Some aspects of the glaciology of the mars ice piedmont, Anvers Island, Antarctica

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SOME ASPECTS OF THE GLACIOLOGY OF THE MARR ICE PIEDMONT, ANVERS ISLAND, ANTARCTICA.

By

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VOLUME ONE

TEXT AND ILLUSTRATIONS

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ABSTRACT

The results of a comprehensive three-year study of the Marr Ice Piedmont, Anvers Island, Antarctica are presented.

The piedmont stands on a low coastal platform ranging from slightly below sea level to 200 m. a.s.l. Ice thickness ranges from 60 to 80 m. at the coastal cliffs to more than 600 m. inland.

Annual accumulation is high. There is a strong relationship between elevation and accumulation rates and a marked variation of accumulation rates from year to year.

Surface ice velocities range from 14 m/year to 218 m/year and there is considerable ice streaming as a result of the subglacial topography.

The mass balance of a representative part of the piedmont is considered to be in equilibrium or possibly, slightly positive. A study of a peripheral ramp shows annual fluctuations of balance and it is hypothesised that there may be a long-term tendency towards a positive regime.

Ice core studies indicate that there is no dry snow facies but all other facies are identified. The saturation line lies at approximately 600 m. a.s.l. and the equilibrium line ranges from 60 to 120 m. a.s.l.

Englacial ten-metre temperatures range from -0.8 °C near the coast to -4.9 °C inland.

Deformation velocities have been calculated and basal sliding velocities inferred. It is hypothesised that basal conditions are not everywhere the same and that parts of the piedmont are frozen to bedrock.

It is suggested that basal sliding and erosion are related and that the piedmont is selectively eroding its bed and accentuating the subglacial topography. Evidence of erosion, debris-rich ice, exists in the piedmont but is below sea level at the coastal cliff. The piedmont is not a "Strandflat Glacier" which is cutting a planed surface at a level controlled by the sea.

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To

ANTARCTICA

my Great White Mistress,
whom I shall always love

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CHAPTER ONE

INTRODUCTION

Anvers Island is located off the western coast of the Antarctic Peninsula between latitudes 64° and 65° south and longitudes 62° 30' and 64° 30' west. It is the largest and most southerly member of the Palmer Archipelago and is rectangular in shape, approximately 60 km long in a northeast-southwest direction and 45 km wide. It has an area of about 2,700 km^2 (figure 1).

Physiographically, Anvers Island can be divided into two distinct parts. The eastern side consists of heavily glacierized mountainous terrain, which is dominated by Mt. Francais, which rises to 2,750 m. The Achaean, Trojan and Osterreith Ranges trend northward and eastward from this point and form a discordant northeast coastline. From Mt. Moberly (1,660 m.), a fourth range trends southwards. Within this mountain system are several valley glaciers, the most notable of which is the Iliad Glacier which flows northward from Mt. Francais to Lapayrere Bay (figure 1).

The entire length of the western side of the island is occupied by an unbroken ice cover, known as the Marr Ice Piedmont (figure 2). This piedmont ranges in width from 7 to 25 km and accounts for approximately half of the total island area. From its terminal ice cliffs on the western and southwestern coastlines, the snow surface rises toward the mountain walls and reaches a maximum elevation of some 850 m.

