks and mountains which lie far inside the accumulation zone, such as the southern extremities of the Prince Charles Mountains.

At the moment it is hard to see how the Antarctic mass budget can be balanced, but at the same time there is no direct evidence that the volume of the inland ice is increasing. When such a big problem is considered with so few data to hand, widely varying interpretations can be put on observations. Thus it is felt that, until more accurate, widespread, and inter-related data become available, the regime and mass economy of the Antarctic ice sheet cannot be firmly established.

9. ACKNOWLEDGEMENTS

The basic glaciological programme of the Australian National Antarctic Research Expedition for the International Geophysical Year was designed to provide as much data as possible on the mass economy of the Antarctic ice cap. A number of subsidiary investigations were also undertaken, and although these were originally intended to supplement the main mass balance studies, some of them were later to provide absorbing studies in themselves. In this paper the results of the mass economy studies are given and a picture of the nourishment and de-icing of the Antarctic ice sheet is built up.

The author, as ANARE glaciologist, carried out field work from Mawson station during 1957 and part of 1958 and was fortunate in being able to travel widely by aircraft and by tractor. He also had the opportunity of visiting the most easterly parts of Australian Antarctic Territory early in 1959 during a summer cruise on M.V. *Magga Dan*. Most of the observations in this work were made by the author, assisted by various expedition members, but some observations made by Australian parties prior to the IGY are also referred to. A skeleton glaciological programme was maintained at Mawson after the author left, supplementary measurements being made by members of the 1958 wintering party to provide a check on the 1957-58 results.

The data were collected in a period of about fourteen months, but it has required an equal length of time to analyse and interpret the results. The expenditure and effort of Antarctic expeditions have sometimes been wasted when lack of funds has prevented scientists from working up their results fully. The author was fortunate, however, in being able to work full-time on data analysis for more than a year after his return from Antarctica, finance being provided by the Australian Academy of Science.

The author would like to express his gratitude to Dr. P. G. Law, formerly Director of the Antarctic Division, for making this material available and to Dr. F. Loewe, formerly Head of the Meteorology Department, University of Melbourne, for his constant interest and advice. He is particularly indebted to Dr. Uwe Radok for guidance and encouragement throughout the work.

10. REFERENCES

AHLMANN, H. W. (1953). Glacier variations and climatic fluctuations. Isaiah Bowman Memorial Lecture, American Geographical Society: New York.

ANDREE, S. A. (1883). Om yrsnön i. de arktiska trakterna, *Kal. Vet. Akad. Förh.* No. 9: 33.

AVSYUK, G. A., MARKOV, K. K., *and* SHUMSKY, P. A. (1956). Geographical observations in an Antarctic "oasis". USSR Academy of Sciences: Moscow.

BADER, H. (1950). The significance of air bubbles in glacier ice. *Journal of Glaciology* **1**: 443-451.

BADER, H. (1954). Sorge's law of densification on high polar glaciers. *Journal of Glaciology* **2**: 319-323.

BAGNOLD, R. A. (1936). The movement of desert sand. *Proceedings of the Royal Society* (A) **157**: 594-620.

BAGNOLD, R. A. (1941). The Physics of Blown Sand and Desert Dunes, Methuen: London.

BARRÉ, M. (1954.) Propriétés electriques du blizzard. Résultats Scientifiques, No. S. IV. 1. Expéditions Polaires Française: Paris.

BAUSSART, M. (1958). Essai de détermination par photogrammétrie de la vitesse superficielle d'un glacier du Groenland. Symposium de Chamonix: Physique du Movement de la Glace. International Union of Geodesy and Geophysics, International Association of Scientific Hydrology: Gentbrugge.

BESCHEL, R. (1956). Translated abstracts of various papers given by G. A. Llano in "Botanical Research in Antarctica". Antarctica in the IGY American Geophysical Union, National Academy of Sciences: Washington.

CHRISTENSEN, L.(1939). Recent reconnaissance flights in the Antarctic. *Geographical Journal* **94**: 192-202.

CROHN, P. W. (1959). A contribution to the geology and glaciology of the western part of Australian Antarctic Territory. *ANARE Reports* (A) **3**. Publication No. 49.

DAVID, T. W. E., *and* PRIESTLEY, R. E. (1914). Geology. Reports on Scientific Investigations, British Antarctic Expedition 1907-09: London.

DEBENHAM, F. (1949). The problem of the Great Ross Barrier. *Geographical Journal* **112**: 196-218.

DIAMOND, M. (1958). Precipitation trends in Greenland during the past thirty years. *Journal of Glaciology* **3**: 177-180.

DOLGOUCHINE, L. D. (1958). Les particularités morphologiques essentielles et les régularités des mouvement des glaciers de la marge de l'Antarctide orientale. Symposium de Chamonix: Physique du Mouvement de la Glace. International Union of Geodesy and Geophysics, International Association of Scientific Hydrology: Gentbrugge.

