BRITISH ANTARCTIC SURVEY

(formerly Falkland Islands Dependencies Survey)

SCIENTIFIC REPORTS

No. 42

GLACIOLOGICAL OBSERVATIONS IN TRINITY PENINSULA AND THE ISLANDS IN PRINCE GUSTAV CHANNEL, GRAHAM LAND, 1958-60

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LONDON: PUBLISHED BY THE BRITISH ANTARCTIC SURVEY: 1964

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(Manuscript received 2nd May, 1961)

ABSTRACT

THIS report discusses glaciological investigations carried out in Trinity Peninsula, Graham Land, during 1958–60. The melt season on a valley glacier and on an ice piedmont is followed in detail and the budgets for each are given. An accurate assessment of the budgets of these ice bodies is hampered by the inaccuracy of measurements of calving and melt-water retention, and variations in meteorological conditions from year to year. However, it is concluded that both ice bodies approximate to a state of equilibrium.

In the southern part of Trinity Peninsula accumulation observations using the pit method and movement studies clearly indicate a decrease in glacierization eastwards. Rime accumulation is important during periods of north-west winds on the west coast, but a strong föhn effect on these winds as they descend the east coast ice piedmont limits accumulation and promotes strong ablation in summer. The result is that the firn line rises from 150 m. a.s.l. on the west side of Trinity Peninsula to between 230 m. and 270 m. a.s.l. on the east side south of Duse Bay, and through 300 m. a.s.l. on the Prince Gustav Channel islands.

Evidence for glacial recession during the present century has only been found in a 75–90 m. retreat of the Depot Glacier snout. Evidence generally suggests there have been no measurable ice-mass changes since the early part of the century. The Larsen Ice Shelf has shown catastrophic calving but this is thought to be a periodic occurrence resulting from extension beyond some point of equilibrium.

There are signs of a more extensive glaciation in the past. In the vicinity of Hope Bay ice extended farther into the bay and was approximately 130–170 m. thicker on land. Evidence in the View Point area and in Prince Gustav Channel suggests a body of ice occupied the channel up to a height more than 300 m. above present sea-level. A series of ice-cored moraines in front of Mount Flora and a series of eight cirques in Prince Gustav Channel with base levels between sea-level and 40 m. a.s.l. suggest at least three stages since the retreat from the glacial maximum.

It is concluded that the most recent recession has resulted from a shift of cyclone tracks southwards and eastwards into the western part of the Weddell Sea.

CONTENTS

I. INTRODUCTION

THIS report describes glaciological work carried out in Trinity Peninsula, Graham Land, during the period December 1957 to April 1960, when the author was senior meteorologist at the Falkland Islands Dependencies Survey station at Hope Bay (lat. 63°24′S., long. 56°59′W.).

The most detailed glaciological investigations were those made in the immediate vicinity of Hope Bay, where accumulation and ablation studies and observations on ice movement were made on the ice piedmont between Hope Bay and Trepassey Bay, and on Depot Glacier (Koerner, 1961). Glaciological work on sledge journeys was mainly confined to investigations of accumulation by the pit method and studies of ice movement on the small valley glacier on Eagle Island, Victory Glacier, Russell East Glacier and the ice piedmont between Victory and Russell East Glaciers. In all these areas of the east coast of Trinity Peninsula a study of the glacial history was made.

Nomenclature

Some of the more important definitions used in this report are listed below. The first two definitions were introduced by Benson (1959) as a result of studies in north-west Greenland.

The dry snow line. The line above which no melt water is formed and therefore no "soaked layers" are formed.

The saturation line. Between the saturation line and the dry snow line melt water does not percolate through more than the current budget year's accumulation.

The firn line. The line separating the area where firn accumulation exceeds ablation, from the ice accumulation area and the ablation zone. Between the firn and saturation lines melt water percolates through more than the current budget year's accumulation. Benson (1959) has referred to these as "soaked facies".

The equilibrium line divides the area of net ice accumulation from the ablation zone. The equilibrium line is the most important datum line on a glacier, since it divides the glacier into areas of net accumulation and net ablation.

The snow line divides the area in which the current budget year's accumulation remains on the surface from the area where it has ablated and exposed glacier ice or a previous budget year's accumulation.

Other terms used in this report have already been defined by J. Smith in *Falkland Islands Dependencies Survey Scientific Report* No. 29.

Methods

All the movement studies were made by simple triangulation methods using a "Tavistock" 5-in. theodolite.

In the vicinity of the Hope Bay station use was made of triangulation points fixed during the Hope Bay local survey carried out by R. R. Kenney in 1955. On Depot Glacier poles below the firn line were intersected from two points on Scar Hills, 407·7 m. apart. Above the firn line on Depot Glacier, on Kenney Glacier and on the ice piedmont between Hope Bay and Trepassey Bay, the poles were resected using the more prominent and well-defined of the 1955 local survey triangulation points.

Ice movement observations during sledge journeys were made in less accurately surveyed areas, where it was necessary to measure a base line on the glacier parallel to the direction of flow. Observations on Depot Glacier are accurate to within 10 sec. of arc, whereas resections on the ice piedmont between Hope Bay and Trepassey Bay are accurate to within 10 sec. of arc for Stakes 3 and 4 and to within 20 sec. of arc for Stake 1. Observations on Russell East Glacier, Victory Glacier, the ice piedmont between these two glaciers, and the valley glacier on Eagle Island are accurate to within 20 sec. of arc.

Glacier snout positions were determined by normal plane-table methods with the use of a telescopic alidade.

Pits never exceeded 5 m. in depth which was the maximum depth from which it was possible to throw snow with a spade. Temperatures in the pits were measured with a "Rototherm" bi-metallic thermometer (accuracy $\pm 1°$ C) as soon as the required level was exposed. Snow or firn samples were taken at the same time with snow samplers similar to the Canadian Snow Survey type, and were weighed on a "Salter" dietary scale to the nearest 1 g. Density values of firn too hard for the sampler to cut into were measured by cutting a small block of firn from the pit wall. The block was then measured and weighed; such densities are subject to an error of ± 10 per cent. Firn and ice layering was recorded and the stratigraphy was measured each time 1 m. of snow or firn had been cleared from the pit.

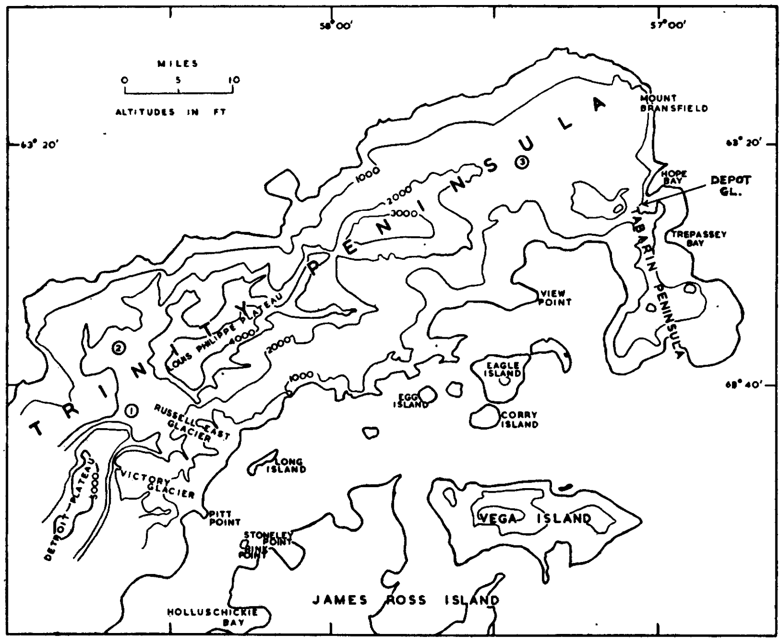

FIGURE 1

Map of part of Trinity Peninsula showing the localities of glaciological observations.

1. Sirius Knoll (Cairn 2, Table VI).
2. Cairn in Marescot Ridge area (Cairn 1, Table VI).
3. Cairn 3 (Table VI).

Contours not shown on James Ross Island.

II. PHYSIOGRAPHY

TRINITY PENINSULA (Fig. 1) lies north-east of a line from Cape Kjellman to Cape Longing and forms the northernmost part of the Graham Land peninsula. It consists basically of east and west coast ice piedmonts which are separated by the plateau. Prince Gustav Channel separates the mainland of Trinity Peninsula from the James Ross Island group.

1. THE PLATEAU

Mount Taylor, which rises to 987 m. and has a steep north-east-facing escarpment, is the north-easternmost outlier of the Graham Land plateau. The true plateau begins north-west of View Point and continues south-westward for about 19 km., reaching a height of a little less than 1,000 m., but it terminates abruptly above Misty Pass. Between Misty Pass and Windy Gap the plateau is more dissected and discontinuous and an outlier forms the southern side of Broad Valley. South-west of Windy Gap the plateau rises gradually to 1,370 m. and terminates above the head of Russell East and Russell West Glaciers. This gap in the plateau is at 830 m. and forms a very level area, bordered on the north-east by the precipitous edge of the Louis-Philippe Plateau and on the south-west by several nunataks rising to 1,000 m. and by the discontinuous edge of the Detroit Plateau. The Detroit Plateau rises quickly to over 1,600 m. south-west of Russell East Glacier and continues uninterrupted as far south as the Marguerite Bay area. The plateau has a very level surface and ends abruptly on its south-east-facing edge in 300 m. ice falls or rock scarps.

2. THE ICE PIEDMONTS

The ice piedmonts of Trinity Peninsula are very distinctive features of the landscape and the one on the east coast, because of its surface regularity particularly above 380 m., provides a regular sledge route as far south as Pitt Point.

a. *The east coast*

North of Pitt Point the ice piedmont is very well defined and reaches a width of 11 km. between the plateau and Prince Gustav Channel. It descends from a height of 800 m. to sea-level in this short distance and its slope steepens rapidly 2 km. from the ice edge before terminating in ice cliffs about 15 m. high. Broad Valley, Russell East Glacier and Victory Glacier flow across the ice piedmont and disturb its continuity, but south of Victory Glacier it becomes more discontinuous and is crossed by ridges which in some cases reach the sea. Between Broad Valley and Russell East Glacier the ice piedmont is interrupted only by occasional nunataks and ridges from the plateau which do not extend into the ice piedmont for more than a quarter of its width. Crevassing is mainly coincident with channelling of the ice, although any irregularities on the surface, presumably caused by irregularities in the bedrock beneath, are usually accompanied by crevassing.

b. *The west coast*

North of Russell West Glacier the ice piedmont is narrower, steeper and less continuous than its eastern counterpart. The west and east coast ice piedmonts coalesce north-west of View Point and the northernmost 15 km. of Trinity Peninsula consist of an ice piedmont of very gentle relief where slopes steepen as the ice edge is approached. Again, channelling causes crevassing. A group of mountains, including Mount Bransfield (765 m.), rises above the general level of the landscape toward the northern tip of Trinity Peninsula.

The ice piedmonts are in many respects similar to Holtedahl's strandflat glaciers which characterize the coastal areas of the Palmer Archipelago (Holtedahl, 1929) and Danco Coast. Bayly (1957) has confirmed Holtedahl's observations on strandflat formation and the dissection of the plateau by head wall erosion of cirque glaciers in the Danco Coast area. However, the origin of strandflat glaciers can hardly be applied to the Trinity Peninsula ice piedmonts, which over the greater part of their area are probably more protective than erosive. The extreme regularity of the plateau margin is in strong opposition to any hypothesis of head wall erosion, since the main agent of erosion produces "a very irregular and angular boundary line between the low foreland and the head wall behind it" (Holtedahl, 1929). Head wall erosion is dependent on sub-aerial agents; the plateau edge is buried beneath an ice fall in many localities but it is exposed in others. Bearing in mind the doubt cast on the rapidity of bergschrund and freeze-thaw erosion by recent researches (Battle, 1960), this variation makes it increasingly unlikely that such a process could be responsible for the headward erosion of the plateau edge more than 10 km. in an even line. A more tenable hypothesis is that the ice piedmonts are a pre-glacial feature.

3. Russell East Glacier

This is a valley glacier (type IV; Ahlmann, 1948) consisting of two main valleys, one of which receives two ice streams from the ice piedmont to the north and then joins the main valley by way of an ice fall. The main ice stream begins in a very broad and level firn area at a height of 762 m. between the Detroit and Louis-Philippe Plateaux. This forms an ice fall as it descends the steep slope north of Mount Canicula and then flows undisturbed for 5 km. before descending another steep slope leading into the main glacier channel. After joining, these two ice streams flow between valley walls, whereas up-glacier the flow is guided largely by sub-surface features which emerge as nunataks in a few areas. The main body of ice is 11 km. long and an average of 1·75 km. wide.

An important valley glacier which is 5 km. long joins the Russell East Glacier network south-east of Mount Canicula but part of this ice flows south-west of Azimuth Hill.

4. Victory Glacier

Victory Glacier, which is 11 km. south-west of Russell East Glacier, is a very broad valley glacier. It is about 14·5 km. in length, 3 km. wide at its narrowest part and 8 km. wide at its widest. At its head the glacier is bounded by the 600 m. high margin of the Detroit Plateau from which snow and ice avalanche down on to the Victory Glacier. In this area, at 610 m., the glacier opens out into a level but steep-sided quasi-amphitheatre. Down-glacier from this the ice is constricted and the valley decreases to half its width with the result that crevasses increase in size and number. Within the next 7 km. three cirques join the valley glacier. 3 km. from Prince Gustav Channel the glacier spreads out and is joined by ice from the piedmont to the north. A rock outcrop, of which Pitt Point forms the seaward part, divides the glacier into two parts one of which flows into the bay south of Pitt Point. The most important outlet forms a snout between areas of slower-moving ice north of Pitt Point. Judging from the amount of calved ice in the sea ice off the northern snout of Victory Glacier in 1959, this must be the most active glacier snout on the east coast of Trinity Peninsula north of Pitt Point.

The cirques have accordant junctions with the valley glacier and the one from the north has eroded back until it is now about 3·8 km. long.

Crevasses are most numerous in the snout areas of this glacier but in its narrow part crevasses 3 km. long and 2·5 m. wide run almost the entire width of the glacier and are spaced at approximately 30 m. intervals.

5. The James Ross Island Group

James Ross Island (approximately 66 km. wide and 72 km. long) is composed of a central ice cap rising to 1,655 m. a.s.l., which drains into the Weddell Sea on the east by a series of steep-sided valley glaciers, and into Admiralty Sound and Prince Gustav Channel to the south and south-west by small valley glaciers and ice falls. The west coast of James Ross Island forms the east coast of Prince Gustav Channel and consists of 60 m. high rock cliffs separated towards the north by three embayments each about 2·5 km. wide. The north-west part of James Ross Island consists of wide tracts of ice-free volcanic rock which bear ample evidence of glacial recession in the form of extensive trains of moraine, many of which are ice-cored. The same features have been found on the coasts of Vega Island.

All the islands of the James Ross Island group are composed of rocks of the James Ross Island Volcanic Group. Of these islands, Beak, Tail, Vortex, Red, Long and Carlson Islands no longer bear any permanent ice cover though most of them show evidence of past ice sculpturing. The tops of Eagle, Corry and Vega Islands are occupied by ice caps thick enough to produce movement. In the case of Vega and Eagle Islands the ice caps drain partly via valley glaciers which terminate at the sea. Egg Island has a very thin, stagnant ice cap.

The valley glacier on Eagle Island is an outlet valley glacier about 2·4 km. long and about 0·8 km. wide. It drains the island's ice cap and its only other source of ice is from a hanging cirque on the west side of the valley. A moraine from the west side of the valley becomes a medial moraine once the cirque ice joins the glacier. The western end of this glacier, down-glacier from its confluence with the cirque, is covered with moraine which in some areas is concentrated into 7 m. high mounds with a steep up-glacier slope standing above a 1 m. wide crevasse.

6. The Ice Piedmont between Hope Bay and Trepassey Bay

Ice cliffs, 30–70 m. high, mark the seaward edge of the ice piedmont, which is morphologically the same as the ice piedmonts described previously (p. 4). Approximately 2,000 m. of the ice margin terminate on land in a slope of 1 in 4·5. This ice piedmont is only crevassed towards the ice edge, in the small cols between The Pyramid and Mount Flora, Mount Carrel and Passes Peak, and where the ice is channelled above Trepassey Bay. Its relatively crevasse-free surface is largely attributable to the sluggish movement and gentle slope of the ice. The ice piedmont often has a surface of bare ice over much of its area, particularly on the north-facing slopes overlooking Hope Bay.

7. Depot Glacier

This fine valley glacier is situated at the head of Hope Bay into which its 20 m. high snout calves. The iceshed at the head of Depot Glacier is 3,430 m. from the snout and is at a height of 350 m. The glacier has an average width of 800–900 m. The lowermost 1,000 m. of the glacier has a very gentle gradient but there is an important break of slope between The Steeple and Blade Ridge where these two ridges approach each other more closely. Between the foot of this slope and its snout, Depot Glacier is joined by two cirque glaciers and Kenney Glacier, a tributary valley glacier 1,800 m. long. There are two very well-defined moraines, a lateral and a medial one. A partly buried moraine, the debris from which frequently falls into the radiation hollow between Scar Hills and Depot Glacier, runs from the confluence of Kenney Glacier and Depot Glacier to the snout.

The cirque between The Steeple and Mount Carrel descends from 366 m. (where it has breached the back wall) to 140 m. in a distance of 900 m., but it has an accordant confluence with the main glacier. This is not so with the west-facing cirque immediately south of Whitten Peak; the floor of this cirque stands about 20 m. above the surface of the main glacier.

III. CLIMATE

TRINITY Peninsula and the islands of Prince Gustav Channel include two climatic regions, the dividing line between the two running along the eastern edge of the plateau. The western area lies in the path of a warm westerly air stream, which on encountering the land of Trinity Peninsula brings cloud and precipitation. The plateau protects the east coast from this air and a drier, colder and less cloudy climate results.

1. TEMPERATURE

The absence of meteorological stations on the west coast of Trinity Peninsula makes a comparison between the west and east coasts difficult. Using Deception Island as representative of the west coast (in fact it is only 60 miles distant) and Hope Bay for the east coast, then the latter area has an average annual temperature about 3° C cooler than at a similar latitude on the west coast. Bransfield Strait, off Trinity Peninsula, is seldom frozen over in winter, and stagnation of air along this coast cannot produce such low temperatures as exist in the temperature inversions of the east coast, where the Weddell Sea current is initially cooler and where in winter sea ice extends out to and sometimes beyond the eastern edge of Tabarin Peninsula, and James Ross and Vega Islands. In addition, the Larsen Ice Shelf forms a favourable area for the institution of temperature inversions farther south.

At Hope Bay there are seldom classic inversions due to the presence of open water either in the bay itself or a mile or so away in Antarctic Sound. Tabarin Peninsula often has a remarkable temperature gradient when inversions occur in Duse Bay, which is usually frozen over for nine months of the year. Vertical gradients of 11° C in 30 m. on the south side of Tabarin Peninsula above Duse Bay are not uncommon on entering an inversion in the winter months and, whereas a calm period at Hope Bay produces a minimum temperature of −20° C, in Duse Bay a temperature between −28° and −35° C is usual.

On the basis of temperature, it may safely be said that the true east coast climate begins in Duse Bay, although Hope Bay and Tabarin Peninsula are far more akin to the east coast than to the west coast in this respect.

Hope Bay experiences its lowest temperatures during south-south-westerly blizzards, whereas south of Tabarin Peninsula the situation is more truly polar in that minimum temperatures coincide with the calm conditions of a temperature inversion. As a result the difference in minimum temperatures between Hope Bay and Duse Bay is approximately 11° C.

The temperature gradient from Hope Bay southward during cyclonic conditions is much smaller than during conditions of temperature inversions, and it is approximately 0·5° C for 10 miles. This is a preliminary figure based on comparisons of temperatures from the Hope Bay station and field parties during southerly blizzards when vertical mixing of air masses is at a maximum, and the figures therefore will be more representative.

Average monthly temperatures rarely rise more than one degree above the freezing point (Fig. 2) and the mildness of a summer depends largely on pressure conditions and the passage of depressions which bring in $_{t}$mT air, without which the average would remain below 0° C. The föhn effect during periods of westerly $_{t}$mT air raises the temperature an additional 3° C and a few days of these winds will raise the monthly average considerably, particularly in winter.