LOCATION AND PHYSIOGRAPHY OF ANVERS ISLAND

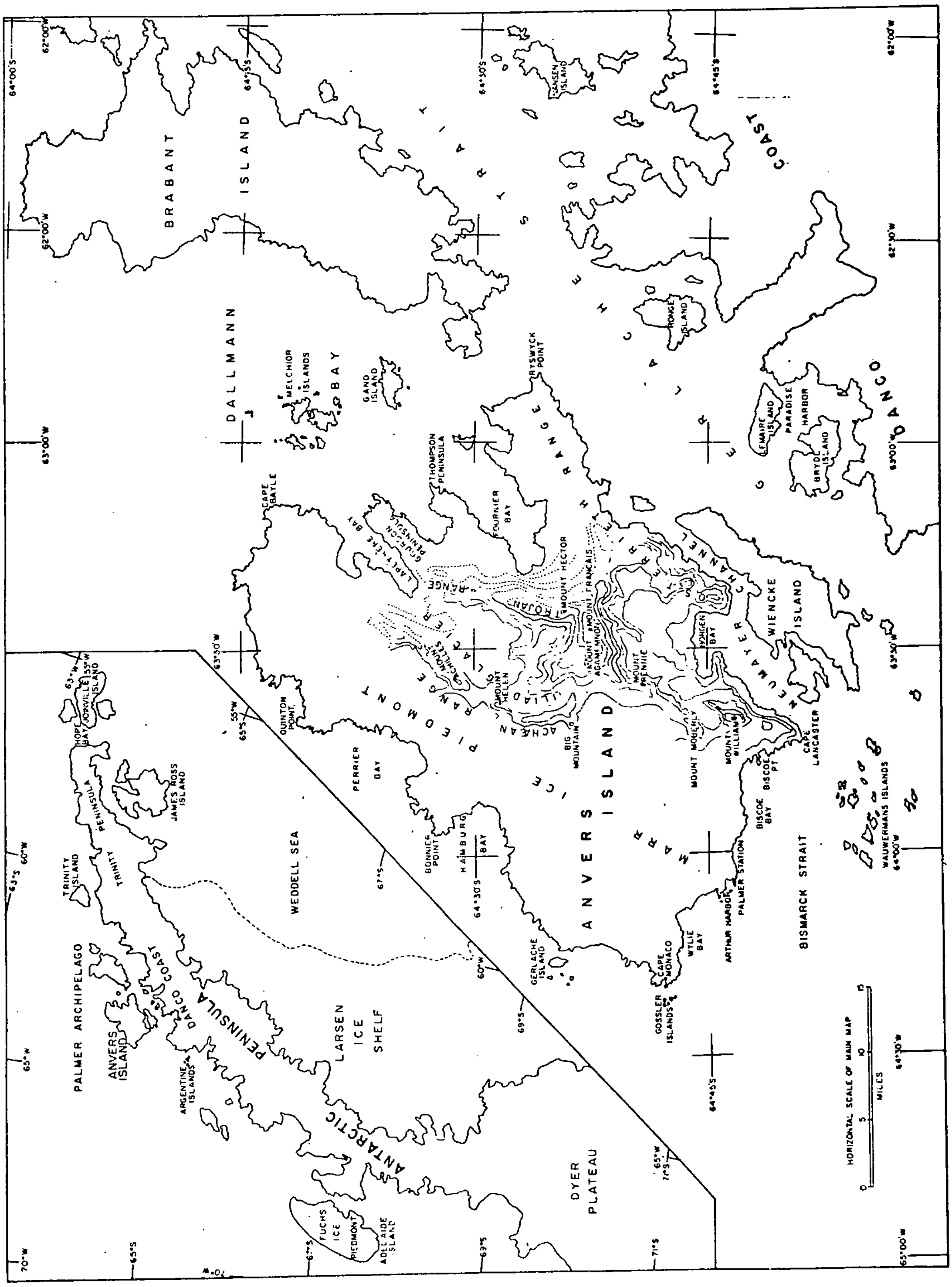


Fig. 1

Fig. 2. Marr Ice Piedmont. View looking east. Note: Sharp physiographic division between mountain ranges and the piedmont, the peripheral ice cliffs, crevasse fields and ice ramps. (US Navy photo)

The surface of the piedmont is not uniformly planar or convex but shows quite pronounced topography. This is particularly evident in the Perrier Bay area and in the southwestern part, where pronounced swells and hollows are frequent. These topographic features, though evident to some degree close to the mountains, are particularly prominent in the central and coastal areas of the piedmont. To a considerable degree, they reflect the subglacial topography, especially in areas where the ice thickness is less than 300 m. The peripheral 2 to 5 km. are heavily crevassed (figure 3) and vehicular access to the inland regions is difficult. In many instances, the pattern of the crevasses in this peripheral zone can be related to the direction of ice movement.

For the most part the ice piedmont terminates in cliffs ranging in height from 20 to 80 m. above sea level. These cliffs pass into small ice ramps in only a few isolated locations where the piedmont terminates on land promontories (figures 2 & 5).

Within the general descriptive framework of the west coast piedmonts, the Marr Ice Piedmont is a good example of its type, though a precise morphological classification according to Ahlman (1948 p 67-69) is not possible due to the lack of topographic information. However, from the limited information which is available from this study, it is probably of Ahlmann's type "A" with an area-altitude graph of the general shape shown in Figure 4.

Ice thickness measurements published by Dewart (1971) and discussed by Rundle (1971) indicate that it is in fact an ICE PIEDMONT, defined by Armstrong et al (1966) as follows:

Fig. 3. Heavily crevassed peripheral ice typical of much of the piedmont. View in vicinity of Wylie Bay, looking NNE.

" Ice covering a coastal strip of low lying land backed by mountains. The surface of an ice piedmont slopes gently seawards and may be anything from 1 to 50 km wide, fringing long stretches of coastline with ice cliffs. Ice piedmonts frequently merge into ice shelves. A very narrow ice piedmont may be called an ice fringe".

GEOGRAPHIC CONTEXT OF ANVERS ISLAND

The Antarctic Peninsula, because of its latitudinal extent and elevation range, exhibits a variety and progression of glaciological conditions and forms an important link between the temperate glacier category in the north and the truly cold polar West Antarctic Ice Sheet to the south. It contains an almost complete range of morphological types from small cirque glaciers and large valley glaciers to ice caps ice piedmonts and ice shelves.