DRYGALSKI, E., *et al.* (1908). Deutsche Sudpolar Expedition, 1901-03: Berlin.

GEORGI, J. (1956). Temperatur-und Dichtemessungen des Grönlandischen Inlandeises. *Annales de Géophysique* **12**: 102-110.

GLEN, J. W. (1955). The creep of polycrystalline ice. *Proceedings of the Royal Society* (A) **228**: 519-538.

GOODSPEED, M. J. (1958). Preliminary report of ice thickness measurements, Mawson 1957-58. Records 1958/27, Bureau of Mineral Resources, Geology and Geophysics: Melbourne.

GOULD, L. M. (1957). Antarctic prospect. *Geographical Review* **47**: 1-28.

GUTENBERG, B. (1941). Changes in sea level, post-glacial uplift, and mobility of the Earth's interior. *Glaciological Society of America Bulletin* **52**: 721-772.

HESSELBERG, T., *and* JOHANNESSEN, T. W. (1957). The recent variations of the climate at the Norwegian Arctic stations. Polar Atmosphere Symposium. Special supplement of the *Journal of Atmospheric and Terrestrial Physics.*

HOFMANN, W. (1958). Bestimmung von Gletschergeschwindigkeiten aus Luftbildern. Bildmessung and Luftbildwesen. Heft 3, Seite 71-88.

HÜTTE. (1955). Engineering Handbook, Vol I: Berlin.

IMBERT, B. (1959). Détermination de l'épaisseur de glace en Terre Adélie. Comptes Rendus des Séances de l'Academie des Sciences **248**: 576-579.

KAPITZA, A. P. (1958). O zavisimosti formy lednikovogo kupola vostochnoy Antarktidy ot rel'efa podlednogo lozha i kharaktera rastekaniya l'da. Informat sionnyy Byulleten', Sovetskoiy Antarkticheskoy Ekspeditsii. l. Izdatel'stvo "Morskoy Transport", Leningrad.

KAPITZA, A. P. (1958). Opredelenie skorosti dvizheniya vyvodnykh lednikov po gofrirovke morskogo pripaiya. *Ibid.*

KNOWLES, P. H. (1945). Glaciology of southern Palmer Peninsula Antarctica. Reports on Scientific Results of the United States Antarctic Service Expedition 1939-41. *Proceedings of the American Philosophical Society* **89**: 174-176.

KOSACK, H. P. (1954). Zur Frage des Wesserhaushalts in der Antarktis. *Polarforschung* **3**: 286-293.

KRICHAK, O. G. (1958). Some results of meteo research carried out by Second Continental Antarctic Expedition of the USSR Academy of Sciences. Symposium of Antarctic Research. New Zealand Department of Scientific and Industrial Research.

LAMBETH, A. J. (1951). Heard Island—Geography and Glaciology. *Proceedings of the Royal Society of N.S.W.* **84**: 92-98.

LAW, P. G., *and* BÉCHERVAISE, J. M. (1957). ANARE Australia's Antarctic Outposts. Oxford University Press: Melbourne.

LETTAU, H. (1939). Atmosphärische Turbulenz. Leipzig.

LILJEQUIST, G. (1957). Energy exchange of an Antarctic snowfield. Scientific Results of the Norwegian-British-Swedish Antarctic Expedition 1949-52. Vol. II, Part 1 C: Oslo.

LISTER, H. (1959). International glaciology in the Antarctic. *The Times Science Review* No. 31.

LOEWE, F. (1956). Études de glaciologie en Terre Adélie, 1951-52. Expéditions Polaires Françaises: Paris.

LOEWE, F. (1956a). Notes on firn temperatures and ablation in Mac.Robertson Land Antarctica. *Journal of Glaciology* **2**: 725-726.

LOEWE, F. (1957). Precipitation and evaporation in the Antarctic, *in* M.P. van Rooy, Meteorology of the Antarctic, Weather Bureau, Pretoria.

LOEWE, F. (1960). Notes concerning the mass budget of the Antarctic inland ice. *In* Antarctic Meteorology, pp. 361-369. Pergamon Press: London. Melbourne.

LOEWE, F., RADOK, U., *and* GRANT, A. (1952). Eine säkulare Klimaänderung im südlichen Indischen Ozean? Berichte des Deutschen Wetterdienstes in der U.S. Zone, No. 42: Bad Kissingen.

MCLEOD, I. *et al.* (1959). (Unpublished observations from Mawson, 1958.) Antarctic Division: Melbourne.

MATHER, K. B. (1960). Katabatic winds south of Mawson. *In* Antarctic Meteorology, pp. 317-320. Pergamon Press: London.

MAWSON, D. (1915). Home of the Blizzard. Hodder and Stoughton: London.