The onset of winter is sharp, coming at the end of March or very early April. At this time of the year the temperature gradient between the east and west coasts must be at its greatest and the year's minimum temperature at Hope Bay may occur as early as May.

Although the annual range of temperature in the Hope Bay area is never as great as in the highly continental Arctic climates of Canada and Siberia (Pepper, 1954), the daily range may be very high in winter. Maximum and minimum temperatures for 1959 showed a daily range of 22° C, and 15° C changes

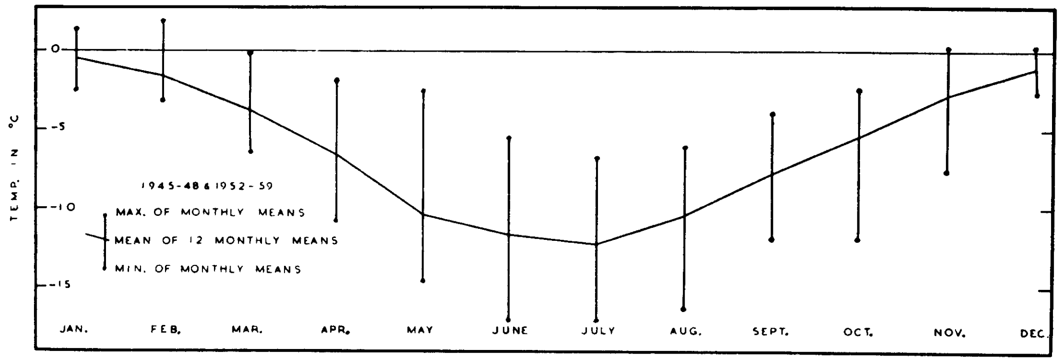

FIGURE 2
Monthly mean temperatures for Hope Bay, lat. 63°24′S., long. 56°59′W.

of temperature occurring with the passage of a front are not unusual. This range is accentuated by the föhn effect on air descending from plateau level. The division of air masses at fronts is sharp and the passage of a front in winter may bring a change in temperature of over 6° C in less than 15 min. This range is much less pronounced in summer when increased insolation raises the temperature of Antarctic air and, although high maximum temperatures do not show a marked rise except in their frequency, minimum temperatures do. The maximum temperature in winter usually rises above 4° C in every month and the occasional outbreak of mT air, which reaches the east coast as a dry föhn wind, e.g. end of June 1959, brings temperatures within a few degrees of maximum summer temperatures. The record maximum temperature for the Antarctic continent of +14° C was recorded at the Hope Bay station in the same month (July) in 1958 as that station's minimum temperature of −30° C was recorded.

These high föhn temperatures have been recorded along the whole of the east coast of Trinity Peninsula, Prince Gustav Channel and the offshore islands.

The variability of temperatures from year to year, as represented by monthly averages (Fig. 2), is most pronounced in winter but, as summer averages are so close to 0° C, variations then are the most important glaciologically. November shows a very marked variability and is surpassed only by June and July. December and January show the least variability and the latter is usually the warmest month.

2. PRECIPITATION

The exact amount of precipitation for this area is unknown due to the impossibility of measuring snowfall in a windswept area where heavy drift in southerly blizzards coincides with snowfall. An examination has been made of the average number of snow days/month over a period of 6 yr. but, since a snow day bears no relation to the amount of snowfall during that day, it has proved fruitless. Observation suggests that precipitation occurs principally at the equinoxes, particularly in spring. Rain falls on approximately 20 days/yr.

The highest rate of snowfall is during northerly winds (Pepper, 1954) belonging to the warm sector of depressions moving up the west coast of Graham Land. Probably the greatest amount of snow during the year is brought by south-westerly winds, particularly those close to and behind a cold front. Snowfall in the latter case, however, is usually accompanied by high wind speeds and consequently little is added to the accumulation, whereas snowfall in northerly air streams nearly always comes with winds less than Force 8.

On the east coast of Trinity Peninsula arien deposits are only occasional phenomena. There are few clear nights for hoar frost development and, as at sea-level there are only 20–30 days of true fog in a year, rime or fog deposit is unimportant. Rime and fog deposits on the west coast of Trinity Peninsula and Joinville Island are very common phenomena and observations by sledge parties suggest that certainly above 200 m. one or the other must occur on approximately 75 per cent of the days in a year.

3. WIND

In the Hope Bay area the dominant wind direction lies in the south-western quadrant, and on the east coast mainland to the south there is an increase in the frequency of westerly and north-westerly winds.

Less is known of the west coast of Trinity Peninsula but observations by sledge parties suggest that north-west is an important wind component.

At Hope Bay the dominant wind is the cold south-south-westerly, which is extremely violent and frequently reaches storm force. It appears to increase in violence on the lee side of Tabarin Peninsula and, although funnelling explains the high speeds down Depot Glacier, it cannot explain its violence at the top of the ice piedmont between Hope Bay and Trepassey Bay, where this wind is far stronger than across the immediate vicinity of the station. This wind seldom exceeds strong gale force south of Tabarin Peninsula, except through areas of funnelling like Admiralty Sound (Nordenskjöld and Andersson, 1905), although it still blows from a similar direction.

Because of its dryness and warmth, the westerly föhn wind is glaciologically one of the most important meteorological factors of the east side of Trinity Peninsula. During periods of föhn winds the moist wind on the west coast is from the north-west and the föhn bank generally lies at about 500–600 m. below the plateau gaps and also along the eastern edge of the plateau. From this limit to sea-level it descends as dry air and usually exceeds strong gale force often dying out before reaching James Ross Island where, however, the temperature is similar to that within the föhn wind. Relative humidities as low as 5 per cent have been recorded (Pepper, 1954) but a more common figure is 50–75 per cent. Wind speeds recorded at the Hope Bay station are very irregular and gusty. The wind becomes more powerful on the ice piedmont between Hope Bay and Trepassey Bay where mean wind velocities of 20–25 m./sec. are typical for a föhn wind. It is on the east coast of Trinity Peninsula, however, that the westerly föhn wind reaches its greatest velocity. Both Russell East and Victory Glaciers form areas of local funnelling and föhn winds reach hurricane force during the passage of a depression in the Weddell Sea. On the east coast ice piedmont the föhn wind is less violent than this but still it frequently exceeds storm force in areas where there is no funnelling effect. This wind is most common in summer when depression tracks move southward, and it is associated with low pressure in the Weddell Sea area.

IV. DEPOT GLACIER

THE ratio of the accumulation area to the ablation area over the whole glacier network is 7 : 5. Table III (p. 21) gives the areas of individual budget regions.

Depot Glacier is subject to winds of a violent nature which, judging from records and personal comparative experiences, frequently, particularly at the equinoxes, reach hurricane force with fairly common mean wind speeds of 30 m./sec. This wind, which usually belongs to the cold sector of a depression, is low in temperature and as a result deflates the snow cover in all but the lee accumulation areas. Snow depth on the glacier rarely exceeds 1 m., apart from the firn accumulation areas, and bare ice may appear at any time of the year subsequent to the deflating action of a south-south-west blizzard. The westerly föhn wind is also funnelled down the glacier and with temperatures close to 0° C ice crystal drift can result after recrystallization of the snow has taken place.

1. SURFACE FEATURES

a. *Moraines*

The glacier has two principal moraines (Fig. 3) consisting mainly of cobbles and boulders with occasional blocks, the largest measuring about 3 m. by 4 m. by 1·5 m. Frost-shatter disposes of the majority of blocks, and one which measured 2·5 m. by 4 m. by 1·5 m. was so shattered by frost action in one year that the largest remnant measured 1·5 m. by 2·5 m. by 1 m.

These moraines are superficial collections of frost-shattered rock rolling down the valley sides and, as they are formed in the ablation region, they remain superficial except for an area opposite the Kenney Glacier where the moraine, for reasons undiscovered, disappears beneath the surface for 330 m. Crevasses and the evidence at the snout present good cross-sections of these moraines, which consist of little more than a single layer of rock resting on the ice. The more bulky western moraine stands 3–4 m. above the surrounding ice surface and the eastern one about 2·5 m. Though there is ice protection, the low altitude of the moraines above the surrounding ice supports the stake measurements in showing the low order of ice surface lowering in the ablation season.

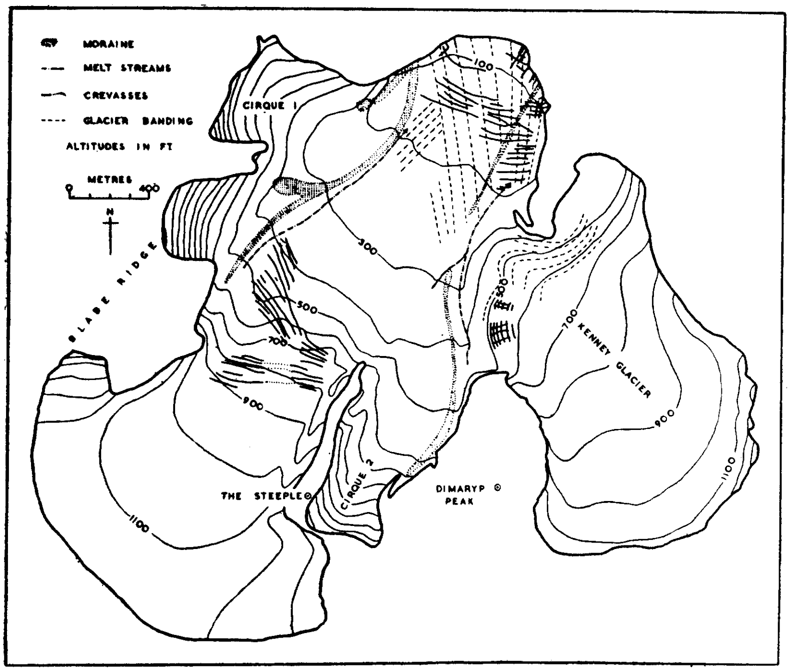

FIGURE 3
The morphology of Depot Glacier.
(Map by R R. Kenney (1955) with minor local modifications.)

The glacier snout, due no doubt to a lee effect, usually has a thin snow cover and the medial moraine is only on the surface at the snout during the height of the ablation season. In 1955, when the glacier was surveyed, it ended superficially 200 m. from the snout.

b. *Melt streams*

No slusher streams were observed on this glacier network and this is believed to be due to the presence of crevasses which take the excess melt water. At no time were snow surfaces as saturated as they were on the ice piedmont slope behind the Hope Bay station. The ice forming this slope has a considerable admixture of dust and pebbles which greatly reduces its albedo and this, together with its north-westerly aspect, increases the amount of melt water which can accumulate on the surface due to the absence of crevasses.

Melt streams formed twice during the summer of 1959–60 and their courses are shown in Fig. 3. The Kenney Glacier never bears melt streams, no doubt because of the presence of crevasses parallel to the contours of the slope.

c. *Glacier bands*

Fig. 3 shows the banding on the surface of Depot Glacier. No explanation has been formulated for the orientation of these bands which would appear in cross-section to be accumulation layers exposed at the surface as an alternation of clear blue and white opaque ice. The bands on the steep forward slope of Kenney Glacier follow the contours.

No ogives are evident on this glacier and in fact none at all have been seen on any of the numerous glaciers in the Trinity Peninsula area. A low firn line may obscure any that form at the base of ice falls in the accumulation regions but ice falls observed below the firn line show very little distortion or waving in the ice at their bases.

d. *Ice ridges*

A detailed cross-section of the surface of Depot Glacier would show two gentle ice ridges rising 7 m. above the surrounding ice surface. The extent of the more prominent of these was plane-tabled and its alignment with the ridge surmounted by The Steeple indicates that it is the surface expression of a rise in the valley bed. Two other valley glaciers, Russell East and Victory Glaciers, which are 64 and 80 km. south of Hope Bay respectively, show similar features though these are accompanied by crevassing of the surface. The crevasses are particularly severe on Victory Glacier. In each case the ice surface ridge is aligned with a nunatak ridge and it seems likely that the ridge continues under the ice and is sufficiently pronounced to cause a reflection of its form on the surface. It serves to indicate that on these valley glaciers it is most unlikely that the cross-section has a simple U-shape.

e. *Crevasse areas*

Depot Glacier has large completely crevasse-free areas which are largely a reflection of the low degree of activity. However, there are three main crevasse groups on the glacier network (Fig. 3): one in the vicinity of the snout, a second on the steep slope of Kenney Glacier overlooking Depot Glacier and a third on the steep slope of the main trunk glacier between The Steeple and Blade Ridge.

The snout area. Crevasses begin 350–400 m. from the snout and reach a width of 2 m. until an area close to the snout itself, where they become much wider. This set of crevasses drains away the surface melt water. Crevassing is most pronounced on the Scar Hills (east) side of the glacier, which is the only area where crevasses parallel to the direction of flow of the glacier become important.

Kenney Glacier. The crevasse area approaches the form of an ice fall and is probably due to the nearness of the valley floor to the glacier surface. This tributary glacier has a series of crevasses 60 cm. wide following the contours of the slope. Many crevasses were filled by re-frozen melt water at the end of the ablation season.

The slope between Blade Ridge and The Steeple. This is subdivided into two subsidiary groups basically aligned along breaks of slope. Individual crevasses reach a width of 10 m. The whole area is one where the ice undergoes much stress and strain as Blade Ridge thrusts a spur into the ice stream, which falls 170 m. in a distance of 700 m.

2. Regime

a. *Firn accumulation*

The main accumulation area forms a part of the ice piedmont which covers most of Tabarin Peninsula. Although no pits were dug above the regional climatic firn line on Depot Glacier, some were dug on Tabarin Peninsula at similar heights. These gave an average annual accumulation ranging from 10 to 20 cm. water equivalent.

The isolated accumulation area below the regional firn line (Fig. 4) lies on a steep slope below an area of maximum wind funnelling between the southern end of Blade Ridge and the ridge of which The Steeple forms the highest part. The carrying power of the glacier's most important and dominant wind, the south-south-westerly one, drops as the funnelling effect decreases and the slope changes abruptly, and surplus snow is deposited. The annual accumulation per unit area in this budget region is more than twice that of the area above the regional firn line and its contribution to the positive side of the budget in 1959–60 was 84 per cent of that in the main accumulation area (Table III). The lee firn accumulation area also lowers the altitude of the equilibrium line to approximately 75 m., because of the formation of superimposed ice from the freezing of melt water running from the firn at the end of the ablation season. Kenney Glacier has no such lee accumulation area and the equilibrium line is therefore as high as 270 m.

Realizing its importance, a pit was dug in the lee firn area in January 1960 to determine more precisely its contribution to the glacier budget (Fig. 5). Calculation of the annual accumulation from stratification in the pit was difficult due to possible percolation of melt water through more than the current year's accumulation, but summer ice layers were quite easily distinguished. The snow level at Stake 15b (124 m. a.s.l.) fell by 137·0 cm. and as the firn was already of high density very little further compaction or retention of melt water might have been expected. This loss of snow and firn with an average density of 0·65 g./cm.^{3} leaves a net accumulation for the budget year 1959–60 of 48·4 cm. of water. This figure should be above the true value as percolation of melt water at the end of the ablation season more than compensated for any increase in firn density after the time the pit was dug in mid-January. This figure compares with 65·0 cm. for 1958–59, 57·0 cm. for 1957–58 and 47·2 cm. for 1956–57.

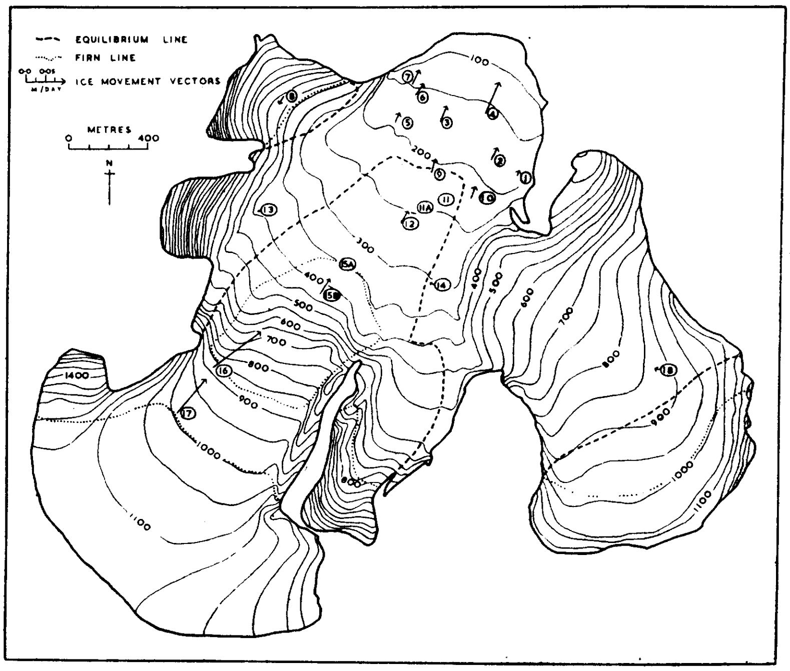

FIGURE 4
Budget regions and ice movement for Depot Glacier, and cirques 1 and 2 (Table III).
Altitudes in ft.
(Map by R. R. Kenney (1955) with minor local modifications.)
Movement rates:

Stake 1	0·026 m./day*	Stake 10	0·055 m./day
Stake 2	0·072 m./day*	Stake 12	0·076 m./day*
Stake 3	0·085 m./day*	Stake 13	0·0099 m./day
Stake 4	0·165 m./day	Stake 14	0·017 m./day*
Stake 5	0·056 m./day*	Stake 15b	0·090 m./day
Stake 6	0·073 m./day	Stake 16	0·298 m./day
Stake 7	0·076 m./day	Stake 17	0·224 m./day
Stake 8	0·047 m./day	Stake 18	0·018 m./day
Stake 9	0·072 m./day		

All the above values are for summer measurements except those marked with an asterisk where values are for measurements over a complete year.

These accumulation figures can be correlated with the summer conditions as represented by monthly average temperatures and the deviation from the mean of 12 yr. monthly averages (Fig. 2). The budget year accumulations for 1956–57 and 1959–60 are comparatively low and in each of the ablation periods of these budget years the average temperature for February rose 6·1° C and 2·0° C respectively above the mean of the averages for this month. The highest accumulation figure in these four budget years is for the budget year ending with the summer of 1958–59. The monthly average temperature for February 1959 fell 2·7° C below the 12 yr. mean as the result of a two-week period under the influence of the south-south-west wind; the November 1958 temperature was 5·6° C above the mean. The budget year ending with the summer of 1957–58 had an accumulation close to the mean of the four years' values. In that budget year the ablation period consisted of a warm November (2·9° C above the mean) and average mid-summer and autumn temperatures.

The lowest variation in mean monthly temperatures is in December and January which are usually the warmest months. Variations in temperatures from one ablation season to another are most pronounced in spring and autumn and from the above analysis of accumulation data it would appear that the meteorological conditions of late summer and autumn rather than spring are of vital importance to the state of the budget.

When Pit 1 was dug in mid-January 1960 the entire winter snow accumulation was wet and high snow densities were recorded, reaching a maximum of 0·7 g./cm.³ at a depth of 1·22 m. The firn of the previous

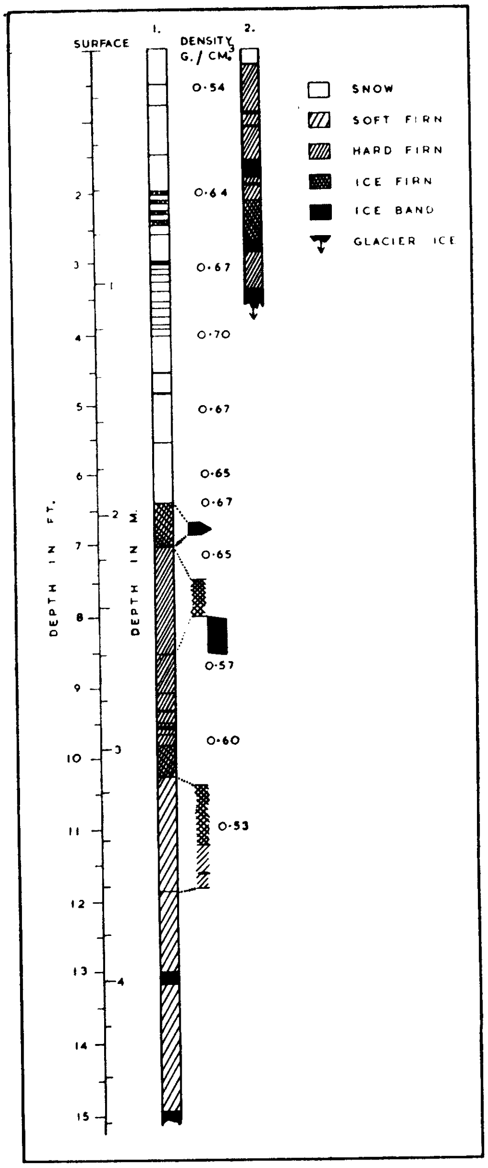

FIGURE 5
Pit profiles on Depot Glacier.
Pit 1. Lee accumulation area, Depot Glacier.
Pit 2. Cirque 1.

two budget years had an average firn density of 0·57 g./cm.³. At the end of the ablation season penetration of the cold wave freezes the snow surface layers and formation of melt water ceases. Melt water in the underlying firn continues to percolate through and run along the surface of impervious ice layers. This consequently reduces the firn density for all except the ablation surface layer.