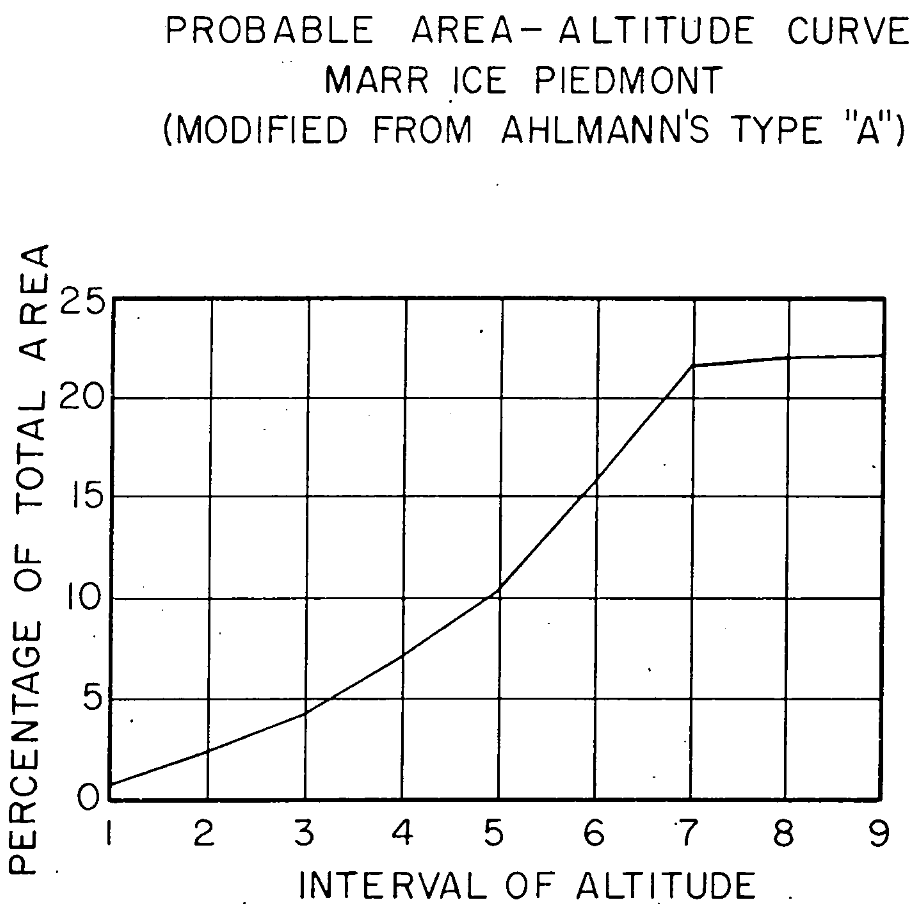

PROBABLE AREA-ALTITUDE CURVE
MARR ICE PIEDMONT
(MODIFIED FROM AHLMANN'S TYPE "A")
PERCENTAGE OF TOTAL AREA
25
20
15
10
5
0
1
2
3
4
5
6
7
8
9
INTERVAL OF ALTITUDE

Fig. 4

Fig. 5. View of Biscoe Bay showing peripheral cliffs, crevasses and Biscoe Point ice ramps (center right).
(US Navy photo)

Its position relative to the cold Weddell Sea in the east and the warmer Bellingshausen Sea in the west, give it characteristics peculiar to itself from the point of view of climatic as well as latitudinal influences on glaciological regimes. For 1,300 km. the eastern shoreline borders the Weddell Sea, which with its massive concentrations of pack ice, acts as a "reservoir of cold" (Linton 1964) and profoundly influences the nature of the east coast glacierization. Thus, the ice shelves which characterize and dominate the eastern seaboard begin as far north as 64° S. and culminate in the massive Larsen Ice Shelf.

By contrast, the Bellingshausen Sea is warmer than the Weddell Sea by as much as 5° C. at similar latitudes, and along the 1,100 km. western coastline ice shelves are not evident until about latitude 69° S. and the Wordie Ice Shelf is reached. North of this latitude, ice caps, ice piedmonts and sea-calving glaciers are characteristic and none of these extends far, if at all, beyond their individual bedrock coastlines. An important, climatically controlled boundary line has been suggested by Robin (1964) to exist along this coast and represents the latitude at which sub-polar glaciers are able to terminate on land. This location is likely to be extremely sensitive to climatic variation and change.

The southerly penetration of the Antarctic Convergence below 60° S. is a dominant climatic influence and meteorological and associated distributions run generally parallel to it. The air temperature along the entire western coastline can rise above the freezing point in all months of the year.

The limited precipitation records from the peninsula also indicate contrasts between the two coastlines. On the eastern side, as little as 10 g/cm^2/ year is suggested by Linton (1964) while on the western side 10 to 20 times this amount is not uncommon. The highest value on record can now be reported from Anvers Island, where the average value for the three-year period 1965-1968 reached over 230 g/cm^2/ year at 850 m. above sea level (Rundle 1967, 1970). A precipitation gradient is evident from northwest to southeast and Linton (1964) suggests a rising snowline from northwest to southeast which is generally opposed to the normal rise of the snowline with decreasing latitude. The deposition of rime by condensation is common throughout the peninsula and reaches proportions large enough to be an important factor in glacier budgets in Trinity Peninsula.

Thus the Antarctic Peninsula emerges as a complex geographic province in which the interaction of glacio-climatic factors must be considered in its overall geographic description. Many individual glaciological and climatological problems await solution in the area and the underlying intent of the work described in this thesis is to shed light on some of the answers.

PREVIOUS GLACIOLOGICAL WORK IN THE ANTARCTIC PENINSULA

Before the turn of the present century much descriptive material had been collected by whaling and sealing ships and by exploratory voyagers but the first glaciological observations in the Antarctic Peninsula were made during the Belgian Antarctic Expedition and Arctowski (1900) briefly described the glaciation of the area. This was followed by the Charcot expeditions, 1903-1905, and Gourdon (1908) made detailed glaciological observations and, from Anvers Island, described "vast flat glaciers descending from Mt. Francais and the Osterreith Range - - - " (p 108). In the following decade the British Imperial Expedition visited the Danco Coast (Lester 1923), and during the Norwegian Antarctic Expedition, Holtedahl (1929) examined the physiography and glaciology of the region and concluded that many of the ice features of the peninsula's west coast were "strandflat" glaciers.

During the British Graham Land Expedition, 1934-1937, the glaciology of the western coast of Graham Land was further investigated by Fleming (1940) who believed that the present ice caps and "fringing" glaciers had been reduced to their present form from a previously more extensive ice sheet. From observations at Argentine Islands he believed that ablation far exceeded accumulation and concluded that the present ice caps and glaciers were ephemeral features, out of phase with the prevailing climatic conditions and would fast disappear.

Following the second world war, the Falkland Islands Dependencies Survey established the scientific station at Argentine Islands in 1947 and a continuous meteorological record has been kept to the present time. Glaciological work was carried out by Roe (1960) from 1958-1960 on Galindez, Winter and Uruguay Islands. He suggested that the formation of superimposed ice, even during a warm summer, is important in the maintenance of the Argentine Islands' ice caps. From measurements between June 1958 and February 1960, he concluded that ice accumulated at the rate of 5.0-7.5 cm/year but because this period covered anamolous meteorological conditions, he believed his observations were not likely to be typical and that if the observations were made over a longer period (10 years) the "picture - - - would probably be much modified". Measurements over a five-month period showed ice movement rates of 20-50 cm./year, 30 meters from the edge of the ice cliff.