MAWSON, D. (1935). The unveiling of Antarctica. Australian and New Zealand Association for the Advancement of Science: Sydney.

MAWSON, D. (1942). Australasian Antarctic Expedition 1911-14. Scientific Reports (A) **1**: 271-272.

MEINARDUS, W. (1938). Klimakunde der Antarktis. *In* W. Koppen and R. Geiger: Handbuch der Klimatologie, Band IV (U): Berlin.

MELLOR, M. (1957). Glacier observations in north-west Spitsbergen. *Journal of Glaciology* **3**: 61-66.

MELLOR, M. (1958a). Australian glaciological contributions in Antarctica—preliminary report. *Journal of Glaciology* **3**: 279-285.

MELLOR, M. (1958b). Photogrammetric flow measurements on Antarctic glaciers. *Transactions: American Geophysical Union* **39**: 1158.

MELLOR, M. (1959a). Gauging Antarctic drift snow. Antarctic Meteorology. Pp. 347-355. Pergamon Press: London.

MELLOR, M. (1959b). Ice flow in Antarctica. *Journal of Glaciology* **3**: 377-385.

MELLOR, M. (1959c). Variations of the ice margins in East Antarctica. *Geographical Journal* **125**: 230-235.

MELLOR, M. (1959d). Sea ice measurements at Mawson and Davis, 1954-58. *ANARE Interim Reports.* (Publication No. 55.)

MELLOR, M. (1959e). Mass balance studies in Antarctica. *Journal of Glaciology* **3**: 522-533.

MELLOR, M. (1959f). Creep tests on Antarctic glacier ice. *Nature* **184**: 717.

MELLOR, M. (1959g). Creep tests on polycrystalline Antarctic ice. (Unpublished paper.)

MELLOR, M., *and* MCKINNON, G. (1959). The Amery Ice Shelf and its hinterland. *Polar Record* **10**: 30-34.

MELLOR, M., *and* RADOK, U. (1959). Some properties of drifting snow. *In* Antarctic Meteorology. Pp. 333-346. Pergamon Press: London.

NYE, J. F. (1951). The flow of glaciers and ice-sheets as a problem in plasticity. *Proceedings of the Royal Society* (A) **207**: 554-572.

NYE, J. F. (1952a). The mechanics of glacier flow. *Journal of Glaciology* **2**: 82-93.

NYE, J. F. (1952b). A method of calculating the thickness of ice sheets. *Nature* **169**: 529-530.

NYE, J. F. (1957). The distribution of stress and velocity in glaciers and ice-sheets. *Proceedings of the Royal Society* (A) **239**: 113-133.

NYE, J. F. (1958). Surges in glaciers. *Nature* **181**: 1450-1451.

PÉWÉ, T. L. (1958b). Quaternary glacial geology of the McMurdo Sound region Antarctica—Symposium on Antarctic Research, New Zealand National Committee for the IGY. Department of Scientific and Industrial Research: Wellington.

PÉWÉ, T. L. (1958b). Quaternary glacial geology of the McMurdo Sound region Antarctica—a progress report. IGY Glaciological Report Series, No. 1. American Geographical Society: New York.

POUND, T. N. (1958). Calibration of Antarctic snow gauges, A.T.M. No. 139. Aeronautical Research Laboratories, Department of Supply: Melbourne.

RENAUD, A. (1949). A contribution to the study of the glacier grain. *Journal of Glaciology* **1**: 320-324.

ROBIN, G. DE Q. (1958). Seismic shooting and related investigations. Norwegian-British-Swedish Antarctic Expedition 1949-52, Scientific Results, Vol. 5: Oslo.

ROSCOE, J. (1956). Antarctic photogeography. Antarctica in the IGY. American Geophysical Union, National Academy of Sciences: Washington.

ROSSBY, C. G., *and* MONTGOMERY, R. B. (1935). Papers in physical oceanography and meteorology. Massachusetts Institute of Technology. **3**: No. 3.

SCHYTT, V. (1953). The Norwegian-British-Swedish Antarctic Expedition 1949-52. Summary of the glaciological work—preliminary report. *Journal of Glaciology* **2**: 204-205.

SCHYTT, V. (1954). Glaciology in Queen Maud Land. *Geographical Review* **44**: 70-87.

SCHYTT, V. (1958). Glaciology II. Norwegian-British-Swedish Antarctic Expedition, 1949-52, Scientific Results. Vol. **4**: Oslo.

SHIOTANI, M., *and* ARAI, H. (1953). A short note on the snow storm. Proceedings of the Second Japan National Congress for Applied Mechanics, 1952. Science Council of Japan: Tokyo.

SHUMSKIY, P. A. (1955). Osnovy strukturnogo ledovedeniia. Izdatel'stvo Akademii Nauk CCCP, Moscow.

SHUMSKIY, P. A. (1957). Reconnaissance in the Antarctic 1956. *Journal of Glaciology* **3**: 56-61.