The other firn accumulation areas of the Depot Glacier network are in the upper reaches of the two cirques and Kenney Glacier. The cirque below Whitten Peak has a very small firn area but the annual accumulation must be considerable as it is supplied by avalanched snow from the cirque walls; snow avalanche fans are often present. A pit dug in this cirque at 140 m. a.s.l. uncovered the layers shown in Fig. 5. Glacier ice was 1·5 m. below the surface and drilling gave a similar depth of firn and snow cover to a distance of 150 m. up the cirque from this point. It was impossible to identify ice layers so close to the firn line when the elevation of the firn line above this point in any one year can remove more than the current year's accumulation. Pit 2 indicates that the climatic firn line lies somewhere below the site of the pit, which places it at approximately 120 m. a.s.l. The shaded position of the cirque and the supply of avalanched snow have brought about this local lowering of the climatic firn line. There is a similar situation in the cirque between Mount Carrel and The Steeple, although here the walls of the cirque are steeper and, as there is very little head wall to the cirque, there is hardly any avalanched snow. Its lee and shaded position, however, favours a local lowering of the climatic firn line which is at approximately 150 m. a.s.l.

b. *Ablation and superimposed ice*

Measurements of ablation were made by fairly routine methods. Holes were drilled and then stakes were hammered into the glacier ice to a depth of 1 m. Measurements of the ice surface relative to the stake were taken whenever possible. Ablation, wind and the differential heating of the surface of the metal stake caused tilting and at any sign of this they were reset and remeasured. Six stakes were set up in February 1959 and the snow and ice level was measured on each. Absence from the Hope Bay station on sledge journeys excluded measurements more than once or twice between the 1958 and 1959 summers, by which time four stakes were missing. One of them was almost certainly buried and two had been snapped off by wind. The stakes were replaced in November 1959 and observations were restarted. The stake measurements (Table I) are represented in histogram form in Fig. 6, from which it is clear that

TABLE I

BUDGET OF ICE LEVEL ON DEPOT GLACIER
(Stake Measurements, November 1959–April 1960)

Stake Number	*Height* (m. a.s.l.)	*Loss* (cm. of ice)	*Gain by Superimposed Ice* (cm. of ice)	*Net Accumulation or Ablation* (cm. of ice)
1	56	53	0	−53
3	53	68	20	−48
11	69	33	31	+ 2
12	82	—	—	at least + 7
18	265	15	5	−10

The ice datum level for Stake 3 is February 1959.
The ice datum level for the other stakes is 24 December 1959 and the negative side of the ice budget is accordingly exaggerated.

substantial snowfalls are liable to interrupt the ablation season which in Trinity Peninsula is rarely continuous over more than a week at a time. Ahlmann (1948) states that temperate glaciers can retain hardly more than 50 per cent of the fluid water present during the ablation season and on Isachsen's Plateau in Spitsbergen the retention figure is only 30 per cent. On sub-polar glaciers 100 per cent of the melt water may be retained. It is suspected that on Depot Glacier not more than 10 per cent of the melt water leaves the glacier in a fluid state and therefore the role of evaporation in ablation may be very important.

i. *Summer 1958–59, winter 1959.* The continuity of measurements between the summers of 1958–59 and 1959–60 is based on readings at Stakes 1, 3 and 12 which remained in the glacier throughout the winter of 1959. The ice level was measured on Stakes 1 and 3 in February 1959. In that month the glacier had a continuous snow cover right to the ice front. The depth of this cover varied between 40·0 and 62·0 cm. and the mean of four stake readings was 53·0 cm. Föhn winds in the middle of March 1959 lowered the snow level and it was further ablated in some localities to bare ice by a violent south-south-west blizzard at the beginning of what would normally be termed the winter accumulation period. In July the snow cover below the firn line was 45·0 cm. thick at 53 m. a.s.l., 10·7 cm. at 65 m. a.s.l. and 22·0 cm. at 82 m. a.s.l. The absence of correlation between the altitude and accumulation is a consequence of drifting. By November the snow cover at Stake 3 (53 m. a.s.l.) was 12·7 cm. but the ice level at the stake was

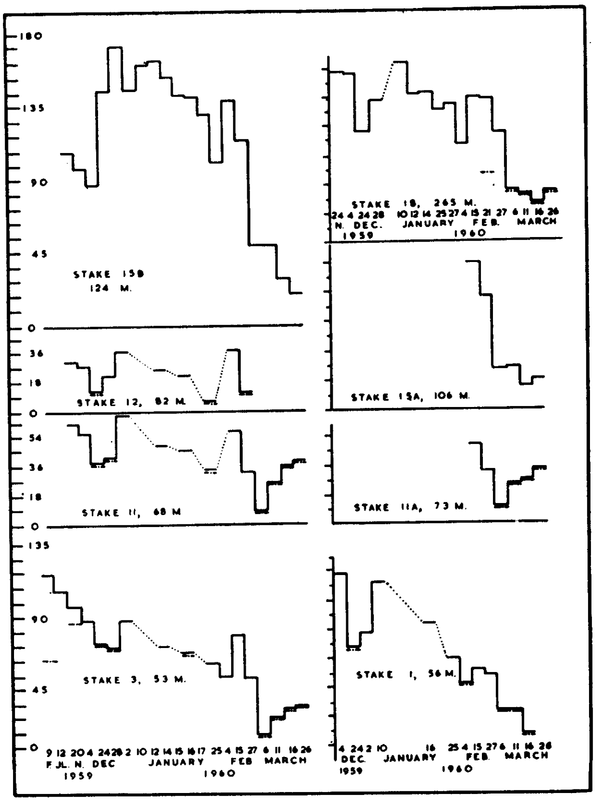

FIGURE 6
Accumulation and ablation measurements for Depot Glacier, February 1959–March 1960. Ordinate scale in cm. All the changing snow and ice levels are relative to the initial measurement in the first bar of each graph. The dash-dot-dash line is the level of superimposed or glacier ice, whether at the surface or under a snow cover.

found to have increased by 18·5 cm. between February and 20 November 1959. The warm period in March 1959 followed a period of accumulation when temperatures were not more than a few degrees below freezing point. The ice surface temperature could not have been very strongly negative and very little superimposed ice could have formed. On 21 November 1959 the temperature at a depth of 15·0 cm. at Stake 15b (124 m. a.s.l.) was −6·0° C and at Stake 3 the ice was covered with firn in which firnification was well advanced. This cover was by then only 12·7 cm. deep and the conditions were clearly favourable for the formation of superimposed ice. The figure of 18·5 cm. is, however, high in comparison with the theoretical figures calculated by Ward and Orvig (1953, p. 165) for the superimposition of ice, given a certain length of melt period and a particular initial ice surface temperature. It seems likely, therefore, that an important thickness of superimposed ice forms during spells of warm föhn weather in winter, mainly when the glacier tongue has a thin snow cover.

ii. *4 December 1959–10 January 1960.* By 24 December ablation had reduced the glacier to bare ice below 85 m. a.s.l. At Stake 3 (53 m. a.s.l.) 10·0 cm. of superimposed ice were ablated and at Stake 1 the ice level on 24 December was 15·0 cm. lower than the February 1959 one. This period of ablation was followed by the first important snowfall of the summer, which arrested ablation for ten days, and the snow mantle it provided covered the ice below 70 m. a.s.l. for sixteen days. Moderate snowfalls on 3, 6, 7 and 8 January added 28·0 cm. of snow above the 115 m. firn line of the lee accumulation area and 25·0 cm. below it (mean of four stake readings). Snow depths showed wide variation from one stake to another, varying between 19·0 and 34·0 cm. The lee accumulation area did not receive a higher accumulation, because the snowfall was accompanied by very little drift which is that area's main source of accumulation.

iii. *10–31 January 1960.* For a period of one and a half weeks after 10 January there were only 2–3 freeze degrees, whereas the minimum number of thaw degrees was 15 on 14 January and the maximum number was 74 on 19 January. The temperature dropped below freezing point only on 15, 16 and 18 January and then only for a few hours (Fig. 7). Stake readings were taken every other day during the first half of the period while camping on the glacier and the second half of the period is represented

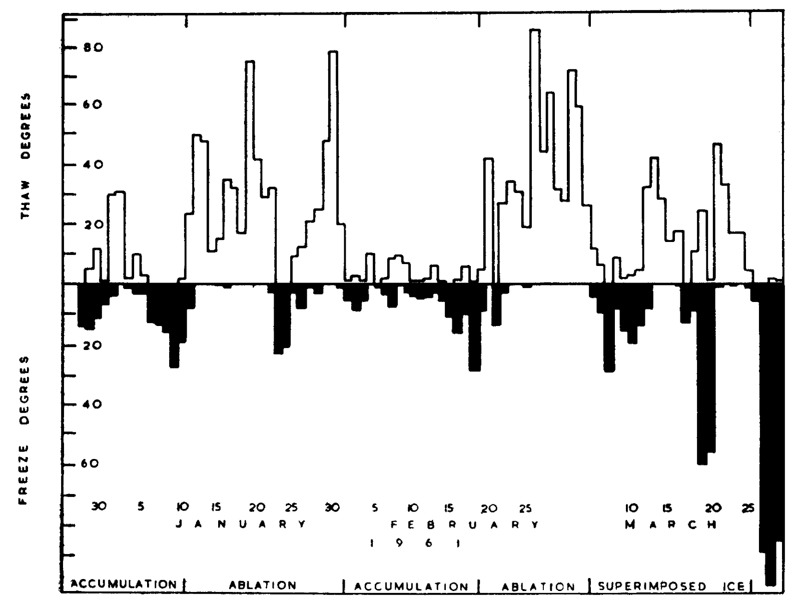

FIGURE 7

Average daily freeze and thaw degrees, January–March 1960, for Hope Bay. 1 thaw degree = 2 hr. × 1·1° C above 0° C, i.e. the temperature remaining at 4·4° C for 4 hr. is equivalent to 8 thaw degrees. 1 freeze degree = 2 hr. × 1·1° C below 0° C.

by stake readings taken on 25 January. By 17 January the snow line had risen to 70 m. a.s.l. and by 25 January it was at approximately 90 m. a.s.l. The ice surface was lowered by 7·0 cm. at 82 m. a.s.l., 4·0 cm. at 69 m. a.s.l. and 7·5–8·9 cm. between 45 and 53 m. a.s.l.

A three-day spell of cold south-south-west winds with snow from 23–26 January slowed ablation which was re-invigorated by the incursion of warm air on 27 January. By 4 February the area of Depot Glacier below 110 m. a.s.l. was reduced to its lowest level up to that date in the ablation season. At Stake 3 (53 m. a.s.l.) all the superimposed ice formed in February–November 1959 was removed.

iv. *1–20 February 1960.* Accumulation recommenced with moderate snowfall at near freezing temperatures on 6 and 7 February and slight snow until 14 February. When Depot Glacier was revisited on 15 February the ice was covered by 9·0–32·0 cm. of snow below 110 m. a.s.l., and 39·0 cm. of snow had accumulated at Stake 15b (124 m. a.s.l.) and 29·2 cm. at 265 m. a.s.l. on Kenney Glacier. This was the last time that the glacier bore a snow cover until the beginning of the accumulation season.

v. *21 February–5 March 1960.* This interval constituted the main ablation period of the budget year. During eleven days the temperature dropped below freezing point for only 1 hr. Relative humidities were generally about 70 per cent, dropping occasionally to 50 per cent. By 6 March the ice surface above 50 m. a.s.l. reached its lowest level and just above the equilibrium line at Stake 11 it was lowered by 25·4 cm. after the 23·0 cm. snow cover had ablated away. Below the equilibrium line, at Stakes 1 and 3, 35·6 and 33·0 cm. of ice were removed. On Kenney Glacier, at 265 m. a.s.l., ice was exposed for the first occasion by 6 March, by which time 10·0 cm. of ice had been ablated. In this period the snow level fell by 91·4 cm. at Stake 15b in the lee accumulation area and by 63·5 cm. just above the firn line.

vi. *5–26 March 1960.* The approach of winter was indicated by alternating cold and warm weather which favoured the formation of superimposed ice below the firn line. The presence of crevasses about 1 m. up-slope from Stake 1 prevented any melt water running on to that area. Stake 1 therefore showed a continuous lowering of ice level. An 8 cm. wide crevasse above Stake 3 filled with ice during the ablation season so that melt water from higher up the glacier froze on the surface around Stake 3 in the latter half of the summer. No open crevasses were seen near Stakes 11, 11a, 12 and 15a. Melt water issuing from the lee firn area started to freeze between 6 and 11 March. By the latter date an average of 24·0 cm. had formed on the glacier tongue, some of it from the freezing of melted snow and rain. Ice continued to build up on the glacier tongue until 26 March and much of it resulted from rain and snow melt water which had frozen in the cold sector weather of 19, 20 and 26 March. The ice surface at the end of this period was exceptionally smooth and it was impossible to walk on without crampons.

An ice layer also formed on the firn surface at Stake 15a (117 m. a.s.l.), almost on the firn line, whereas the snow level at Stake 15b (124 m. a.s.l.) continued to fall up to 26 March. The level at Stake 15a always showed a smaller drop than Stake 15b, probably because of the greater density of firn at the lower level resulting from melt-water accumulation. Snow accumulation at the beginning of the winter period was deflated by a south-south-west blizzard at the end of the first week in April.

The Kenney Glacier tributary differs markedly from Depot Glacier in one important respect—aspect. It does not receive the full impact of the south-south-west wind which has a less deflating effect, but in summer this is counteracted by its north-west-facing aspect. At 275 m. a.s.l. the glacier ice was firn-covered until 11 February 1960 up to which time more than one year's accumulation lay on the surface (48·0 cm. of firn at one point near the stake and 60·0 cm. at another 20 m. above). Ice was exposed some time between 27 February and 6 March and in the next five days there was a continued ablation of 1·2 cm. of ice while superimposed ice was forming at Stake 4 on Depot Glacier which is farther from a firn area. Superimposition of ice did not exceed ice ablation until some time after 16 March. With no lee firn areas, the ablation area of Kenney Glacier is twice as large as its accumulation area, whereas on Depot Glacier itself the accumulation area is more than twice as large as the ablation area. Kenney Glacier is therefore the most recessive part of the Depot Glacier network.

The area of bare ice below the main accumulation area on Depot Glacier was relatively inaccessible from the Hope Bay station and no accurate measurements of superimposition of ice were made. From isolated and somewhat discontinuous stake readings it appears that it is at least as much as on the area below the lee firn area.

Stake 12 was measured regularly until 27 February, after which time the final figure is an extrapolation based on Stakes 11 and 11a, the latter between Stakes 11 and 12, which was measured regularly from 15 February onwards.

3. Movement

The Depot Glacier area had been surveyed by R. R. Kenney in 1955 which meant that intersections and resections could be observed using his triangulation points. Stakes which could be intersected accurately were observed using two of Kenney's occupied stations on Scar Hills, and all other stake positions were fixed by theodolite resections.

The major problem in connection with the movement observation scheme was the loss of stakes due to ablation, burial by snow, snapping off in a wind, collapse of moraine cairns due to frost-shatter and tilting. For the first four reasons, readings were very discontinuous over the first year (1959) and the main value of the results is in the comparison of average velocities taken over a period greater than two months and in some cases more than a year.

An attempt was made to determine the rate of increase in movement in summer by observing at fortnightly intervals but the collapse of cairns and the overall inaccuracy of the observations (the theodolite used at the time, personal error and the poorly balanced triangles involved) rendered many results invalid.

Average movement has been determined for 18 points on the glacier including a transverse profile and a longitudinal profile (Fig. 4). The longitudinal profile was not set ideally because of the time involved in observing all safety precautions in a crevassed area. Points of interest on these averages are:

i. Movement on the transverse profile increases in speed towards the centre of the glacier.
ii. The rate of movement increases as the snout is approached.
iii. Movement is perpendicular to the contours in all cases except in the cirque below Whitten Peak and where the moraine disappears under the ice surface opposite Kenney Glacier.

The lateral moraines show the line of direction of ice movement and both bend in towards the centre of the valley as the cirque and Kenney Glacier tributaries join the main channel. Towards the snout the moraines and all but one of the stakes diverge from one another and an increase in crevassing results, although there are hardly any longitudinal crevasses. Almost all the crevasses are nearly parallel to the contours.

The maximum movement measured was at 300 m. a.s.l. on the top of the steep slope forming the lee accumulation area, where movement was 0·298 m./day, and above this at 330 m. a.s.l., below the regional firn line, movement was 0·224 m./day (both are summer measurements). This pattern is comparable with measurements on North Atlantic coast glaciers where movement was found to be at a high rate towards the head of some glaciers (Ahlmann, 1948).

Movement at a position 70 m. from the rock wall is low and Stake 13 towards the centre of the north-west lateral moraine moved at only 0·0099 m./day despite the presence of a shallow ice-filled cirque behind it. This is very suggestive of thin ice and hence a shallow form of cross-section.

The moraine stake, Stake 14, opposite Kenney Glacier showed a low order of movement (0·017 m./day) and in a direction in line with the forward movement of Kenney Glacier; this stake does not coincide with the 1955 survey of the position of the moraine.

Stake 18, the only one with values for movement on Kenney Glacier was situated at a height of 290 m. a.s.l., towards the centre of the channelled flow, and yet the movement was slow and, assuming that in two areas of similar slope differences of movement depend largely on differences in the thicknesses of the ice, then it is clear that Kenney Glacier is considerably thinner than Depot Glacier above its tributary confluences. This agrees well with the hypothesis that this glacier is receding at a rate greater than that of any other part of the glacier network.

The thickness of the ice shown on the profiles in Figs. 8 and 9 has been computed using the measured

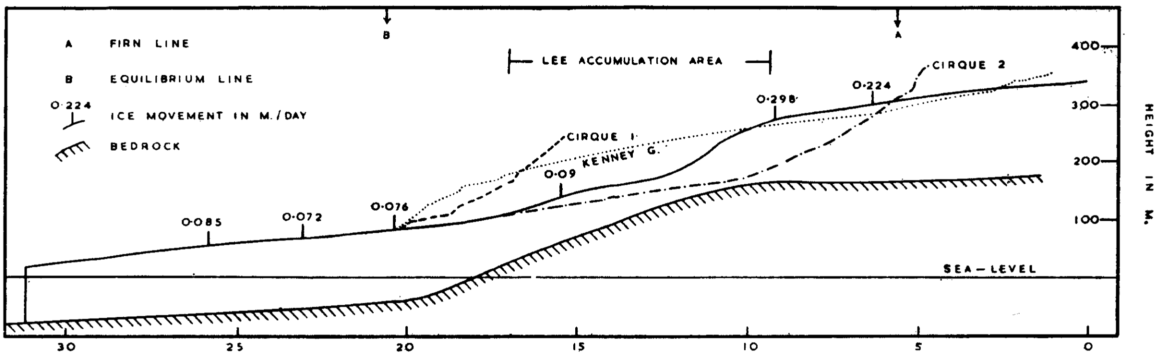

FIGURE 8
Surface and bedrock longitudinal profiles of Depot Glacier.