In the South Shetland Islands, the establishment in 1948 of a base at Admiralty Bay, King George Island, led to glaciological observations by Hattersley-Smith (1948, 1949) who concluded, from the very low movement rates observed, that "Camp Glacier" is very inactive. Four accumulation poles set out by Chaplin (1949) showed snowfall of 186 cm. between May and October 1949. Similar observations by Stansbury (1961a) in 1959 showed accumulation of 140 cm. during the winter. Noble (1959, 1965) assessed the budgets of Flagstaff Glacier and Stenhouse Glacier from March 1957 to March 1958. The budget of Stenhouse Glacier was almost balanced but Flagstaff Glacier had a decidedly negative budget. High movement rates were observed on Stenhouse Glacier; 1 meter/day near the terminal cliffs and 0.5 m/day up-glacier. Movement of Flagstaff Glacier was too slow for short-term measurement and the results of annual movement studies are regarded as unreliable (Noble 1965, p 9).

Stansbury (1961b) studied the Hodges Glacier, South Georgia, from November 1957 to September 1959 and concluded that this glacier had a markedly negative budget, continuing a slow trend of galcier shrinkage of the past few years. Movement rates were high for this cirque glacier; 0.5 cm/day near the snout and 62 cm/day on the steep back walls. From Signy Island, South Orkney Islands, Maling (1948) reported that rime formation above the cloud base is an important accumulation factor and because it also forms on rock outcrops, it increases the albedo and retards ablation.

In Trinity Peninsula, extensive glaciological work was carried out by Koerner between December 1957 and April 1960 (Koerner 1961, 1964). He concluded that the glaciers in Trinity Peninsula have shown few signs of recession during this century. From a study of the regimes of Depot Glacier and the ice piedmont between Hope Bay and Trepassy Bay, he concluded that accumulation exceeds ablation by sublimation, evaporation and runoff and that calving of the ice front is responsible for maintaining equilibrium. Accumulation by rime deposits is important on the ice piedmonts of the west coast of Trinity Peninsula and on Joinville Island.

Thomas (1963) carried out more detailed glaciological observations on Galindez Island between August 1960 and January 1962. Snow depth and density were measured every few months at a rectangular stake pattern and the study showed that over a number of years the budget state of Galindez Island was one of equilibrium with minor oscillations in surface level related approximately to the cyclic variation in the annual mean temperature. Continuation of this work by A.J. Scharer (see Sadler 1968) to March 1963 showed that mean annual net accumulation from 1958-1963 was 20 g/cm^2 at one site.

Sadler (1968), continuing the work on Galindez and Skua Islands, showed that 66% of the gross accumulation was removed by ablation and that 22% of this was the result of calving at the ice cliffs. A net accumulation of approximately 19 cm. of water equivalent was recorded from which it is apparent that these islands' ice caps are not in equilibrium but have slightly positive budgets. In the Argentine Islands the very small annual ice movement is too low to allow calving to create equilibrium.

Bryan (1965, 1968) and Dewar (1967) have conducted glaciological investigations on Adelaide Island and it appears that the Fuchs Ice Piedmont is at or close to equilibrium. Whether it is a relic from a formerly more extensive ice sheet is not known but from the high rates of accumulation and the apparently active state of the regime, that ice piedmont seems capable of independent existence.

PREVIOUS WORK ON THE ICE PIEDMONTS AND THEIR GEOMORPHOLOGICAL SIGNIFICANCE

First serious attention was drawn to the ice piedmonts by Holtedahl (1929) who, after investigating the Marr Ice Piedmont on Anvers Island, suggested that the masses of foreland ice on the exposed coast of Graham Land are strandflat glaciers that have cut and are still cutting planed surfaces or "strandflats" at a level controlled by the sea. He regarded the Marr Ice Piedmont, the Fuchs Ice Piedmont, the Handel and Mozart Ice Piedmonts as mature strandflat glaciers which have eroded all nunataks they may have surrounded and have almost completely cut away their mountain headwalls in areas such as north Adelaide Island. He believed the strandflats were not cut on the less exposed coasts because the fjords were filled with stagnant ice where full cutting power was not developed.

He suggested that the Marr Ice Piedmont was once more extensive and that it may at some time in the past have been supported by the sea in places (i.e. like an ice shelf). The Fuchs Ice Piedmont is thought by Dewar (1967) to have been thicker and more extensive and to have reached seaward for at least 11 miles (18 km.) to the Amiot Islands.

Fleming (1940) concluded that the present ice piedmonts and ice caps along the western coastline were the remaining land supported margin of a former coastal ice shelf and that the seaborne area had broken away and been lost in comparatively recent years. He did not however, discuss the origin of the sea level platforms on which the ice piedmonts rest except that he found no evidence that these "fringing" glaciers eroded their own platforms. In fact, he reported that they had been found resting on, but not eroding, unconsolidated beach deposits (p 98).

A significant feature of the ice piedmonts and other "fringing" glaciers is that they generally terminate in sea-calving cliffs at their own individual bedrock coastlines and bedrock is generally visible at the base of the ice cliffs. Englacial moraine has rarely if ever been reported in the ice cliff sections even where bedrock between ice and the sea can be safely examined at close quarters (figure 6). This point has been stressed by Dewar (1967).

It has been suggested by Koerner (1961, 1964) that the ice piedmonts in Trinity Peninsula rest on pre-glacial pavements because he considered that the regular linearity of the mountain headwalls behind the

Fig. 6. Ice cliff in Lowdwater Cove on the northern side of Palmer Station, showing clean, debris-free ice resting on bedrock.

ice piedmonts could not have been produced by a strandflat cutting glacier which had eroded its pavement for as great a distance as 6.2 miles (10 km) from the coast. From his observations on the Fuchs Ice Piedmont, Dewar (1967) regarded this point as even more significant because in places the mountain headwalls are 10 miles (16 km) from the coast. On Anvers Island the distance is even greater though the linearity of the mountain headwall is not so strongly defined.