SHUMSKIY, P. A. (1958). The mechanism of ice straining and its recrystallization. Symposium de Chamonix: Physique du mouvement de la glace. International Union of Geodesy and Geophysics, International Association of Scientific Hydrology: Gentbrugge.

SHUMSKIY, P. A. (1958a). *In* Symposium on Antarctic Research. New Zealand National Committee for the IGY Department of Scientific and Industrial Research: Wellington.

SIPLE, P. A. (1958). Man's first winter at the South Pole. *National Geographic Magazine* **113** 439: 478.

STEINEMANN, S. (1954). Flow and recrystallization of ice. General Assembly of Rome. International Union of Geodesy and Geophysics, International Association of Scientific Hydrology. Tome 4.

STEINEMANN, S. (1958). Résultats expérimentaux sur la dynamique de la glace et leurs corrélations avec la pétrographie des glaciers. Symposium de Chamonix: Physique du mouvement de la glace. International Union of Geodesy and Geophysics, International Association of Scientific Hydrology: Gentbrugge.

STEPHENSON, P. J., *and* LISTER, H. (1959). Preliminary results of the glaciological work on the Trans-Antarctic Expedition 1955-58. *Journal of Glaciology* **3**: 426-431.

SWITHINBANK, C. (1955). Ice Shelves. *Geophysical Journal* **121**: 72-75.

SWITHINBANK, C. (1957a). Glaciology. I. The morphology of the ice shelves of western Dronning Maud Land. Norwegian-British-Swedish Antarctic Expedition, 1949-1952. Scientific Results. **3** (A): 1-37. Norsk Polarinstitutt: Oslo.

SWITHINBANK, C. (1957b). Glaciology. I. The regime of the ice shelf at Maudheim as shown by stake measurements. *Ibid*. **3** (B): 41-75.

SWITHINBANK, C. (1958). Glaciology. I. The movement of the ice shelf at Maudheim. *Ibid*. **3** (C): 77-96.

SWITHINBANK, C. (1959). Glaciology. I. The morphology of the inland ice sheet and nunatak areas of western Dronning Maud Land. *Ibid*. **3** (D): 99-116.

TRESHNIKOV, A. F. (1958). Novye dannye o vysotakh vostochnoiy Antarktidy. Informatsionnyy Byulleten', Sovetskoy Antarkticheskoy Ekspeditsii, l, Izdatel'stvo "Morskoy Transport": Leningrad.

VICKERS, W. W. (1958). Snow accumulation on the Ross Ice Shelf and the Victoria Plateau. Symposium on Antarctic Research. New Zealand National Committee for the IGY Department of Scientific and Industrial Research: Wellington.

VIALOV, S. S. (1958a). Regularities of glacial shields movement and the theory of plastic viscous flow, Symposium de Chamonix; Physique du mouvement de la glace. International Union of Geodesy and Geophysics, International Association of Scientific Hydrology: Gentbrugge.

VIALOV, S. S. (1958b). Regularities of ice deformation, Symposium de Chamonix. *Ibid*.

VORONOV, K. (1958). K geomorfologii vostochnoy Antarktidy, Informatsionnyy Byulleten' Sovetskoy Antarkticheskoy Ekspeditsii, 1. Izdatel'stvo "Morskoy Transport": Leningrad.

WADE, F. A. (1945). The physical aspects of the Ross Shelf Ice. *Proceedings of the American Philosophical Society* **89**.

WAGNER, G. (1950). Einführung in die Erd-und Landschaftsgeschichte. F. Rau: Öhringen.

WARNER, L. A. (1945). Structure and petrography of the southern Edsel Ford Ranges, Antarctica. Reports on Scientific Results of the United States Antarctic Service Expedition, 1939-41. *Proceedings of the American Philosophical Society*. **89**: 78-122.

WEGENER, A. (1909). Danmark Ekspedition til Grønlands Nordøstkyst 1906-08: Meddelelser om Grønland, Band 42, No. 2.

WEXLER, H. (1958). Some aspects of Antarctic geophysics. *Tellus* **10**: 76-83.

WRIGHT, C. S. (1925). The Ross Barrier and the mechanisms of ice movement. *Geographical Journal* **65**: 198-214.

WRIGHT, C. S., *and* PRIESTLEY, R. E. (1922). Glaciology: British (Terra Nova) Antarctic Expedition 1910-13: London.

MISCELLANEOUS: Some information was gained from news items appearing in the New Zealand Antarctic Society's magazine *Antarctic* and from progress reports given in the IGY supplements of the *Transactions, American Geophysical Union*. The Smithsonian Meteorological Tables, 1951, were used for general reference.

ANARE SCIENTIFIC REPORTS. Series A (IV) GLACIOLOGY, Publication 97. M. Mellor (1967)