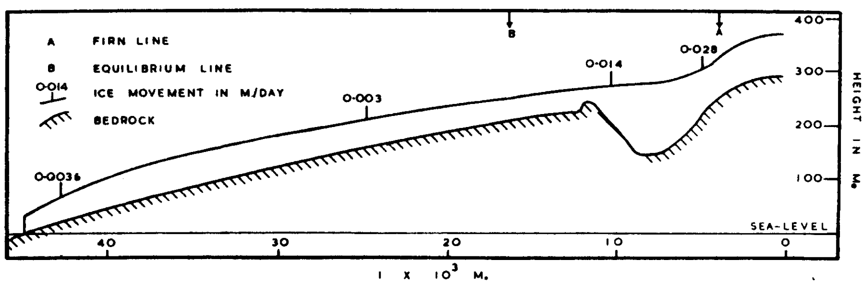

FIGURE 9
Surface and bedrock longitudinal profiles of the ice piedmont between Hope Bay and Trepassey Bay.

velocities of ice flow and the angle of slope of the glacier surface at the point of the measured velocity.

From experimental work on the Jungfraujoch, Switzerland, Perutz (1950) found his results fitted the formula

$$\frac{dy}{dt} = \frac{\tau^{2\cdot 3}}{550},$$

where y = shear strain and τ = shear stress.

For a velocity gradient of dv/dt the rate of change of shear strain with time, $d\gamma/dt$, can be expressed as $d\gamma/dt = dv/dh$.

Therefore, $$\frac{dv}{dh} = \frac{1}{550} \cdot \left\{\frac{phg \sin \alpha}{10^6}\right\}^{2\cdot 3},$$

since $\tau = pgh \sin \alpha$, and where
p = density of ice in g./cm.^{3} (= 0·9 g./cm.^{3}),
g = acceleration due to gravity ($\sim$ 1,000 cm./sec.^{2}),
h = ice thickness in cm.,
α = angle of slope of ice surface,
v = velocity in cm./sec.

Putting $A = \frac{1}{550\,(10^6)^{2\cdot 3}} \cdot (pg \sin \alpha)^{2\cdot 3}$ and integrating from 0 to h, then

$$v = A \int_0^h h^{2\cdot 3}\, dh = A\,\frac{h^{3\cdot 3}}{3\cdot 3} + \text{constant.}$$

When $h = 0$, $v = 0$, therefore the constant = 0, and $h = 1\cdot 435 \left(\frac{v}{A}\right)^{\frac{1}{3\cdot 3}}$.

From this the following ice thicknesses have been calculated:

Depot Glacier	Stake 3	144 m.
	Stake 15b	67 m.
	Stake 17	157 m.
Kenney Glacier	Stake 18	64 m.
Ice piedmont between Hope Bay and Trepassey Bay	Stake 4	83 m.
	Stake 3	139 m.
	Stake 2	$\sim$ 48 m.
Eagle Island	Average	$\sim$ 70 m.

a. *Periodic variation*

A series of fixes was taken on a bamboo stake, Stake 6, on the north-west side of the glacier and the resultant triangle was better balanced than others observed over this period (December 1958–March 1959). Movement was found to be far from regular and the rate over the eleven days between 5 and 16 January 1959 was 136 per cent the rate between 16 January and 1 February 1959 (0·077 m./day and 0·049 m./day, respectively).

The more extensive measurements during the summer 1959–60 show similar results. The moraine cairn observations have therefore been discarded because of their lower reliability as a result of the collapse of cairns, etc., but fixes at Stake 3 (58 m. a.s.l.) show irregularity of rate of movement, though of a lower order than in the previous case nearer the valley wall and nearer the snout of the glacier. The rates of movement of Stake 3 were as follows:

25 November–30 December 1959	0·132 m./day
30 December 1959–26 January 1960	0·105 m./day
26 January–1 March 1960	0·152 m./day
1–31 March 1960	0·118 m./day

No correlations have been found between these rates of movement and the meteorological conditions characterizing each period.

On Styggedalbreen in Norway the difference between the velocity during one period and the succeeding was found to be on an average 10 per cent, reaching a maximum of 30 per cent (Ahlmann, 1928). This variation is low compared with the periodic variations on Depot Glacier, where at Stake 2 it is 57 per cent. This high value is almost certainly the result of the position of the stake near the glacier snout where calving may have had an effect on ice movement for a short distance up-glacier. However, the same explanation cannot account for the value of 47 per cent for Stake 3. On a glacier where movement is slow, such as Depot Glacier, a small change in the rate of movement can result in a large change in the percentage value.

b. *Seasonal variation*

A similar examination of points on the glacier for winter/summer variations produced some anomalies but there was a strong indication that movement is more rapid in summer than in winter.

The medial moraine cairn (Stake 2) showed a movement 50 per cent greater in summer than in winter, a reliable result as the figures refer to periods of 128 and 124 days respectively. Similarly, a stake fixed in the middle of the glacier (Stake 3) had a movement rate of 0·06 m./day between May and October 1958, as opposed to a stake in almost the same position measured in the summer 1959–60 which had a movement rate of 0·125 m./day. Another stake in the same locality gave an annual movement value of 0·08 m./day, which as an overall figure agrees with the mean of the seasonal figures.

Stake 1, the nearest stake to the valley side, showed identical rates of movement over the 245 days from March to November 1959 and the ensuing 96 days from November 1959 to the end of February 1960. Similarly, the lateral moraine Stake 5 showed no appreciable seasonal variation; 0·054 m./day for December 1958 to March 1959 (109 days), 0·061 m./day from March to November 1959 (245 days) and 0·058 m./day from November 1959 to March 1960 (128 days).

From these observations it therefore seems that seasonal variation increases towards the centre of the glacier which is in direct opposition to the findings of observers on some Northern Hemisphere glaciers, where seasonal variations increase towards the glacier margin (Ahlmann, 1948).

c. *Snout movement and calving*

Regular plots on a plane table were made from three triangulation points once a month during the summer of 1960 to determine the rate and time of commencement of calving. The summer retreat of the glacier (30–35 m.) was almost the same as the forward movement over a year, which supports the evidence of the annual plots of the snout positions in showing that terminal recession is very slight. Some calving took place as early as December 1959, but the greatest activity occurred in the months of February and March 1960. Calving generally ceases on this glacier some time in April; in 1959 sea ice formed in June in front of the snout and stayed in until October, yet it contained no calved glacier ice. This is a situation which does not exist off the snout of Arena Glacier, which in both the winters of 1958 and 1959 discharged a large volume of ice into the bay, ice which was incorporated in the much disturbed sea ice off its snout.

d. *Conclusions*

The movement rates are low when compared with the movement of ice in other valley glaciers (Table II). This is partly a reflection of the small size of Depot Glacier which consequently has a small regime. The ice thicknesses (p. 19 and Fig. 8) and movement rates indicate that Kenney Glacier in particular has a very inactive regime and if this arm of the glacier network does have a negative budget then stagnation of the ice there is not a far distant occurrence. The concurrent thinness of the ice piedmont east of Mount Flora and The Pyramid shows that local meteorological conditions are the most important causal factor for the slight regime of Depot Glacier. Of these meteorological conditions the deflative effect of the south-south-west wind is the greatest single factor.

4. BUDGET

The budget of Depot Glacier (Table III) has been calculated on the evidence of the stake measurements, pit profiles and snout plots. The density of glacier ice has been taken as 0·9 g./cm.^{3} and the glacier's cross-sectional area has been computed graphically using the 1955 local survey map and the Admiralty hydrographic chart (Plan 3213), which gives the soundings in front of the snout. The soundings do not

TABLE II

GLACIER MOVEMENT RATES

Glacier	Rate of Movement (m./day)	Reference
Depot Glacier	0·075	
Russell East Glacier	0·53	
Valley glacier on Livingston Island, South Shetland Islands	0·23	Hobbs, 1959
Unnamed glacier,* King George Island, South Shetland Islands	0·40	Hattersley-Smith, 1948
Tasman Glacier, New Zealand	0·15–0·305	Speight, 1940
Hoffellsjokull, Iceland	0·55–2·11	Thorarinsson, 1939

* Lat. 62°06′S., long. 58°22′W.; ice piedmont south-west of Lussich Cove.

give a complete cross-section and the gaps have been filled in by extrapolation based on a knowledge of surface ice features.

The net accumulation due to firn in Cirque 2 and Kenney Glacier and the net accumulation by superimposed ice in the cirques and on Kenney Glacier are all based on comparisons with other areas of the glacier; the accumulation values are small and introduce only a small possible error. The ablation on

TABLE III

BUDGET OF THE DEPOT GLACIER NETWORK (SEE FIG. 4)

Locality	Area ($m.^2 \times 10^3$)	Water Equivalent of Ice or Snow Accumulated or Ablated (m.)	Volume of Water Equivalent (m.)
Net accumulation by firn			
Lee firn area	409	0·48	196,320
Main accumulation area	1,166	0·20	233,200
Cirque 1: avalanche area	47	0·50	23,500
rest of firn area	133	0·18	23,940
Cirque 2	142	0·30	42,600
Kenney Glacier	360	0·20	72,000
Net accumulation by superimposed ice			
Main valley	514	0·04	20,560
Cirque 1	56	0·04	2,240
Cirque 2	276	0·04	11,040
Kenney Glacier	294	0·04	11,760
TOTAL	3,397		+637,160
Net ablation			
Main area of Depot Glacier	742	0·46	341,320
Secondary area of Depot Glacier	352	0·025	8,800
Kenney Glacier	1,507	0·40	602,800
TOTAL	2,601		−952,920
Calving	26	111·00	−2,886,000
BUDGET			−3,201,760
SPECIFIC BUDGET			= −0·534 m./unit area

Kenney Glacier is based on measurements at Stake 18 and discontinuous readings at four other stakes sited down the centre of the glacier. The secondary ablation area of Depot Glacier refers to the higher parts of the ablation zone and includes the very steep ice apron south-west of Cirque 1.

A comparison of photographic evidence from 1903 (Nordenskjöld and Andersson, 1905), 1945 (James, 1949) and 1960, and the evidence of the comparatively static position of the ice front indicate that Depot Glacier is in a state of relative equilibrium. Reasons for the disagreement between historic evidence and the measurements of 1959–60 lie in a possibly large exaggeration of the net ablation, a slight under-estimation of the accumulation and a variation in ablation seasons from year to year.

a. *Retention of melt water*

Kenney Glacier is the most important area in connection with the ablation figures as the net ablation on this tributary glacier is greater than the net ice ablation on the whole of the remaining glacier ($6 \cdot 03 \times 10^3$ to $3 \cdot 50 \times 10^3$ m.^{3}). There were few melt streams on the main body of this glacier during the ablation season. A small amount of melt water left the glacier through the moraine near Scar Hills and a small stream flowed down the steep slope above Depot Glacier and into the south-eastern melt stream of that glacier. This loss to Kenney Glacier can hardly be more than 5 per cent of the total gross ablation and a large percentage of melt water must be retained in crevasses. Crevasses parallel to the contours are mainly concentrated towards the centre of the glacier and are about 15 cm. wide, although below 190 m. a.s.l. some crevasses reach a width of 1 m. Most of the narrower crevasses were filled with melt water by the end of the ablation season but some of the larger ones remained open. A high percentage of melt water has to cross the path of one or other of these open crevasses, which implies that an equally large bulk of melt water did not leave Kenney Glacier in a fluid state, because it is unlikely that these crevasses have connecting channels with an outlet to the sea.

A similar situation exists on Depot Glacier where the two major melt streams flow into crevasses 400 m. from the ice front. Melt water running at the sides of the glacier and discolouration of sea-water beyond the ice front were never observed. Crevasses must therefore retain a high percentage of the gross ablation but as the volume of the crevasses on Kenney Glacier is insufficient to retain the total melt water formed in the ablation season, evaporation must play an important part in the ablation process. In fact, the föhn effect on air descending from 600 m. a.s.l. on the west produces clear skies and a low relative humidity, both of which strongly favour evaporation.

b. *Calving*

In the calculation of the volume of ice calved from the ice front it has been assumed that the front calves equally from the glacier surface to the glacier bed below sea-level. In fact, this is unlikely to be the case because of a decrease in the rate of movement with increasing depth. A calculation of the amount of ice moving through a line of cross-section at the regional firn line (303 m. a.s.l.), assuming that velocity is constant with depth, gives a figure far in excess of the glacier's annual accumulation above the firn line. To calculate a volume in agreement with the accumulation figure and a state of equilibrium, the necessity to assume a considerable decrease in velocity with depth arises. A similar situation applies to calculations of the movement of ice from the two Depot Glacier cirques and it is therefore safe to assume that velocity decreases with increasing depth (which is in agreement with the mathematical consideration of the problem by Nye (1952)). This means that the seaward edge of the glacier snout slopes backwards below sea-level giving a substantial overhang. The figure calculated for the amount of ice lost by calving may therefore be far in excess of the actual volume lost. It is this factor more than any other which has produced the apparent anomaly of a very negative budget on a glacier which does not show signs of the recession the figures demand.

c. *Variation of summer weather conditions*

It is important to bear in mind that Table III refers to one budget year only and it is a characteristic feature of the area that one year may be strikingly different from the next. The summers of 1958–59 and 1959–60 were quite dissimilar and in the former there could have been very little surface ice ablation (the snow line was at sea-level for most of the summer) but an appreciable amount of firn and superimposed ice accumulation.*

V. ICE PIEDMONT BETWEEN HOPE BAY AND TREPASSEY BAY

THIS ice piedmont (Fig. 1) lies immediately to the south of the base, where it terminates on land in a steep slope. With a gently sloping surface it extends from 400 m. a.s.l. backed by The Pyramid and Mounts Carrel and Flora and ends for 90 per cent of its length in 30 m. high ice cliffs.

The purpose of the study of this ice piedmont was to determine to what extent it is a relic feature, by discovering the state of its present budget.

Six stakes were set up and the movement, accumulation and ablation were measured at each one for

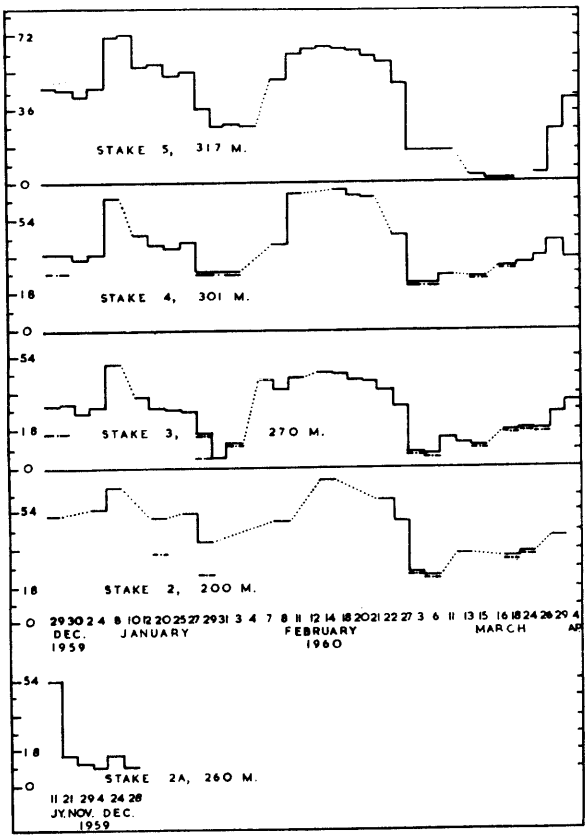

FIGURE 10

Accumulation and ablation measurements on the ice piedmont between Hope Bay and Trepassey Bay, July 1959–April 1960. Ordinate scale in cm.

All the changing snow and ice levels are relative to the initial measurement in the first bar of each graph. The dash-dot-dash line is the level of superimposed or glacier ice, whether at the surface or under a snow cover.

the period 28 December 1959 to 4 April 1960 (Fig. 10). Two of these stakes were visited on only a few occasions and one provided a reference to the ice level at the end of the previous summer, whereas the other four were visited about once every three days.

1. REGIME

Ablation and accumulation

The average height of the firn line at 300 m. a.s.l. excludes most of the piedmont from a firn cover,

* The altitude of the snow line in the summer of 1958–59 was low over a very wide area, e.g. at Wilkes station (lat. 66°S., long. 110°E.) the snow line was 350 m. lower than in 1957–58 (Hollin and Cameron, 1961).

which is restricted to the two col areas in the lee of the mountain slopes. It was soon evident that the ice surface would suffer a regular lowering if this was the sole source of nourishment and for this reason the *ice level* on the stakes was measured both when it was snow-covered and snow-free.

Absence from the Hope Bay station on a sledge journey prevented measurements of the superimposition of ice beneath the winter snow in early autumn, but a small amount must form before the ice reaches melting point. This amount will be of a low order because the snow cover on the ice piedmont is seldom more than 0·3 m. in depth and there must be only a short time lag between the basal snow cover and surface ice layers reaching melting point. This compares with the average of 96·5 cm. of snow on the Barnes Ice Cap in Baffin Island at the beginning of the ablation season. Baird (1952) found that superimposition of ice took place below the snow cover and amounted to 12·7 cm. over the ablation period. Baird cites four definitive characteristics of what are termed "Baffin type glaciers":

i. Insufficient altitude to reach the local firn line.
ii. Great residual cold in the ice.
iii. Light precipitation.
iv. Usual Arctic environment of short cool summers and long cold winters.

The ice piedmont considered in this section fulfils these conditions, but only to a certain degree. The summers are open to considerable variation (cf. Grinnell and Terra Nivea Ice Caps in Baffin Island (Mercer, 1956)) and the effect of the local south-south-westerly wind can cause great variations in the snow cover at the end of the winter. In addition, the second factor is not absolutely satisfied as the surface ice in some years must quickly reach the melting point. Melt periods can occur at any time of the year and this means that, if one of these warm spells coincides with a thin snow cover and if the period is short enough, then superimposed ice will form, whereas if the snow cover is thicker the melt produced will be spent entirely on firnification and the formation of ice layers within the snow cover.*

i. *Melt season sequence.* The snow cover in February–March 1959 was approximately 50·0 cm. deep but a south-south-west blizzard in April 1959 reduced the ice piedmont to bare ice below 250 m. a.s.l. and considerably reduced the snow cover at altitudes above this. By July 1959 the snow depth reached 53·0 cm. and much of this was again ablated by blizzards between then and late October. By 21 November 1959 the snow cover at 270 m. a.s.l. was only 20·0 cm. deep and remained at about this depth throughout December.

a. *November 1959 to 11 January 1960.* There was moderate snowfall on 3, 6, 7 and 8 January and slight snowfall on 4, 5 and 9 January, which covered the ice piedmont with 16·0 cm. of snow at 200 m. a.s.l. and 31·0 cm. at 300 m. a.s.l. Accumulation increased with altitude as precipitation was associated with only slight drifting. This snow cover was not completely removed again until 27–29 January.

b. *11 January to 1 February 1960.* The first important thaw period began on 11 January when there were 22 thaw degrees as opposed to 8 freeze degrees (Fig. 7). The surface soon became composed of new wet snow overlying slush and the flow of melt streams from the edge of the ice recommenced. There were 48 thaw degrees with no freeze degrees on 12 January, and slusher streams were abundant down the ice slope behind the station. This unusually warm period continued for nine days when the daily thaw degrees never dropped below 17 and freeze degrees never rose above 2. At the end of the period the surface below 200 m. a.s.l. consisted of ankle deep slush composed of a top layer of ice crystals each measuring more than 1·0 cm. in length overlying a water-saturated crystal layer resting on glacier (or superimposed) ice. Above 200 m. a.s.l. the surface consisted of a similar depth of crystals with very little free water present between the crystals. At 300 m. a.s.l. the surface was composed of wet firn.