The degree of activity of the piedmonts and other fringing glacial features appears to vary from one part of the peninsula to another, but the sea-calving type generally seems to be close to an equilibrium condition. This was Koerner's (1961, 1964) conclusion regarding the piedmont between Hope Bay and Trepassy Bay where the positive side of the budget is compensated for by calving at the ice front. Photographs taken by Bodman in 1903 and a sketch map by Duse (in Nordenskjöld & Andersson 1905) showed no significant difference in elevation and areal extent and suggested support for the equilibrium hypothesis.

The relatively inactive regime of the piedmont between Hope Bay and Trepassy Bay led Koerner (1961, 1964) to conclude that the piedmont is a relict feature and that "as such it is a survival from the most recent glacial recession" and emphasised the similarity with "Baffin-Type" glaciers which are postulated as survivals from before the post-glacial climatic optimum, since "it is difficult to reconcile their reappearance since this time in the light of present-day precipitation" (Baird 1952, p 9). Thomas (1963) and Sadler (1968) suggest a relict nature for the ice caps of Galindez and Skua Islands in the

Argentine Island group because their very low elevation should not support the ice cover at present in existence. A similar argument might be given for the Wauwermann Islands in Bismarck Strait (figure 6).

From these considerations it is apparent that, even though much work has been accomplished, basic glaciological information about the Antarctic Peninsula in general, and about its middle latitudes in particular, is limited and in some respects almost completely lacking. From the point of view of the so called "fringing glaciers" and the ice piedmonts in particular, four major questions arise and are the main subject matter of this thesis. These questions relate to the mass balance, state and activity of regime, the origin of the subglacial "platforms", the age and origin of the present ice cover and the degree of protection or denudation caused by the ice at present in existence.

Fig. 7. The low-elevation, permanently ice-covered Wauwermann Islands in Bismarck Strait. Possibly these are relics of a former more extensive ice cover. Note the "Roches Moutonnes" form suggesting east-west (l. to r.) ice movement.

CHAPTER TWO

NATURE OF THE INVESTIGATIONS

Introduction

In January 1965, Arthur Harbor, on the southwest coast of Anvers Island, was the construction site of "Palmer Station", a research installation of the United States Antarctic Research Program (USARP). Palmer Station served as the base of operations for glaciological investigations on the Marr Ice Piedmont by personnel from the Institute of Polar Studies, The Ohio State University.

The work in this thesis is the outcome of that fieldwork carried out while the author was Principal Investigator of glaciological and meteorological research at Palmer Station, between February 1965 and January 1967. It also draws on information obtained by Mr. R.A. Honkala, the author's successor, and his collegues, who continued some of the field measurements until January 1968. This work, sponsored by the National Science Foundation, Washington DC, formed a part of the overall United States Antarctic Research Program.

No previous work had been carried out on the Marr Ice Piedmont so the initial effort was a pilot project with no foundation of previous data on which to build. Thus, as a first consideration during the course of preparation for the fieldwork, a comprehensive investigation or "case study" was regarded as warranted. The program was, therefore, directed first toward obtaining basic field data and other supplementary information for a detailed study of the mass balance and regime of this (probably sub-polar) ice piedmont and to correlate these observations with present knowledge of glaciological conditions throughout the Antarctic Peninsula. Particular emphasis would be given to surface velocity determination, an aspect inadequately covered in many previous glaciological studies in the area. An ultimate goal was to apply and relate these data to current glaciological theory which might assist in establishing the position and significance of the ice piedmont within the wider spectrum of glaciological problems in the Antarctic Peninsula.

Initially, in the absence of first-hand knowledge, it was proposed to extend the study over the entire piedmont. The naiveté of this plan quickly became apparent. Foul weather, equipment and vehicle breakdown, together with time spent on logistic matters and the necessities of station operations, severely restricted the amount of scientific work that could be accomplished in the field.

Chronology of Field Activities

Field work was begun on February 3 1965 and except for two short interruptions, due to personnel change-over in January 1966 and January 1967, was continuous until January 1968.

Soon after activities began it was realized that the extremely inhospitable climate would prevent a detailed study of the entire piedmont which in area amounts to approximately 1,350 km^2. Consequently, the southwestern "lobe" was selected as a feasible study area, probably representative of the whole piedmont, and which was readily accessible by surface vehicle from Palmer Station. This area is bounded by the coastline from Cape Monaco to Cape Lancaster, the mountains from Cape Lancaster to the vicinity of Mt. Francais (actually to the unnamed "Big Mountain" in figure 1) and by the high divide from this mountain to Cape Monaco and amounts to approximately 380 km^2. This area appears to be a distinct drainage basin, with ice flowing outward into Wylie Bay, Bismarck Strait and the Neumayer Channel (figure 1).

Accumulation Stake Networks

During early 1965, two principal lines of bamboo poles were set to measure snow accumulation and ice ablation. In Figure 8, these are shown as the Main and Mountain lines and the Neumayer and Top-21 lines. The latter extended northward as the Top Line Trail. They were designed to obtain representative data from the low coastal areas to the higher interior parts. In addition, five stake farms, of 25 stakes each set 100 m. apart in five lines 100 m. apart were set to measure areal variability of snow accumulation. These are marked A, B, C, D and R in Figure 8.

Late in 1965, the number of poles was increased to over 600 as shown in Figure 9. Measurement of this network, with the exception of the stake farms, which were cannibalised for replacement stakes, was made several times a year until January 1968.

Ice Movement Stations

Fifteen ice movement stations were established during April and May 1965. Nine were in the peripheral ice around Arthur Harbor and six were further inland.

Beginning in late September 1965, the number of stations was increased to 68 and extended over the study area as shown in Figure 9.