The greatest surface lowering occurred between the measurements of 8 and 12 January but this was largely the result of the compaction of loose snow. Ablation was interrupted by low temperatures which affected the area for four days and froze the surface layers. A re-invasion of warm air restarted ablation and there was an important lowering of the surface between 27 and 29 January, largely arising from the effect of moderate rainfall on 28 January. High wind speeds (average approximately 12 m./sec.) with a low relative humidity (less than 70 per cent for most of the day) on 29 January and rapid melt-water run-off from 26 January onwards helped to dry out the surface above 150 m., where by 30 January the surface was almost free of the slush that had been ankle deep a week previously. To determine the thickness

* At nearby Eagle Island the ice surface temperature on the ice cap at 300 m. a.s.l. in late October 1959 was −8·8° C under a 45 cm. cover of dense firn, of which the upper 10·0 cm. had already reached the melting point. Superimposition of ice under the winter snow cover has already been mentioned on p. 15 in connection with Depot Glacier, where it was found to be quite important.

of the firn on the ice the surface near Stake 2 was probed and a pit was dug nearby. This showed that the ice level at Stake 2 had been lowered by 15·0 cm. between 20 and 29 January despite the fact that there had been an 18·0–20·0 cm. cover of coarse loose firn throughout the period. At Stake 3 (270 m. a.s.l.), on 29 January, there was an 11·0 cm. cover of large ice crystals while at the same time bare ice was exposed at Stake 4 (300 m. a.s.l.). The layer in contact with the ice surface was saturated and the top layer of ice was probably formed in the March 1959 ablation period at the end of the cold summer of 1958–59 and during the winter of 1959 when slow percolation of melt water and subsequent re-freezing produced air-rich superimposed ice (Schytt, 1955, p. 52–57). Once the surface ice reached melting point interstitial ice thawed out leaving loose ice crystals lying on the unaffected ice below.

High temperatures and ablation continued until 2 February and by the time of the last stake measurements in this period (31 January) the ice piedmont below 270 m. a.s.l. had reached almost its lowest level of the summer. Relatively low temperatures on 2 and 3 February froze the ice and snow surface and a 6·0 cm. layer of slush with a thin ice crust collected at Stake 3.

c. *2–20 February 1960.* The second and most important snowfall of the summer occupied this period and it gave the ice piedmont a continuous snow cover which lasted until the end of February 1960. 10·0 cm. of snow fell in a moderate breeze on 6 February and by 7 February the snow depth had increased to 23·0 cm. at Stake 3 (270 m. a.s.l.). Settling resulted in a 5·0 cm. drop in snow level by 8 February but further moderate snowfall on 8 and 9 February increased the snow cover by another 9·0–22·0 cm. Slight snowfall continued to increase the snow level so that by 14 February there were 30·0 cm. of snow resting on ice at 225 m., 35·0 cm. at 270 m. and 41·0 cm. at 300 m. a.s.l. This snow formed an important protective cover against the most pronounced ablation period of the 1959–60 summer which began when the February snow cover was almost at its thickest.

d. *21 February to 5 March 1960.* Six measurements of the glacier surface level were made during the main ablation period of 20 February–6 March. An average lowering of surface level in cm./day is shown in Table IV.

TABLE IV

LOWERING OF SNOW OR ICE SURFACE IN CM./DAY
(20 February–3 March 1960)

Date	*Average of 4 Stake Readings* (cm./day)	*Average Daily Thaw Degrees*
20–22 February 1960	2·44	16·8
22–27 February 1960	2·90	39·0
27 February–3 March 1960	5·40	49·0

The accumulative effect of the high temperature was mainly responsible for the increasing rate of ablation as the period continued. The increasing rate represented by the values in Table IV is an underestimate of the true gross ablation, because surface lowering between 27 February and 3 March was of denser snow (and ice towards the end of the whole period) than that constituting the snow cover at the beginning of the ablation period on 20 February.

By 6 March the ice piedmont below the firn line had reached its lowest level of the 1959–60 summer. This was the result of run-off and slusher flow which had reduced the surface to bare ice below the firn line at 305 m. a.s.l. However, the ice was not in every case at its lowest level. At Stake 2 (200 m. a.s.l.) the ice level was 4·0 cm. above the level of ice at the end of the first ablation period (on 29 January). An additional layer of ice had been formed from refrozen slush which had accumulated after 29 January and frozen between 31 January and 3 February. At 270 m. a.s.l. 6·0 cm. of frozen slush had been added to the surface and 4·4 cm. of this layer were ablated during the February–March ablation period.

Between 3 and 6 March very little ablation took place and the latter date marked the end of the main ablation period, during which the surface had been lowered at a rate of 3·5 cm./day (cf. 1·9 cm./day in the first ablation period).

e. *5–26 March 1960.* In this period ablation continued below the firn line but accumulation by freezing of melt water and rain on the ice surface exceeded ablation. An alternation of warm and cold spells of weather (Fig. 7) in early and mid-March favoured the formation of a smooth ice surface from re-frozen melt water formed in that area in the preceding warm spell, or water running from the firn above 305 m. a.s.l. The latter quantity is very small as any increase of the ice level in this period was usually of a similar magnitude at each stake, one of which (Stake 2) was 1·5 km. from the firn line. A certain quantity of melt water was formed whenever warm air covered the area after a period of snowfall at relatively low temperatures. More important than snow melt was the supply of water from rainfall. Rain and sleet in the second half of March was frozen on to the ice surface in intervening cold periods of weather and formed the most important source of ice accumulation.

Above the firn line the firn surface continued to drop for another ten days and accumulation did not exceed ablation until the onset of winter conditions at the end of March.

Table V gives the results of ice measurements, although the datum level of ice includes a small amount of ice formed at the base of the winter snow cover in the spring. The measurements place the equilibrium line for 1959–60 between 240 and 250 m. a.s.l. As the ice budget values above and below this altitude are both less than 4·0 cm., the height of the equilibrium line is only an approximation.

TABLE V

ICE BUDGETS ON THE ICE PIEDMONT, DECEMBER 1959–APRIL 1960

Stake Number	*Height* (m. a.s.l.)	*Loss* (cm. of ice)	*Gain by Superimposed Ice* (cm. of ice)	*Net Accumulation or Ablation* (cm. of ice)
1	73	19·0	0·0	−19·0
2	200	15·0	13·5	− 1·5
3	270	11·4	12·0	+ 0·6
4	301	6·3	7·6	+ 1·3

The ice datum level at Stake 1 is February 1959.

The ice datum level at the other stakes is 29 December 1959 and the negative side of the budget is accordingly exaggerated.

ii. *Melt and slusher streams.* Melt streams form an important factor in ablation on the north-west-facing slope of the ice piedmont. They flow intermittently between late November and late March and are dependent on the temperature and snow cover. There are three major outlets: one via Lake Boeckella, one parallel to and along the edge of the ice margin and one 100 m. from the ice margin.

Whenever a period of warm föhn air occurs at the same time as there is a snow cover of 15·0 cm. or more, slusher streams break out down the steep ice slope behind the Hope Bay station. They are liable to occur at any time of the year, and in July 1958 they occurred after a very warm spell when the temperatures exceeded 10° C for a few hours and stayed above freezing point for a few days.

The preliminary conditions necessary for the initiation of these streams is the complete saturation of the snow cover over a large area of the ice piedmont. In March 1958 the more level top of the ice piedmont became covered with a 30·0 cm. layer of slush and small slushers could be started on the steep slope by artificially creating a small channel to allow some free movement of the saturated mass. The slusher stream collects all the slush in its path, which addition increases the size of the head, and a bare ice path is left in its wake. Its start is partially attributable to the formation of excess water, which can no longer be held by the snow and which flows away creating a small initial channel, which induces further movement until the whole mass starts on the move. Its demise can be brought about half way down the slope if the snow lying in its path is able to absorb the excess water.

These streams have been observed to reach a width up to 10 m. and one had sufficient momentum to break through a bolted wooden door of a hut 20 m. from the ice margin. On the day of their occurrence they give a pulsating rhythm of flow to the major outlet streams, and slush fans left on the rock quickly ablate. The back step of a slusher stream bed is well marked (Plate Ia) as a large proportion of water drains out of the undisturbed snow and leaves a small vertical step standing above bare ice. The bed of a slusher stream receives a significant cover of superimposed ice at the end of the melt season and this together with the drift infilling means that the following season's slushers will not necessarily follow the same course. Slusher streams have also been described from North East Land (Glen, 1939), Baffin Island (Ward and Orvig, 1953) and north-east Greenland (Lister, 1958).

2. Movement

At Stake 1 movement was measured over a period of 21 months but at Stakes 3, 4 and 5 observations covered a period of 3 months in the summer of 1960.

The movement of unchannelled ice is of a very low order and at Stake 2 it was so small as to be unmeasurable with the standard of accuracy employed. At Stake 1, near the ice front, it was 0·004 m./day, which might be given as an upper limit for Stake 2. This, plus the infrequent calving at the ice front, partly accounts for the static nature of the ice piedmont, which, if one can judge from photographs given by Nordenskjöld and Andersson (1905) has altered little in its horizontal extent since 1903.

The slightest amount of channelling increases the rate of flow substantially. The two stakes above Nobby Nunatak (only Stake 4 was on a steep gradient) turned away from the regional contours and into a narrower line of flow between Nobby Nunatak and Mount Flora. This same channel shows similar surface flow patterns to those of a valley glacier where flow is at a maximum towards the centre (this was deduced from visual observations on marker poles placed across this channel).

The main area of channelling on the ice piedmont is into Trepassey Bay, where ice from this particular part of the piedmont joins that from broad ice channels draining other parts of Tabarin Peninsula. The high rate of flow is suggested by the relatively high degree of crevassing without any increase in slope.

It seems likely that these channels are draining their sources faster than accumulation can compensate for the deficit. The source between Mount Flora and The Pyramid gives the appearance of a negative regime, since steep snow slopes sweep down from the firn region above 330 m. a.s.l. into the broad depression where movement was measured.

3. Conclusions

An estimate of the regime of the ice piedmont for the year 1959–60 gives a slightly negative budget of $-5{\cdot}0$ cm. Calving of ice from the ice front has been discounted on the grounds that no calving of the ice front was observed from the position of the Hope Bay station in two and a half years. Calving occurs occasionally and preserves the verticality of the ice cliffs but, since the only evidence pertinent to ice movement and calving is for the area represented by Stake 1, any estimation would possibly introduce large errors. Calving and its part in increasing the negative side of the budget must, however, be borne in mind, and will be considered later.

Variation of ablation conditions from one summer to another cannot be too strongly emphasized and during an average summer it is not unlikely that the equilibrium line lies closer to the altitude of maximum area at 150 m. a.s.l. (Fig. 11). A graphical interpretation of the regime below the firn line (Fig. 11) illustrates a rapid increase in ablation below 150 m. a.s.l. In the part of the ice piedmont represented by stake measurements the high ablation below 150 m. a.s.l. is largely a result of the increased effect of radiation on a north-facing slope where dust and moraine debris contribute to lowering the albedo of the surface. If a stake profile had been set in the ice piedmont above Trepassey Bay, the ablation curve would almost certainly have been less steep below 150 m. a.s.l. Accumulation by drift is important below 130 m. a.s.l. south-west of Stake 1 and the summer of 1959–60 was the only time that the surface around Stake 1 was reduced to bare ice. During periods of ablation when bare ice was exposed in wind-scoured areas above 230 m. a.s.l. and at lower altitudes between Nobby Nunatak and the edge of the ice piedmont near Grunden Rocks, firn or snow covered the ice east and south-east of Stake 1. The line between bare ice and firn was clearly marked and was situated about 100 m. west of Stake 1. It coincided with the line beyond which crevasses were an important feature of the surface and the firn cover may be accounted for by the fact that melt water drained into the crevasses so that formation of deep slush and ablation by slusher streams was prevented. The stake profile therefore represents that part of the ice piedmont which undergoes maximum ablation and, since the measurements were taken during a very warm summer,

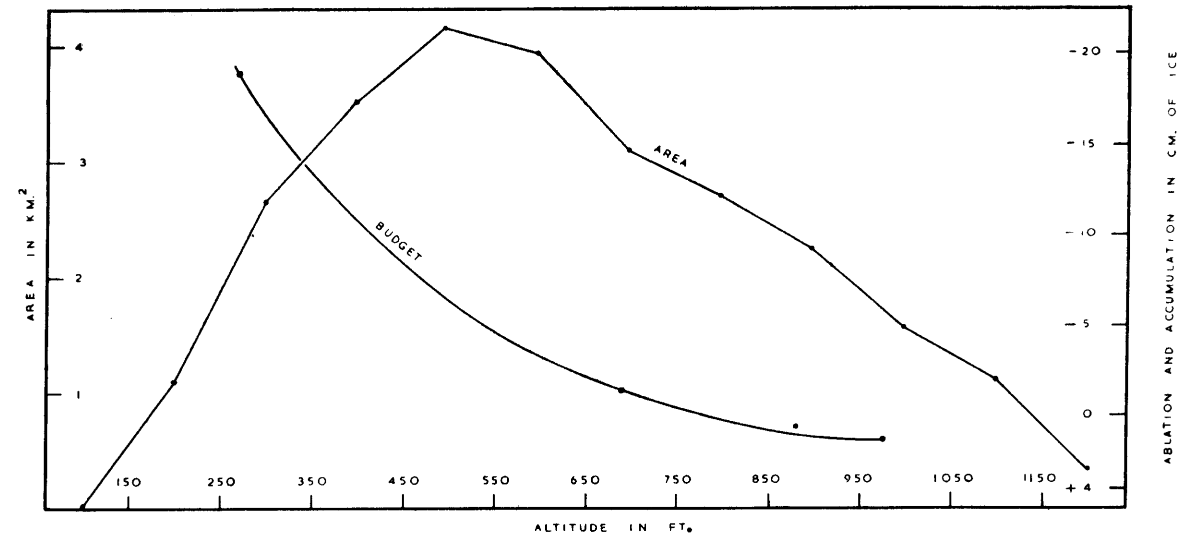

FIGURE 11
Diagram showing the area distribution curve and annual ablation per unit area in relation to altitude above sea-level on the ice piedmont between Hope Bay and Trepassey Bay.

the ablation situation is depicted in its most extreme form. Despite this the ablation figures are not of a large scale: a gross ablation at 70 m. a.s.l. of 17·1 cm. of water. The net ablation is something less than this as a certain amount of melt water is taken up by crevasses. The ice piedmont as a whole has a regime very close to equilibrium and if calving is excluded from the calculations the budget over a period of years must be slightly positive. A state of equilibrium results from a loss of ice by calving. It is interesting to note that the regime of Depot Glacier in 1959–60, excluding the loss of ice by calving, was −5·0 cm. of water which has been shown to include exaggerated values of net ablation. It seems likely, therefore, that glaciers in the vicinity of Hope Bay have a positive regime when judged on the basis of all factors of accumulation and ablation except calving. This positive side of the budget is compensated for by calving at the ice front and this is a means by which glaciers are maintained in a state of equilibrium. Photographs taken by Bodman in 1903 and a sketch map of Hope Bay by Duse (Nordenskjöld and Andersson, 1905) show no significant differences in the horizontal and vertical extent of the ice piedmont north-east of Mount Flora. This supports the hypothesis suggested by the budget measurements and calculations for 1959–60.

Calculations of ice thickness (p. 19) show that the ice piedmont is very thin and the greater part of it between Hope Bay and Trepassey Bay must be between 50 and 100 m. thick.

The low degree of activity of the ice piedmont's regime and its state of equilibrium indicate, as was originally expected, that the ice piedmont between Hope Bay and Trepassey Bay is a relict feature. As such it is a survival from before the most recent general glacier recession (a characteristic of most of the "Baffin type" glaciers (Baird, 1952)).

VI. TRINITY PENINSULA AND THE ISLANDS IN PRINCE GUSTAV CHANNEL

WORK on the glaciers of Trinity Peninsula and the islands in Prince Gustav Channel (Fig. 1) during sledge journeys was by necessity of a preliminary nature. It was concerned first with observations on the movement and accumulation of mainland and island glaciers and secondly with recording evidence relevant to the glacial history of the area.

The main discovery of the first investigation was a decrease in glacier activity eastward. The regionally dominant snow-bearing wind has a westerly component and, whereas it increases the accumulation on the west coast of Trinity Peninsula, it promotes ablation on the east coast which it reaches as a föhn wind with humidities below 60 per cent and very occasionally below 10 per cent.

1. Regime, Accumulation and Ablation

Pit digging and visual observations placed the firn line somewhere below 150 m. a.s.l. on the west coast of Trinity Peninsula, between 230 and 270 m. a.s.l. on the east coast south of Duse Bay, at 300 m. a.s.l. on Tabarin Peninsula and above 300 m. a.s.l. on the east coast islands.

In winter the föhn wind favours snow retention to a certain extent by cementing the winter snow at the lower levels, the supply being drift from the plateau and col surfaces. In the Hope Bay area the dominant strong wind is the cold south-westerly which causes considerable deflation and can erode and remove the winter snow cover far more rapidly than can the föhn wind by melting. It is believed that this is the main reason for the greater height of the firn line in this area.

a. *Rime and fog deposit*

Accumulation by rime and fog deposit is important on the ice piedmont of the west coast of Trinity Peninsula and on Joinville Island. Rime and fog deposit is associated with low stratocumulus and stratus cloud which envelopes these areas for the greater part of a year. This cloud bank usually ends along the eastern margin of the plateau and sometimes descends on to the higher areas of the east coast ice piedmont. In the plateau gaps at the heads of Russell East and Russell West Glaciers, Windy Gap and Misty Pass the cloud descends to a lower level, reaching approximately 550 m. a.s.l. The cloud lies in these cols for a greater part of the year than on the Detroit and Louis-Philippe Plateaux and accordingly rime and fog deposit accumulation must be greater there.

Following a suggestion from Professor D. L. Linton, a sledge journey was made to Mount Bransfield (756 m.) to examine the ice formations of the north-west or windward parts of the mountain. These were found to consist of huge lobes of ice, some measuring about 15 m. by 30 m., the thickest part protruding 5–10 m. from the contact with the mountain. Some of these lobes (Plates Ib and IIa) were detaching themselves from the mountain under their own weight and crevasses 1 m. wide were found within 7 m. of the summit of Mount Bransfield. Rime ice and fog deposit clearly produce an immensely active regime on this mountain which lies directly in the path of the moist north-west winds and which is covered by orographic stratocumulus cloud for much of the year.

Sledge journeys to areas 60 km. farther south showed that the entire ice piedmont on the west coast of Trinity Peninsula is affected by rime and fog deposit during periods of low cloud. Every mountain in this area presents a similar appearance to Mount Bransfield. Mount Canicula, at the head of Russell East Glacier, was examined in some detail and some of the rime ice lobes were found to have broken off from the west-facing side of the mountain and were lying at the foot of the slope. One large block measured about 20 m. by 4 m. by 4 m. and was in the process of becoming incorporated in the *névé* of Russell East Glacier. Very small rock exposures were visible on the north-west side of Mount Canicula and it is probable that they represent areas from which rime ice lobes have recently broken away from the rock face. At the summits of mountains or along ridge tops the north-west sides support overhanging lobes of ice and on Marescot Ridge there are occasional small towers of ice on the mountain tops, but none of them approach the proportions of the larger rime "mushrooms" in the Monte Fitz Roy region of Patagonia (Lliboutry, 1953).

In certain places on the north-west side of Mount Canicula the rime ice surface is fluted, each rib measuring more than 1 m. and up to a maximum of 10 m. in length. The most typical surface pattern is similar to a crocodile skin in appearance and the individual protuberances decrease in size with increasing horizontality of the slope.

Cores from steep west-facing slopes have totally different profiles to the pits dug on the level glacier nearby. A pit dug on the top of a rime ice knoll near Sirius Knoll, and a core from a steep west-facing slope on the northern shoulder of Mount Bransfield both showed ice and dense firn layers within 10·0 cm. of the surface. A pit on the top of Wimple Dome (518 m. a.s.l.), 150 m. below the other sites, showed thicker firn and snow layers overlying ice. Pits dug to a depth of 1·8–2·2 m. on the level surface of Russell West Glacier and in the Marescot Ridge area uncovered very few ice layers and these were thin. These pits have been interpreted as showing that rime ice and fog deposit accumulation is the dominant process on steep west-facing slopes but is insignificant on more horizontal surfaces. Rime ice layers in Pit 3a (700 m. a.s.l.), including the rime ice surface, were very distinct. Each layer consisted of parallel 2–3 mm. wide pillars of rime ice inclined west-north-west at an angle of about 45°. These layers formed an insignificant part of the accumulation but, as a very thin layer of rime often forms on the windward edge of small irregularities in the surface, its major contribution to accumulation lies in its ability to fix the accumulation and thereby resist drifting. Glen (1941) came to a similar conclusion concerning rime ice accumulation on West Ice in North East Land. Ahlmann (1929) found that rime ice contributed a greater part to the accumulation than snow in the mountains of the Jotunheimen of Norway. At the heads of Russell East and Russell West Glaciers rime ice and fog deposit on stakes showed that, while accumulation on the glacier surface was less than 1 mm., accumulation on a vertical stake surface was often more than 40 mm. The rate of accumulation on vertical stakes sometimes amounted to 1 cm./hr., the rate decreasing as the accumulation increased. The accumulation of rime and fog deposit on stakes increased proportionately with height due to greater air circulation away from the ground. Rime and fog deposits form on the windward edge of small irregularities but the complete absence of large skavler makes the accumulation by this process insignificant.