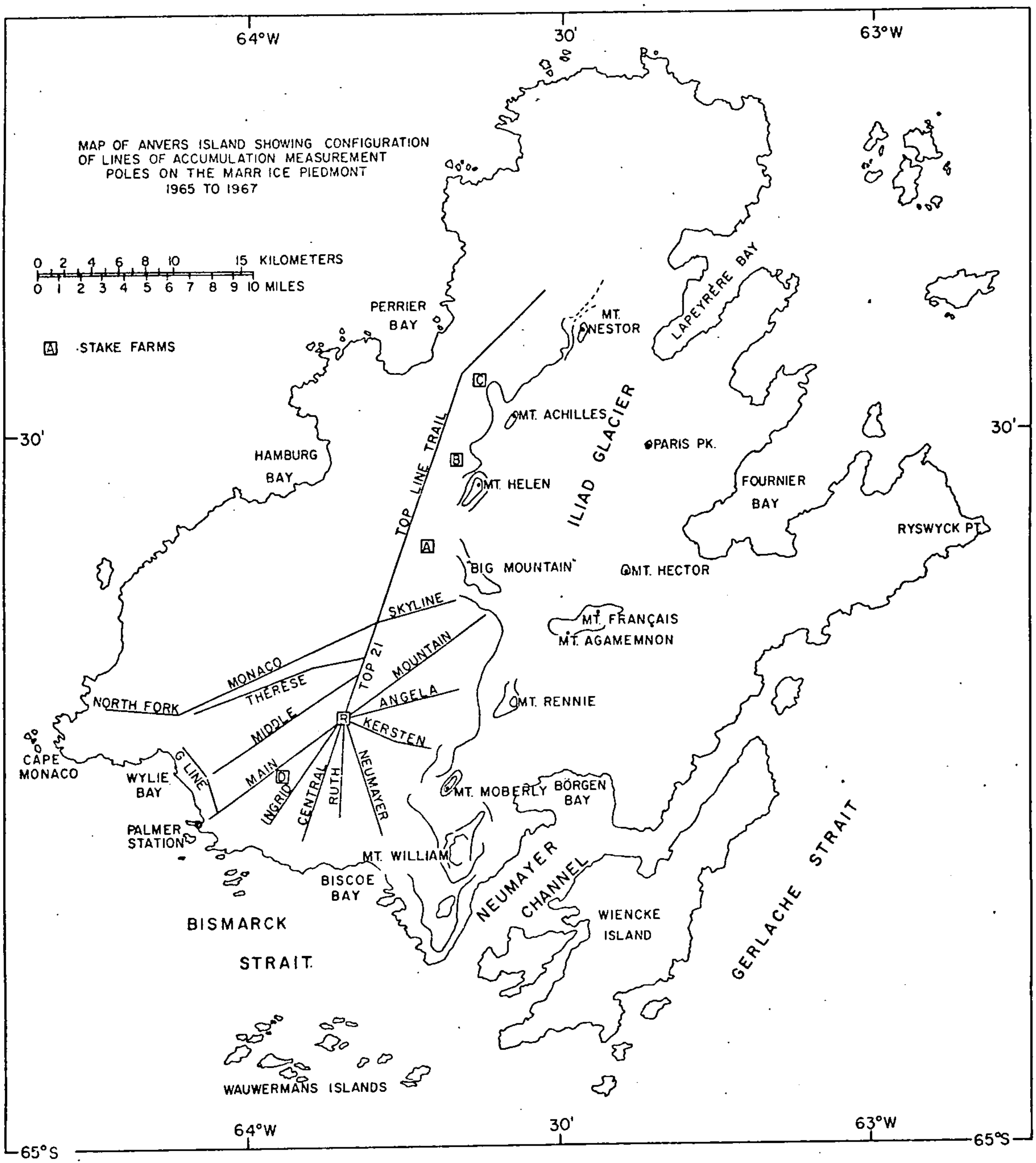

MAP OF ANVERS ISLAND SHOWING CONFIGURATION
OF LINES OF ACCUMULATION MEASUREMENT
POLES ON THE MARR ICE PIEDMONT
1965 TO 1967
0 2 4 6 8 10 15 KILOMETERS
0 1 2 3 4 5 6 7 8 9 10 MILES
STAKE FARMS
64°W
30'
63°W
65°S
PERRIER BAY
HAMBURG BAY
MT. NESTOR
LAPEYRÈRE BAY
MT. ACHILLES
ILIAD GLACIER
PARIS PK.
MT. HELEN
FOURNIER BAY
RYSWYCK PT.
TOP LINE TRAIL
BIG MOUNTAIN
MT. HECTOR
SKYLINE
MT. FRANÇAIS
MT. AGAMEMNON
MONACO
THERESE
TOP 21
MOUNTAIN
NORTH FORK
MIDDLE
ANGELA
MT. RENNIE
KERSTEN
CAPE MONACO
WYLIE BAY
G LINE
MAIN
INGRID
CENTRAL
RUTH
NEUMAYER
MT. MOBERLY
BORGEN BAY
PALMER STATION
MT. WILLIAM
NEUMAYER CHANNEL
GERLACHE STRAIT
BISCOE BAY
WIENCKE ISLAND
BISMARCK STRAIT
WAUWERMANS ISLANDS