An attempt was made to examine the rate of accumulation by rime ice and fog deposit by making periodic measurements of the dimensions of snow cairns built to serve as survey triangulation points. Table VI shows the response of these snow cairns to the climatic conditions but the variation in their reaction poses the problem as to whether only solid ice cairns would give representative figures.

TABLE VI

RESPONSE OF SNOW CAIRNS TO RIME ICE CONDITIONS

Cairn	*Date Built*	*Measurements*	*Date Revisited*	*Measurements*	*Result*
1	September 1959	—	3 weeks' interval	—	Considerable deflation
2 (Sirius Knoll)	September 1959	10 ft. 9 in. high ~3 ft. 0 in. high 18 ft. 6 in. high	3 weeks' interval	Same	Slight deflation on leeward side (S.E.) and very slight rime on N.W.
3	February 1959	8 ft. high 4 ft. square	November 1959	6 ft. high 2 ft. 6 in. square	Deflation
4 (Mount Bransfield)	January 1959	8 ft. high 6 ft. square	March 1960	8 ft. high 8 ft. square	Accumulation

b. *Pit profiles*

On Trinity Peninsula the interpretation of firn layering is difficult because of the lack of meteorological data for this area and the occurrence of random ice layers. Determination of the snowfall for the winter of 1959 up to the time the pits were dug is straightforward but the interpretation below that must be done with caution (see Figs. 12 and 13).

In Pit 1a firnified layers of snow at 223 cm. have been taken to represent the summer period of 1958–59. These denser firn layers separated two layers of fairly loosely packed firn. The average grain size in the lower layer was 1·0 mm. although several grains just exceeded 2·0 mm. The summer of 1958–59 was atypically cold with the average temperature in February 3° C below freezing point. For this reason there is only one very thin ice band present, at the beginning of the ablation period, and the late warm period in March is represented by 2·5 cm. of dense firn and ice firn, and 5·0 cm. of very dense firn. The exceptionally warm July weather in 1958 is represented by a 3·2 cm. band of ice firn containing two ice lenses at a depth of 2·74 m. Below this the first ice band appears (summer 1957–58) and this is separated from a second by 36·0 cm. of dense firn.

The relation of this pit to the saturation line is difficult to determine since the pit only penetrates three summer periods. From this slender evidence, however, it seems likely that the saturation line (which moves in accordance with summer conditions) was below this level (700 m. a.s.l.) in the summer of 1958–59 but was above it in the summer of 1957–58. This is suggested by the horizon of 14·0 cm. of low-density firn towards the base of the 1958 budget year's accumulation. No similar variation was noted at the base of the 1957 budget year's accumulation and it can be assumed that melt water percolated throughout this layer.

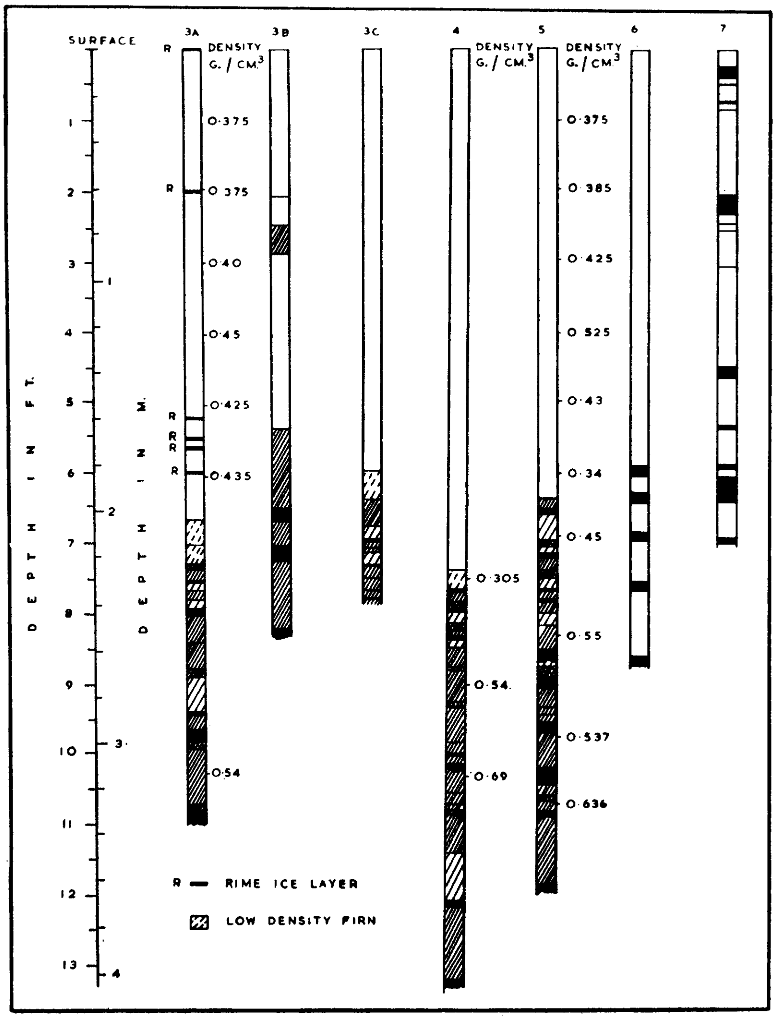

FIGURE 12
Profiles for Pits 3–7 on Trinity Peninsula. In pit profiles 6 and 7 ice banding only is described.

Benson (1959) found the saturation line 460 m. above the firn line in north-west Greenland at lat. 77°N. On Russell East Glacier the firn line is at approximately 230 m. a.s.l. and the iceshed where Pits 3a, b and c were dug lies 457 m. above this. The fluctuation of the saturation line above and below this level suggests that its average height lies between 760 m. and 610 m. a.s.l. (Pit 4) which agrees remarkably closely with Benson's figures. The latent heat released by the formation of rime ice increases the average temperature of the firn and must be responsible for raising the height of the climatic saturation line. Almost the whole of the east coast ice piedmont lies below the saturation line but on the west coast of Trinity Peninsula the difference in altitude between the firn and saturation lines must be greater because of the higher average air and glacier temperatures and the greater accumulation at lower altitudes.

In Pit 3a the density of winter snow gradually increases to a depth of almost 1·22 m. where it shows a slight decrease. The highest firn density recorded in the pit was 0·54 g./cm.^{3}, although some of the thinner ice firn and ice layers exceed this value. These low values illustrate the small amount of melt at this altitude which is accounted for by the fact that much of the pit is composed of winter snow and percolation facies, rather than saturation facies.

Pits 3a, b and c (Fig. 13) were all dug in the broad, level area at the head of Russell East and Russell West Glaciers and yet there is surprisingly little correlation between them. Pit 3b was dug near the foot of the Louis-Philippe Plateau escarpment; the thicker ice bands suggest greater summer melt despite the shadow effect produced by the south-facing escarpment in summer. Pits 3a and c were dug no more than 1 km. apart yet again there are few correlations. This is unusual for an area where the slope is uniform important, the freezing of "freezing rain". A pit dug after the occurrence of the latter phenomenon uncovered a 5·0 cm. ice layer just below the surface; several other ice layers present in the same pit (533 m. a.s.l.) must also have been formed close to or even at the surface, because the temperature in the pit never exceeded $-4\cdot7^\circ$ C which was at a depth of 30·5 cm. (Fig. 13).

The warm, moist north-west wind of the west coast of Trinity Peninsula descends the east coast ice piedmont as a dry wind and its temperature increases at the adiabatic lapse rate for dry air, i.e. $\sim 3^\circ$ C/300 m. This föhn wind is glaciologically important as it brings thawing temperatures in winter to altitudes below 400 m. (below which thin ice layers and ice lenses are recognizable throughout the winter snow cover) and in summer it becomes the most important factor in ablation. During stable conditions on the east coast ice piedmont moist north-west air forms low cloud (base level at 150 m. a.s.l.) over the west coast but east of this the cloud is limited to the plateau and plateau cols. In September 1959 a difference of $11\cdot1^\circ$ C was recorded on Russell East Glacier between the east coast air mass which lay below 460 m. a.s.l. and the moist west coast air mass which descended Russell East Glacier as a moderate westerly gale ($-8\cdot3^\circ$ C) and caused local snowfall. The development of a depression over the western Weddell Sea breaks down these conditions and the west coast air then descends to Prince Gustav Channel as a föhn wind. In winter this wind transports a large percentage of snow in the form of drift from higher to lower altitudes. A similar situation has been found to exist on the Greenland Ice Sheet (Diamond, 1960): snow drifts to lower levels where higher temperatures reduce the snow's ability to drift. The maximum accumulation, therefore, is not coincident with the greatest altitude but lies to leeward of it. This is the situation on Russell East Glacier where winter accumulation was found to be greatest at Pit 5 at an altitude of 430 m. a.s.l. (Table VII; Fig. 13). Winter snow densities are higher at Pit 5 than at Pit 3a which is 170 m. higher up-glacier, because of drifting snow at higher temperatures across the lower pit area. Although maximum winter accumulation is found at the lowest of the three Russell East Glacier pit site levels, maximum net accumulation in the budget year 1958–59 was found at the intermediate level, i.e. at 600 m. a.s.l. Minimum net ablation, as measured by subtracting the accumulation of 1958–59 from the unaffected winter snow accumulation of 1959 which gives a reasonable representative figure for comparative purposes, is at the highest pit level.

The interpretation of the layering in Pit 4 (600 m. a.s.l.) is difficult. Random ice layers and fairly dense firn indicate saturation facies and the percolation of melt water down to a previous year's accumulation. This means that the values given in Table VII must be treated with some caution. The increase in numbers and thicknesses of ice layers between this pit and Pit 5 (430 m. a.s.l.) is very marked and is an indication that the edge of the föhn bank usually lies between these two altitudes. From Prince Gustav Channel the cloud is often seen to extend just down-glacier from Pit 4, which area is therefore frequently in cloud when stable air conditions exist on the east coast ice piedmont of Trinity Peninsula.

Density increases regularly with increasing depth in the thicker firn layers with low density, early winter snow lying on top of the autumn firn. The low density of early winter or late autumn snow may be caused by the presence of a steep temperature gradient resulting from the sharp onset of winter. Water vapour then rises from the warmer early winter or late autumn snow to colder layers above where it recondenses and freezes. In Pits 3 and 4 the late autumn firn horizon is represented by open crystalline firn of low density and loose structure. Similar horizons have been described from the Lake Hazen area of northern Ellesmere Island (Hattersley-Smith, 1960). In Pits 3 and 4 these were more coarsely granular, with a few grains up to 5·0 mm., than the denser winter layers where grains averaged 1·0–2·0 mm. At 700 m. a.s.l. these horizons occurred throughout the pit profile, whereas this is not the case in Pits 4 and 5 below the saturation line. Here the low density firn horizons, in which the crystal size was on an average 2·0 mm., occur only at the base of the current winter's snow.

The Pit 4 horizons have been interpreted with less certainty than those of Pit 5. The low temperatures and high snowfall of February 1959 divided the 1958–59 ablation season into two periods and at Pit 5 two distinct ice layers have been taken to represent these. At Pit 4, 170 m. above Pit 5, the layering is more random and as at Pit 3a layers of ice firn alternate with layers of loose crystalline firn. The accumulation for certain budget years has been calculated by assuming that one budget year extends from the upper surface of one distinct ice layer to the upper surface of the next. The budget year including the cold summer of 1958–59 has been assumed to extend from the upper surface of the ice layer representing the summer of 1957–58 to the base of the unfirnified winter snow where the average crystal size is 1·0 mm.

The climatic firn line lies approximately 150 m. below Pit 5 and for detail on the accumulation between the firn line and 430 m. a.s.l. reference must be made to pits dug on Victory Glacier 11 km. to the south.

The 1958 winter snow accumulation at Pit 10 (300 m. a.s.l.) on Victory Glacier was 150 cm. by early November, which indicates an important decrease in accumulation with decreasing altitude below 430 m. a.s.l. Annual layering was easily recognized and budget year accumulations conform reasonably well with the trend on Russell East Glacier (Table VII). Unfortunately the firn densities calculated at the pit site have had to be discarded because of errors arising from the method employed, and an average firn density of 0·55 g./cm.^{3} has been used instead.

The irregularity of the firn line is evident from the profile in Pit 9 (350 m. a.s.l.) which was dug in the middle of Victory Glacier a distance of 1 km. up-glacier from Pit 10. In Pit 9 117 cm. of firn rested directly on blue ice containing very few air bubbles. This in turn rested on 4·0 cm. of white bubbly ice. This ice band sequence was repeated on top of 16·0 cm. of dense ice. Next there followed a layer of 8·4 cm. of hard dense firn in which the crystal size increased from 1·0 mm. at the top of the horizon to 2·0 mm. at the bottom. A further excavation of 30·0 cm. uncovered indistinctly banded dark ice. Schytt (1955, pp. 52–57) has interpreted both clear, blue, bubble-free ice and white, bubbly ice as superimposed ice, the former forming in the autumn and the latter in the spring. It is clear from the sequence of horizons that the area is one which has often in the near past been between the firn and equilibrium lines and has recently remained mainly above the firn line. The firn lying on the superimposed ice is equivalent to about 64·0–70·0 cm. of water which is three times as much as the annual accumulation in the budget year 1957–58 in Pit 10. The firn profile suggests that in recent years the firn line has remained below this pit area in many budget years. This places a recent short-period climatic firn line below the pit height. Below the banded superimposed ice is dense glacier ice indicative of the ablation area. The whole pit profile suggests increasing accumulation in probably the last few decades (the length of the period cannot be deduced accurately because of the possible ablation of one or more budget horizons whenever the area fell below the firn line). During this period the equilibrium line and then the firn line has dropped below this altitude.

The two profiles of Pits 9 and 10 serve to illustrate the broken character of the firn line in an area where summers show a marked variation, a variation to which the firn line will closely respond.

Ice layering is evident throughout the winter snow of Pit 10 (300 m. a.s.l.), but pits dug to a depth of 1·20–1·80 m. above 500 m. a.s.l. showed an absence of ice layers. Together with the data from Russell East Glacier this suggests that non-freezing temperatures are the exception in winter above 400 m.

To complete the investigation of winter snow depths pits were dug to the glacier ice level at 100 m. a.s.l. on Russell East Glacier and also on the sea ice. In the latter case determination of the amount of accumulation was difficult because by October, when the measurements were made, some of the early snow had thawed and re-frozen on to the sea ice surface. The results are shown in Table VII.

TABLE VII

ACCUMULATION ON THE EAST COAST OF TRINITY PENINSULA

Locality	*Height* (m. a.s.l.)	*Accumulation, Winter 1959* (cm. water equivalent)	*Budget Year, 1958–59* (cm. water equivalent)	*Budget Year, 1957–58* (cm. water equivalent)
Russell East Glacier	700	87·8	40·6	?
Russell East Glacier	600	97·9	41·2	35·5
Russell East Glacier	430	101·4	35·0	27·5
Victory Glacier	300	—	—	20·3
Russell East Glacier	100	36·0	below firn line	
Sea ice	0	20·0	below firn line	

The water equivalents for the last three heights have been arbitrarily determined from the density of similar snow or firn layers in the first three pits.

c. *Tabarin Peninsula*

A profile characteristic of the *névé* of a glacier is that from a pit 1 km. (Pit 11) south of the south-facing edge of Brown Bluff at a height of 460 m. (Fig. 13). The accumulation for the budget year up to and including the first two weeks of December 1959 was 81·0 cm. water equivalent.

Three pits were dug between the 300 and 340 m. contours (Fig. 13) in the expectation that this would give a good indication of the height of the climatic firn line which was believed to lie between these two altitudes. In the highest of this series of pits (Pit 12; 335 m. a.s.l.) the ice surface was 1·12 m. below the snow surface and the firn cover represented 68·9 cm. of water. The firn profile above the glacier ice consisted of 38·0 cm. of winter accumulation and 74·0 cm. of firn and ice layers. The density of the firn increased with increasing depth to a maximum of 0·75 g./cm.^{3} just above the glacier ice. This site level must fall below the firn line only after a warm summer (cf. Pit 2 on Depot Glacier, and Pit 9 on Victory Glacier). The other two pits (13 and 14) at a height of 305 m. a.s.l. had a much thinner firn cover which suggests that the firn line often rises above 305 m. a.s.l. Ahlmann (1948) has interpreted pit profiles similar to the one at 335 m. as belonging to an intermediate zone between the accumulation area proper and the ablation area. He places the firn line at the lower edge of the intermediate zone. On Tabarin Peninsula the firn line lies above the two lower pit sites (300 m.) on a sufficient number of occasions to place that altitude in the ablation region of the ice piedmont.

The problem of the altitude of the climatic firn line unfortunately has no straightforward solution in this area. Despite the thin cover of firn and snow at Pit 14 in late October 1959 and at Pit 13 in early December, the two areas were never reduced to bare ice by the autumn of 1960, although the summer was one of the warmest on record. Similarly, between January 1958 and April 1960 the ice piedmont between Mineral Hill, Ridge Peak and Trepassey Bay was always snow-covered and bare ice only appeared in isolated areas. This phenomenon is effected by the transport of drifting snow to low altitudes by the south-south-west wind and by periods of snowfall continuing throughout the ablation period. These effects have already been considered for the ice piedmont between Hope Bay and Trepassey Bay (p. 27), where the south-south-west wind reaches its greatest violence. South of this on Tabarin Peninsula the south-south-west wind and the föhn wind are sufficiently strong to transfer a large percentage of the accumulation to lower levels but far less is transported out to sea. This process applies also in summer, although to a lesser degree, when, but for the effect of drifting snow, the lower levels of the glacier would receive no accumulation at a time when accumulation continues higher up the glacier. Therefore, particularly below the firn line, accumulation does not decrease regularly with decreasing altitude and the snow line does not necessarily move up the ice piedmont in the classical manner with the advance of the ablation season.

Superimposed ice is important in the maintenance of a balanced budget but the thinness of the snow and firn cover means that it forms principally in early spring, before the ice surface temperature reaches 0° C. There cannot be a very big time lag between the formation of excess melt water and the elevation of the ice surface temperature to 0° C. The ice surface temperature on 3 December 1959 at Pit 12 was −4·9° C and the upper 50 per cent of the snow cover had reached melting point, but no excess melt water had reached the ice surface to form a new layer of superimposed ice. In autumn with the influx of the cold wave the firn cover is at a lower temperature than the ice surface and melt water is frozen within this cover while melt water also continues to run along the ice surface either into crevasses or on to exposed ice below the local firn line where it freezes as superimposed ice.

The part of the ice piedmont south of Trepassey Bay has a more active regime than the ice piedmont between Hope Bay and Trepassey Bay. There is more crevassing on the more southerly of the two ice piedmonts despite a similarity of slope, and this indicates a higher rate of flow. The ice piedmont between Hope Bay and Trepassey Bay has been shown to be almost stagnant over much of its area, except for channelled lines of flow (p. 27), and calving at the ice front is a very infrequent phenomenon. This is certainly not the case on the ice piedmont east of Mineral Hill, Buttress Hill and Lizard Hill, and this greater activity is thought to be entirely the result of lower scouring and transporting activity of the south-south-west wind here in comparison with its effect on the ice piedmont between Hope Bay and Trepassey Bay.

d. *Islands in Prince Gustav Channel*

Work, other than visual observations, on both accumulation and the firn line was restricted to

Eagle Island (Fig. 1), one of the more northerly of the islands. Approximately 7·0 by 6·5 km. in size, it carries an ice cap which feeds a small valley glacier 2–3 km. long debouching into the sea. The whole of the island lies below the local firn line, as a pit dug on the top of the ice cap at 305 m. a.s.l. in October 1959 (Pit 15; Fig. 13) showed 38·0 cm. of very dense firn overlying 15·0 cm. of white (superimposed) ice, overlying dense glacier ice. The top of the ice cap seems to lie between the firn and equilibrium lines and no doubt in most years it has a slightly positive budget though loss of ice through movement and melt lower down-glacier must give the whole glacier a negative budget. A small island ice cap cannot benefit very much from snow drifting from higher altitudes, because of the small areal extent involved. In October 1958, when the snow cover on Trinity Peninsula was at a maximum depth, the valley glacier on Eagle Island was carrying only between 30·0 and 60·0 cm. of snow. The snow line begins to move up this valley glacier earlier than on the mainland to the north and west.

Visual observations on the ice caps of Corry and Vega Islands in October 1959 showed bare ice toward their summits and no doubt a similar situation to that found on Eagle Island exists there.

2. Movement

But for the loss of stakes this investigation would have been more comprehensive. In the case of Russell East Glacier and the ice piedmont between it and Victory Glacier the original observations were rendered useless because of the disappearance of all stakes over the 1958–59 summer, thus necessitating the fixing of a new set in April 1959.

The movement studies confirmed the increasing inactivity of glaciers with increasing distance south-east of the plateau and, although no figures are available for the glaciers of the west coast of Trinity Peninsula, their severe crevassing suggests a high order of movement. Table VIII shows this in comparative form. The line of stakes on Russell East Glacier crossed the up-glacier end of the narrow valley trough where there is a constriction and consequent acceleration of ice coming from the three main accumulation areas. Victory Glacier is only represented by a single stake situated above the firn line where there is very little constriction, and hence it shows a lower order of movement.

Table VIII

COMPARISON OF ICE VELOCITIES FOR TRINITY PENINSULA GLACIERS

	(m./day)
Russell East Glacier	0·53
Victory Glacier	0·25
Ice piedmont between Victory and Russell East Glaciers	0·10
Depot Glacier	0·08
Small valley glacier on Eagle Island	0·03
Ice piedmont between Hope Bay and Trepassey Bay	0·004

Depot Glacier, which is 64 km. to the north-east, shows a rate of movement one-seventh that of Russell East Glacier, though it should be pointed out that its accumulation area is much smaller. Similarly, however, the ice piedmont between Hope Bay and Trepassey Bay has a movement rate about one-twentieth that of the ice piedmont between Victory and Russell East Glaciers.

Distance from the plateau, i.e. the summit level of Trinity Peninsula, is the most important factor in connection with this decrease in glacier activity, as the föhn produces a precipitation shadow effect. Thus, the Eagle Island valley glacier movement rate is about one-third that of Depot Glacier and about one-twentieth that of Russell East Glacier. Calculations of ice thickness for the valley glacier on Eagle Island give a figure of 70 m. (p. 19) which is considerably thinner than for Depot Glacier. Calculations have not been made for Victory and Russell East Glaciers because their surface slopes were not known to a sufficient degree of accuracy but the velocity is such that it makes the thickness of these two glaciers much greater than either Depot Glacier or the valley glacier on Eagle Island. Clearly this cannot be attributed entirely to a precipitation shadow effect, since the size of the accumulation area is important, but the former has a vital influence.

The direction of movement on each of the glaciers is perpendicular to the contours and the only other point of interest is the comparison between the rates of movement of the ice piedmonts and the valley glaciers. The latter show the higher rates for channelled flow, a ratio of about 1 : 5.

3. TEMPERATURES

Temperatures were recorded in pit walls with "Rototherm" thermometers, and a hole drilled at the bottom of some pits gave temperatures for approximately a further metre. No temperatures were taken for depths below 4·5 m. and therefore the depth of penetration of the cold wave is not yet known.

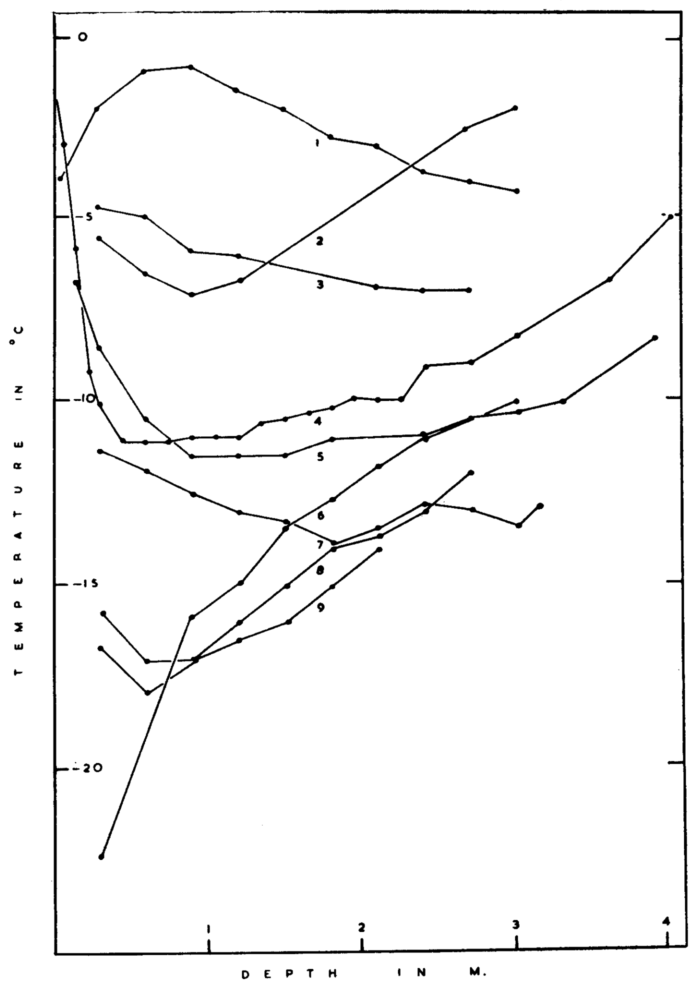

FIGURE 14

Temperature profiles for Russell East Glacier, Victory Glacier and the northern part of Trinity Peninsula.

1. Tabarin Peninsula, 460 m. a.s.l., 22 December 1959.
2. Victory Glacier, 350 m. a.s.l., 14 May 1959.
3. Cairn 3 (Fig. 1), 533 m. a.s.l., 16 March 1959.
4. Russell East Glacier, 430 m. a.s.l., 9 October 1959.
5. Russell East Glacier, 600 m. a.s.l., 14 October 1959.
6. Russell East Glacier, 700 m. a.s.l., 5 September 1959.
7. Russell East Glacier, 700 m. a.s.l., 20 October 1959.
8. Russell East Glacier, 700 m. a.s.l., 9 September 1959.
9. Russell East Glacier, 700 m. a.s.l., 11 September 1959.

It seems likely that below 500 m. a.s.l. the ice below a depth of 10 m. is at pressure melting point for the year, because the curves for 350 m. a.s.l. on Victory Glacier and 430 m. a.s.l. on Russell East Glacier in October are close to and trend towards 0° C (Fig. 14).

The temperature differences at a depth of 3·3 m. between the three Russell East Glacier heights (700, 600 and 430 m. a.s.l.) are rather large and in excess of the dry adiabatic lapse rate for air. A similar high adiabatic lapse rate has been found to exist in dry air whenever the westerly föhn is blowing.

VII. GLACIAL HISTORY

THERE is an abundance of evidence in the deglaciated areas of Trinity Peninsula and the islands of the James Ross Island group of a far more extensive glaciation in the past. *Roches moutonnées*, striae, erratics, perched blocks, abandoned cirques and moraines all indicate that ice was formerly considerably thicker on the mainland and on James Ross Island. Ice from both these centres filled Prince Gustav Channel and flowed north to find an outlet between Vega Island and Tabarin Peninsula.

1. Recent Recession

a. *Trinity Peninsula*

Studies on Depot Glacier and the ice piedmont between Hope Bay and Trepassey Bay have shown that recession during the present century is either non-existent or very slight. The snout of Depot Glacier provides the most useful information because it has been plane-tabled in 1945, 1956, 1957, 1958, 1959 and 1960. To these maps can be added a sketch survey by Duse in 1903, which, though it is of a lower standard of accuracy than those made by the Falkland Islands Dependencies Survey, serves to show the lack of terminal recession on this glacier during the present century. It also shows that the ice edges forming the north-west coast of Hope Bay and the landward termination of the ice piedmont between Hope Bay and Trepassey Bay were earlier in much the same position as in 1960.

While the terminal variation of Depot Glacier has been small since 1903, between 1945 and 1960 there has been a recession of 75–90 m. Much of this has been at the lateral extremes of the snout. This retreat is in fact very slight when compared with world-wide glacier recession outside Antarctica.

The ice edge forming the north-west coast of Hope Bay was plane-tabled in February 1959 and the resulting map was compared with the Hope Bay local survey of 1955. This, similarly, shows no overall recession though the ice edge of the important channelled outflow from Arena Glacier retreated 130 m. between 1955 and 1959. Arena Glacier has a comparatively large catchment area below the north-east face of Mount Taylor and therefore it has a fairly active regime. Its snout, which is the most active in the local area, pushed out beyond the 1955 limit during the winter of 1958 and must have exceeded a certain critical point of equilibrium, because during the summer 1958–59 it was the centre of very active calving and retreated over 130 m. in that period.

In 1959 plane-table maps were made of the southern snout of Victory Glacier and of the ice front immediately north of Pitt Point; photographs were also taken of the snout of Russell East Glacier. These have been compared with aerial photographs taken in 1956 by the Falkland Islands and Dependencies Aerial Survey Expedition and the results show there has been very little change in snout positions.

The overall evidence shows that the glaciers of Trinity Peninsula are in a state very close to equilibrium and probably have been so for the greater part of this century.

b. *Larsen Ice Shelf*

Unlike the glaciers of Trinity Peninsula, the front of the Larsen Ice Shelf has not remained static during the past few years but has broken back at least 15 km. in some areas north of Robertson Island. There is a certain amount of doubt as to the former positions of the ice front north of Robertson Island, arising originally from the position given by Nordenskjöld in October 1902 which he gave as running from Robertson Island in a northerly direction close to the southern end of Lindenberg Island. At the present time Lindenberg Island is well within the ice shelf and is the centre of an area of contorted and disturbed ice. A line of very wide and continuous rifts follows the line Nordenskjöld gave for the ice front and it seems possible that the rifts in the ice, running from Lindenberg Island to Christensen Nunatak, may have confused the issue of differentiating between his "low ice terrace" to the south and the sea ice he had travelled on from Cape Foster to Robertson Island.

In the mid- and late 1940's the ice front extended from the vicinity of Cape Foster in a south-westerly direction somewhere east of Cape Longing to Robertson Island. At Cape Longing the map prior to 1959 recorded "ice front not seen". Some time during the summer of 1957 the ice front broke back to Cape Longing and in the winter of 1958 it extended from Cape Longing in a south-south-west line to Robertson Island. North of Cape Longing it trended east-north-eastwards to Cape Foster. In November 1959 it was clear that the ice front had retreated still further and magnetic bearings were taken from the front between Cape Longing and Nygren Point and from Cape Longing on to the ice front running to the west, south of the cape. The point where it turns southward to Robertson Island was estimated. This showed that approximately 450 km.^{2} had broken out north of Cape Longing and a similar amount to the south. One of the main areas of calving south-west of Cape Longing had been formerly marked on maps as "crevassed and broken". Rifts in the ice shelf approximately 10 m. wide and running parallel to the ice front suggest that further extensive calving will occur in the near future.

Such a retreat is not in sympathy with the mainland glaciers to the north and this may be explained by the extension of the Larsen Ice Shelf beyond some point of equilibrium. It is interesting to note that, although the northern front of the ice shelf in George VI Sound (between Alexander Island and Graham Land) retreated an average of 2·4 km./yr. between 1936 and 1949, it altered very little in position between 1948 and 1958 (Procter, 1959).

2. Recession from the Glacial Maximum

Striae are evident in many of the ice-free areas and have been recorded together with the orientation of *roches moutonnées* to denote the local ice direction and former thickness of glacier ice.

a. *Hope Bay area*

The Hope Bay area (Fig. 15) is littered with angular frost-shattered debris. A large proportion of this debris has been derived from the Mount Flora massif and mostly represents the deposition of superficial morainic material. The moraines on Depot Glacier are superficial features nourished by frost-shattered debris falling from the valley sides and, since a large proportion of this falls in the ablation area, it remains on the glacier surface. The efficiency of glacier erosion under present (and probably some of the past) climatic conditions must be lower than that of periglacial agents such as frost-shatter. This largely explains the low ratio of ice-smoothed to angular morainic debris. A crevasse parallel to and 30 m. from the valley sides is often found on the less active of the Trinity Peninsula glaciers (e.g. on the valley glacier on Eagle Island and on Depot Glacier) and some show evidence of shearing between the ice adjacent to the valley side and the main body of ice. This suggests that, at the surface at least, ice in contact with the rock is almost static or very sluggish.

In the area between Seal Point, Lake Boeckella and the northern end of Scar Hills the exposed bedrock is almost devoid of striae though it does bear the smooth surfaces characteristic of glaciation. A dearth of striae in this area when there are such a great number immediately east of Scar Hills suggests that the area has been deglaciated for a much longer period. Striae are common again 100 m. from the landward edge of the ice piedmont between Hope Bay and Trepassey Bay.

The directions of glacial striae in the Hope Bay area are shown in Fig. 15; they indicate that Kenney Glacier, besides joining Depot Glacier near its present confluence, formerly extended on to Scar Hills. Scar Hills have undergone a tremendous amount of glacial plucking which has created steps 1 m. high on the north-east slope; these steps are not at right angles to the direction of movement suggested by glacial striae. Glacial erratics and perched blocks are now strewn over the surface of Scar Hills (Plate IIb). Some of the ice from Kenney Glacier flowed parallel to and south-east of Scar Hills and was joined by a marginal glacier at the foot of the north-facing slope of Mount Flora. Striae on the northern shoulder of Mount Flora at an altitude of approximately 400 m. a.s.l. suggest that ice from the cirque has at some time in the past spilled over to join the extended Kenney Glacier.

The valley walls of Depot Glacier itself bear the marks of a former greater thickness of ice. The sheer rock wall of The Steeple bears no striae but its smoothness and a sharp break in slope just below the ridge top is strongly suggestive of ice up to this level, which demands an increased thickness of 130 m. of ice at this point. Striae on this ridge and the spur extending from its northern end exist only below 273 m. a.s.l.

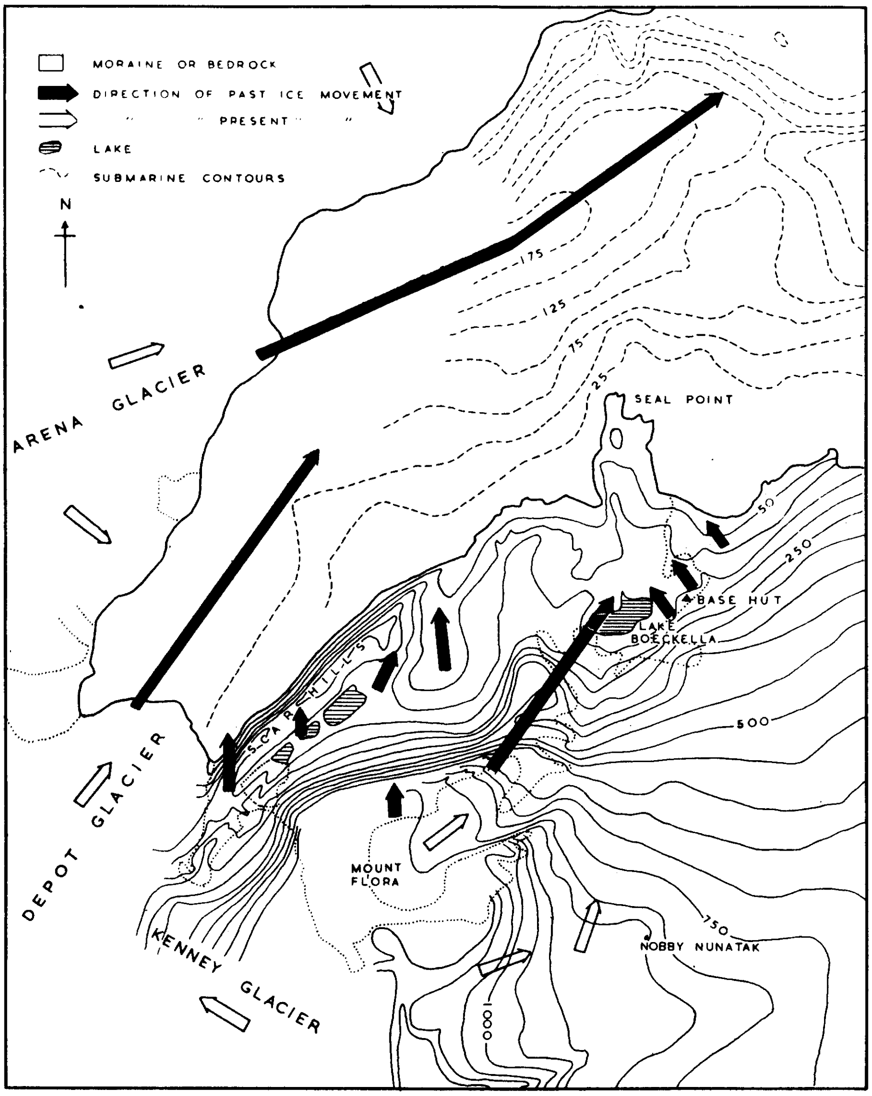

FIGURE 15
The direction of past ice movement in Hope Bay and on the ice-free area south of Hope Bay. Contours are not shown above 850 ft. (259 m.) on Mount Flora. Altitudes in ft. Submarine contours in fathoms.
(Map by R. R. Kenney, 1955.)

The south-east spur of Whitten Peak has been truncated by a former "Depot Glacier" about 140–170 m. thicker than at present, and a glacial overflow channel (Eddy Col) points to a small part of the ice now flowing into Duse and Hope Bays joining Arena Glacier when the ice was thicker. At this stage of glacierization Arena Glacier appears to have been the dominant glacier in Hope Bay, because the submarine topography of Hope Bay consists of a deep trough aligned more towards Arena Glacier than Depot Glacier, which must have been deflected slightly to the north-east.

i. *Mount Flora cirque moraines.* The Mount Flora cirque provides the most detailed evidence of glacial history, as it is fronted by a series of ice-cored moraines (Plates IIIa and b). Frost-shatter, the most important erosional process in this area, provides an abundant supply of material in the summer to the cirque ice which is more liberally strewn with rock than any of the surrounding glaciers. As a result of this a comprehensive series of moraines with steep north-facing slopes has been built up. A plane-table map of the two moraines nearest the cirque was made on a scale of 1 : 5,000 and the moraines were differentiated into terminal and lateral lines. Beyond these two moraines differentiation is complicated by the ice piedmont which has itself transported some of this material farther north (note the northward trend of the western end of the moraines).

A linear collection of blocks and boulders to the east of Lake Boeckella and the northernmost moraines on the ice piedmont are assumed to represent the maximum extent of the first major advance of the cirque ice. The other moraines represent two stages of retreat or re-advance. Each of these moraines consists mainly of cobbles and boulders and on the steep forward slopes any small material has been washed to the base of the moraine or out via a melt stream. Much of the morainic material is segregated into individual groups of blocks and boulders, the whole feature measuring as little as 10 m. by 3 m. and standing 4 m. above the general ice level due to the protection from ablation afforded by the rock. Some of these, consisting of one rock type, are assumed to represent the shattering of one block. The terminal moraines are composed principally of separate rock mounds, the north-east-facing slope being very steep and close to the maximum angle of repose for boulders. The lateral moraines, however, are more continuous possibly because of less dissection by superficial drainage.

Two lines of moraine which occur at the foot of Kenney Glacier east of Scar Hills are believed to be contemporaneous with the youngest of the Mount Flora cirque moraines. Although some of the moraines at the foot of Kenney Glacier have no ice cores, they have a similar rock composition to the youngest cirque moraine. A considerable amount of morainic rubble submerges the southward continuation of Scar Hills and spills over on to Depot Glacier, which has transported some of the material to the sea.