Fig.8

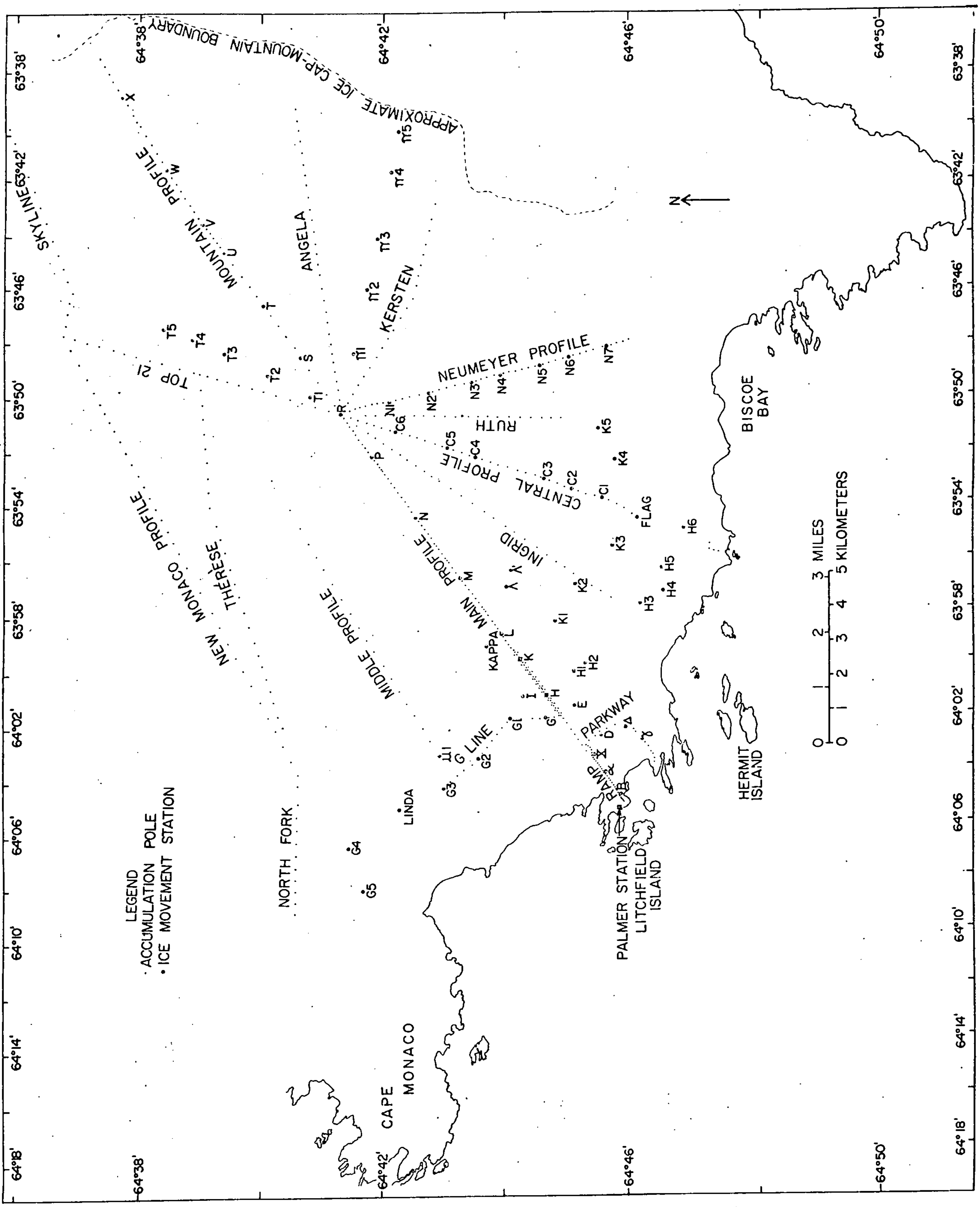


Fig. 9. Network of accumulation and ice movement stations.

Most of these stations were re-surveyed during late 1966 and early 1967 and a third survey was carried out by Honkala, late in 1967. Intermediate surveys were made at several of the stations in the peripheral area.

Associated Investigations

Snow pit studies to depths up to 3 meters were made throughout the period of field investigation to obtain data on snow density, accumulation, ice temperature and facies zonation. In addition, deeper investigations were made by coring to depths of 10 to 12 m.

On the steep ice ramp behind Palmer Station, a detailed study was attempted of superimposed ice formation and ice ablation. This was augmented during 1966 by investigations of the micro-climatic regime of the ramp and its correlation with the ramp's internal structure. Associated with this study was a small meteorological station at 300 m. elevation on the piedmont, which was originally established in April 1965.

Attempts were also made to produce a topographic map of the study area but proved abortive. Poor field working conditions, limited equipment and personnel and lack of available time, prevented this. However, the magnitude and direction of surface slope and the position of the contour was obtained at most of the ice movement stations.

A standard program of meteorological observation was established at Palmer Station on February 1 1965 and was continued throughout the entire period of investigation.

L.E. Brown engaged in geological investigations, with emphasis on the glacial aspects and G. Dewart, working on an independent National

Science Foundation Grant, conducted an investigation into ice thickness and bedrock configuration.

CHAPTER THREE

PHYSICAL CHARACTERISTICS OF THE STUDY AREA

Surface Configuration and Elevation

The surface of the piedmont has not been adequately surveyed and mapped and the only aerial photographs were taken by the Falkland Islands Dependencies Survey (now British Antarctic Survey) in 1956 and by United States Navy and Coast Guard aviators, in support of USARP, between 1963 and 1967. Two map sheets showing the southwestern coastline and peripheral crevasses have been produced by the British Antarctic Survey (1964) but no detailed topographic maps of the interior are available.

Spot elevations, accurate to an estimated ± 3 meters, are however available from the ice movement survey. The surface profile from the base of the Norsel Point ramp behind Palmer Station, to a distance inland of about 5 km, was obtained by levelling during the winter of 1966. In addition, rough elevations were obtained by altimetry during Dewart's gravity and magnetic survey in January 1967. An attempt to produce a plane table map, by L.E. Brown during 1966, proved to be too large a task.

Ice Thickness and Bedrock Configuration

There have been no seismic investigations on any part of the Marr Ice Piedmont and attempts to measure ice thickness using radio-echo sounding equipment, by Dr. Charles Swithinbank of the British Antarctic Survey, during January 1967, were largely unproductive. The performance of the radio-echo sounder in "warm" ice is poor and the main cause of the failure of the equipment on Anvers Island is attributed to the relatively high englacial temperature. However, some ice thickness information was obtained from poor but readable film on one short flight traverse (figure 10).