The alignment of the striae from the south-east to north-west (Fig. 15) indicates that the ice piedmont between Hope Bay and Trepassey Bay has retreated very little in its landward termination in comparison with the retreat of the Mount Flora cirque. This is in fact what is to be expected and it is the lee and shaded position of the cirque which has prevented a more rapid recession. It is of interest to note that a cirque on Eagle Island with a very similar aspect, elevation and outline to the Mount Flora cirque is completely ice-free, no doubt because of its more easterly position.

ii. *Pluvial interglacial.* Gullying is apparent on many of the higher rock outcrops, particularly the west-facing slope of Dimaryp Peak. A formation of this nature is unlikely to develop under the present climate when there is very little free water in summer to erode such gullies. These gullies start quite close to the ridge top but the lower parts of the small ridges separating each gully have been planed off by re-advancing ice in the cirque west of Dimaryp Peak.

The broad and shallow valleys which now carry the sporadic summer melt water are likely to have been formed at this period, as the streams which occupy them in summer are misfits and run for parts of their courses on ice, where no erosion of rock can take place.

The relationship of this pluvial interglacial to the moraines is difficult to establish. Dissection of the two youngest moraines below the Mount Flora cirque brings the melt water into contact with ice which erodes at a vastly different rate to rock. These melt channels are younger than those of the pluvial stage and it may tentatively be suggested that the latter were initiated some time after the first cirque moraine was deposited but before the second. The truncation of the base of the small inter-gully ridges similarly suggests a date preceding the formation of the youngest of the Mount Flora cirque moraines. The thickness of ice which existed at the time of formation of the oldest cirque moraine must have greatly exceeded the thickness of ice which truncated the Dimaryp Peak gully ridges. This suggests that the glacier re-advance, which truncated the ridges formed in the pluvial interglacial, was contemporaneous with the advance of the cirque which formed the second of the three Mount Flora cirque moraines. The pluvial stage immediately preceded this advance.

b. *View Point area*

In the View Point area (Fig. 16), 27 km. south-west of Hope Bay, a second group of ice-free rocks which bear the marks of glaciation is exposed. Notes on the alignment of striae and the orientation of *roches moutonnées* were used to interpret the past glacial history. Glacial striae were plotted by magnetic compass and formed three groups. One set at 115–135° mag. has been overridden by the second set at 140–150° mag. It is suggested that the former were made when ice moved up Prince Gustav Channel (though no evidence in the vicinity of View Point has been found to indicate this) and deflected the ice flowing over the View Point area. This ice was itself of sufficient thickness to move independently of local surface features. With a lowering of the level of ice in Prince Gustav Channel the View Point stream could then flow undeflected and join the channel flow via an ice fall. The fact that the older set of striae still exists and has not been scoured away by the later flow of ice suggests that the period of undeflected flow was short and represents a rapid retreat.

This was followed by further retreat of the ice until the higher land was devoid of ice and the relief exerted itself sufficiently to channel ice down the shallow valley flowing past the nunatak on which the View Point station stands. This ice is now believed to be stagnant; the whole area lies below the firn line and the ice (deduced from soundings at the ice edge) is only 25–30 m. thick.

The above sequence of events would demand an increase of at least 300 m. in the thickness of ice in the valley immediately west of View Point. It would also demand ice in Prince Gustav Channel up to a height greater than 300 m. a.s.l. This consequent thickening does not exceed estimates given for other localities in Graham Land (Nichols, 1960).

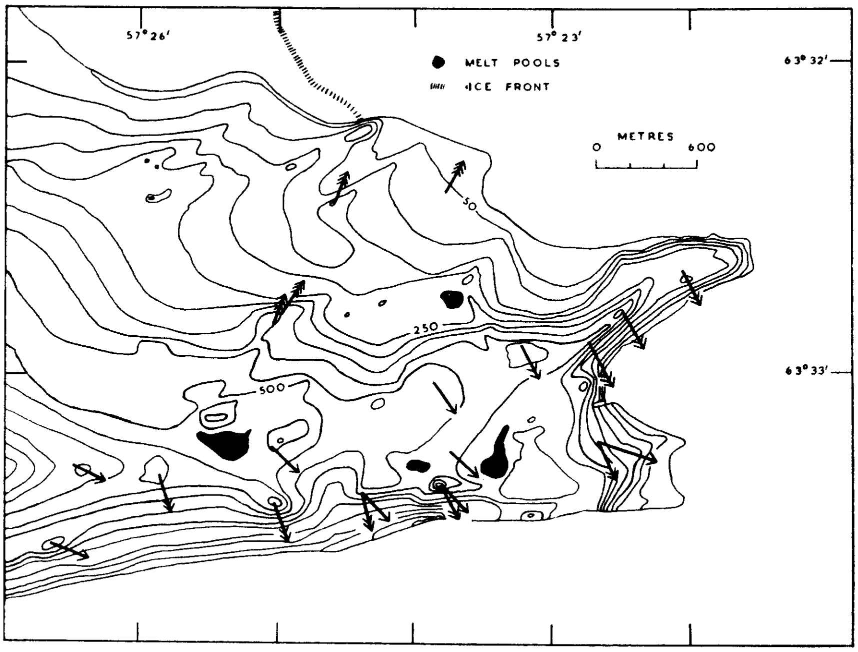

FIGURE 16
Direction of past ice movement in the vicinity of View Point. Altitudes in ft.
The oldest set of striae are represented by arrows with one barb, the youngest by three barbs.
(Map by J. Madell, 1956.)

c. *Islands in Prince Gustav Channel and the adjoining mainland*

Farther south down Prince Gustav Channel there is additional positive evidence for a former channel ice stream. Glacial striae, smoothing and erratics on Long Island point strongly to the extension of Victory Glacier over this island; this indicates a minimum height of 160 m. above present sea-level for this ice. Erratics of mainland origin have been found on the west and north-west coasts of James Ross Island (Bibby, 1961), which illustrates the advance of mainland ice across Prince Gustav Channel as it moved north. Until an accurate knowledge of the distribution of erratics on James Ross Island is gained, it will not be possible to determine how far mainland ice encroached on to James Ross Island. An ice shelf still occupies the southern part of Prince Gustav Channel and the Sjögren Glacier tongue extends across it to Persson Island, which was itself once covered by this same ice stream, i.e. to a height of at least 130 m. In a similar way an ice shelf in the northern part of Prince Gustav Channel could transport mainland morainic material to the southern areas of James Ross Island.

Evidence for the former extent of ice on the mainland is scanty because of the inaccessibility of some nunataks and the extensive frost-shatter that has destroyed much evidence on others. On Azimuth Hill there are striae in line with the present ice flow direction at a height of 50 m. above the present ice level and the Pitt Point outcrop is similarly striated 30 m. above the ice. The angularity of the other nunataks bears witness to the amount of frost-shatter they have undergone.

The retreat subsequent to the major ice advance was itself followed by a re-advance due to lowering of the firn line. This sequence is deduced from the existence of the following series of cirques noted on the islands of Prince Gustav Channel:

i. Two at the northern end of Egg Island with base levels at 30–40 m. a.s.l.
ii. One on Corry Island with a base level at sea-level.
iii. One at the southern end of Eagle Island.
iv. A very broad, shallow cirque feature on Tail Island.
v. A well-developed cirque on the south side of Holluschikie Bay with a base level 30–40 m. a.s.l.
vi. Two small cirques on the south side of the bay between Rink and Stoneley Points with base levels 30–40 m. a.s.l.

All of these cirques, with the exception of the one on Tail Island, are quite well developed with steep head walls and morainic debris strewn on the cirque floor. All except the cirques between Rink and Stoneley Points and in Holluschikie Bay are inactive as glaciers and very little dead ice is present. From a view point in the bay the cirque in Holluschikie Bay appeared to be slightly active with a very negative budget.

"Cirques form at or close to the firn limit because they demand physical conditions that occur only near that limit" (Flint, 1957), i.e. the orographic snow line lies at or slightly above the cirque floor. The level of the floors of these cirques may therefore be assumed to be the height of a former firn line. These cirques cannot be contemporaneous with the maximum glacial stage, since the area containing them was at that time covered by foreign ice which would have greatly modified them if they had been formed prior to that stage. They are not extensively developed and the length of the stage would appear from this not to be of great length yet sufficiently recent to permit two of them to retain some dead ice. This glacial stage may be contemporaneous with the oldest or the second of the Mount Flora cirque moraines. From this stage there has been a fluctuating retreat, the firn line rising 200–300 m.

VIII. CONCLUSIONS

GLACIERS on Trinity Peninsula have shown few signs of recession during this century. A study of the regimes of Depot Glacier and the ice piedmont between Hope Bay and Trepassey Bay strongly suggests that accumulation exceeds ablation by sublimation, evaporation and run-off, and that a state of equilibrium is maintained by calving of the ice front. Evidence of the relative rate of calving is found in the amount of calved ice incorporated in the sea ice off various glacier snouts. This shows that Russell East and Victory Glaciers calve much more than Depot Glacier and accordingly have more active regimes. Similarly, the relatively inactive regimes of the ice piedmonts are reflected in their low rate of calving.

The decrease from the maximum glaciation has seen the removal of at least 300 m. of ice from some areas and considerably more from Prince Gustav Channel which acted as an outflow channel for the east coast mainland ice and the ice from the west coast of James Ross Island. This can be compared with the removal of 700–1,000 m. of ice from the Marguerite Bay area (Nichols, 1960). No evidence has yet been found of the complete swamping of the area with ice. Subsequent advances and retreats have been on a much smaller scale.

The cause of the most recent ice retreat, during which the Mount Flora cirque has retreated from its second moraine, seems to have resulted from a shift in the track of depressions. The amount of winter precipitation is not the operative factor in the determination of the height of the firn line at the end of the ablation period. This is mainly dependent on three factors:

i. Snowfall through the summer.
ii. The frequency of summer föhn winds.
iii. In the vicinity of Hope Bay the frequency and violence of the south-south-west wind.

The ablation season is rarely continuous for more than a week at a time and may be held up by periods of sub-freezing weather, when the bare ice is covered with a protective mantle of cold snow. In the summer of 1959–60 there were two such periods in the months of January, February and March. The second of these periods arrested the lowering of the ice level for three weeks.

The föhn wind in summer raises the temperature above 4° C and lowers the relative humidity to that for dry air (less than 60 per cent). It is associated with clear skies along the east coast, and altocumulus lenticularis is the most common form of cloud, although this covers only 2–3 oktas of the sky. Therefore, the most important ablation periods occur during föhn conditions.

An increase in the frequency of these winds must cause an increase in the ablation. The decrease in glacierization south-eastwards from the plateau has already been attributed to the föhn effect. There is, however, little evidence of recent glacier recession on the west coast of Trinity Peninsula. It is thought that this contrast in the recessional characteristics between the east and west coasts of Trinity Peninsula has been brought about by an increase in the frequency of westerly warm sector winds. This increase must have been brought about by a change in the tracks of depressions mainly southwards (Smith, in press) and also eastwards, so that more depressions have followed a southerly track in the Weddell Sea. While such an increase would promote recession on the east coast it would not have the same effect on the west coast where westerly winds bring orographic precipitation.

IX. ACKNOWLEDGEMENTS

THE writer is indebted to personnel of the Falkland Islands Dependencies Survey station at Hope Bay between 1958 and 1960 for the general assistance they gave on sledge journeys and also for pit digging carried out in areas not visited by the author. The surveyors were most co-operative and loaned surveying equipment and advised the author on its use. Without this assistance movement studies would not have been possible. The mathematical interpretation of glacier slope and velocity to ice thickness was kindly undertaken by J. Ashley. Discussions with Dr. J. W. Glen and Professor D. L. Linton have been very helpful. Thanks are due to Dr. R. J. Adie who supervised the work and who deserves most of the credit for its presentation. The facilities of the Department of Geology were kindly made available by Professor F. W. Shotton. Finally, gratitude should be expressed to D. McCalman who encouraged the author to pursue glaciological research and who organized duties at the Hope Bay station to allow a maximum amount of time in the field.

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PLATE I

a. The path of a slusher burst on the ice piedmont near Hope Bay, showing the re-frozen melt-water surface superimposed on the bed of the slusher stream. Note also the high back wall of the burst. 21 July 1958.

b. Rime ice lobes facing the windward (north-west) direction below Mount Canicula at the head of Russell East Glacier (750 m. a.s.l.). 18 October 1959.

a

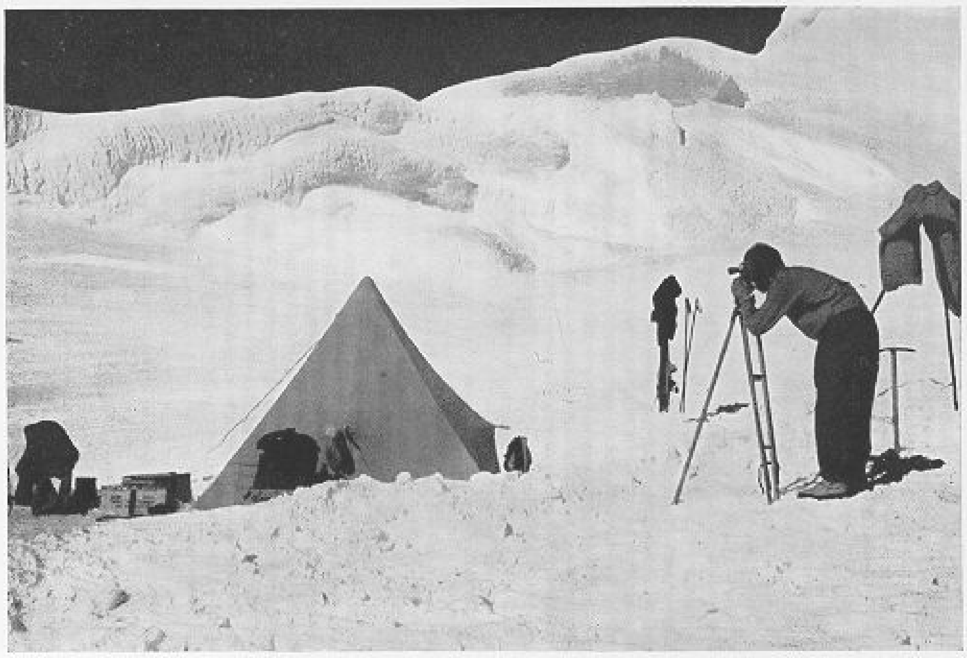

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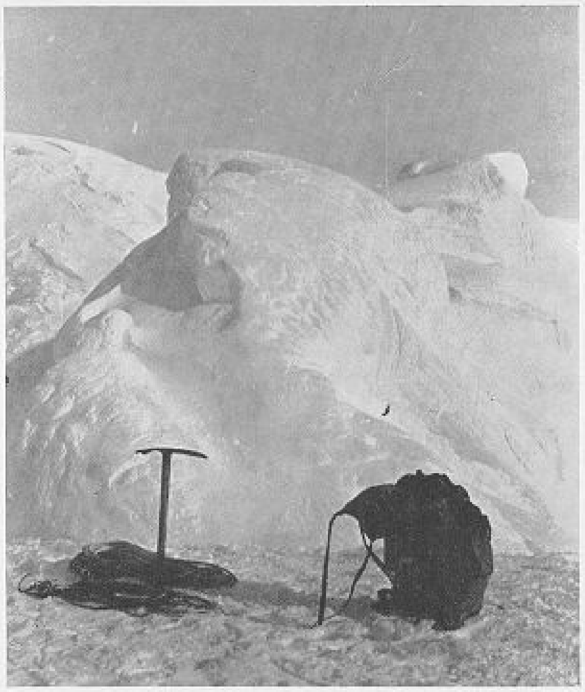
PLATE II

a. A 75 ft. (22·9 m.) high lobe of rime ice near Sirius Knoll (900 m. a.s.l.), Trinity Peninsula. Photograph taken from a similar lobe approximately 200 m. distant. 20 September 1959.

b. The snout area of Depot Glacier, showing the crevassed area and moraines; viewed from the east on the north-west slope of Mount Flora. Scar Hills lie at the foot of the steep slope in the foreground. 10 February 1960.



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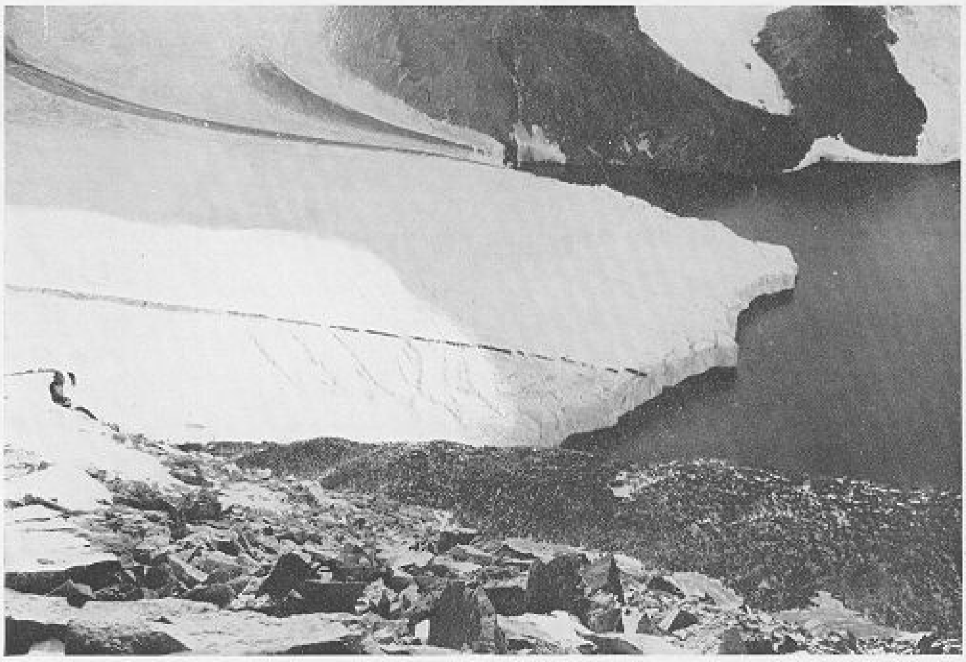

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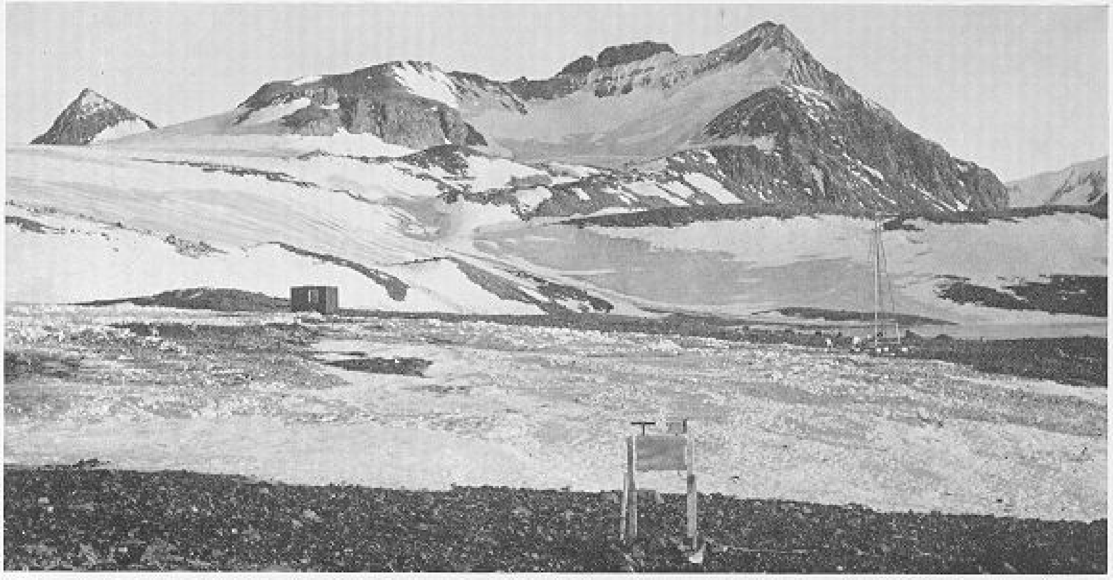
PLATE III

a. View of the ice-free area south-east of Hope Bay and the margin of the ice piedmont from the upper slopes of Mount Flora. The hummocky ground in the middle distance is formed of old moraine from the Mount Flora cirque glacier. 11 February 1959.

b. Mount Flora and its cirque glacier; viewed from the Hope Bay station. 20 July 1958.

a



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