Ice thickness over the rest of the study area was determined by G. Dewart, using gravimetric techniques (Dewart 1971). These thickness values are based upon Bouguer ice anomalies, that is, the gravity anomalies due to assumed density differences between average rock and ice. The author is grateful for the following information on the project.

The surface of the piedmont was taken as essentially flat and its depth small in relation to its lateral extent. The assumption was made that at each observation point (for the most part, observations were made at the ice movement stations) the ice consisted of an infinite slab resting on an infinite horizontal rock surface. The difference in the Bourguer gravity anomaly between a point on the ice and a base station on bedrock (Palmer Station), after certain corrections, was assumed to be due to the presence of the ice and rock layers under the observation point. With values assumed for the ice and rock, the height of the rock surface with respect to the elevation of the base station and the thickness of the ice could be found:

$h = g/0.04185(r - i)$, where h = ice thickness, g is the difference in Bourguer anomalies and r and i are the densities of rock and ice respectively.

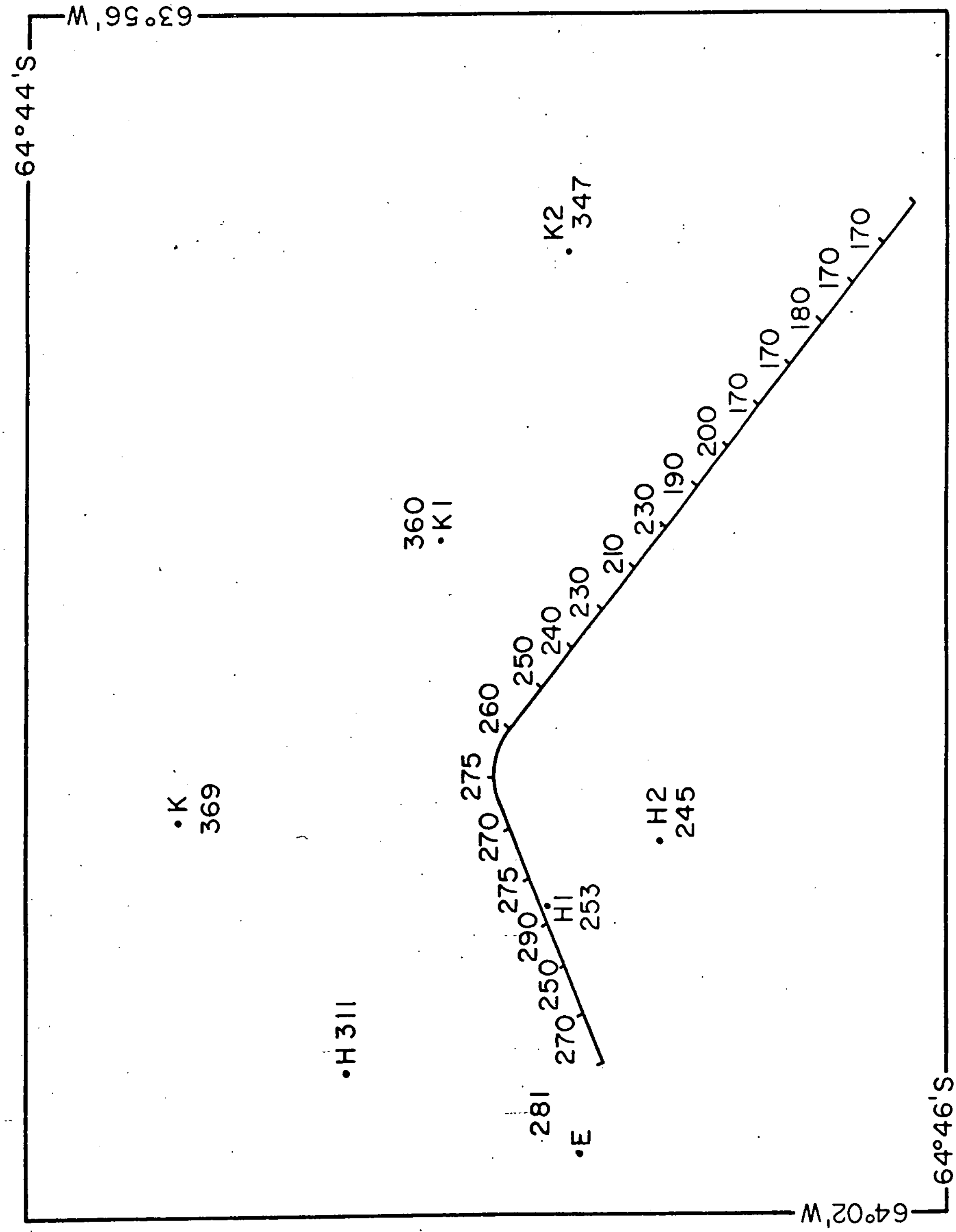


Fig. 10. Ice thickness values (meters) from radio-echo sounding flight traverse, January 1967. (From an original, courtesy of Dr. C.W.M. Swithinbank).

Station elevations were measured by altimeter. Pressure variation control was provided by the barograph records at Palmer Station. Pressure gradient effects were probably small due to the fact that the most distant gravity station was only 25 km. from Palmer Station. Dewart estimated altitudes as accurate to $\pm$ 10 meters

The regional gravity anomaly was estimated from the data in figure 11. Unfortunately, the only bedrock stations were those along the coast. In the interior of the piedmont, especially near the massif, the regional gradient may change significantly.

The topographic effect of the ice surface was regarded by Dewart to be negligible but an estimated correction (using the Hammer zone method) was made on the relatively steep slopes near the coast. The subglacial terrain effect was also estimated where there appeared to be significant bedrock features. The topographic effect of the mountains could also be only roughly estimated in the absence of a detailed topographic map.

The density of the ice was assumed to be 0.9 g/cm^3. The mean density of forty specimens of rock collected along the southwest coast was 2.67 $\pm$ 0.12 g/cm^3. Samples from the altered assemblage and from the Andean Intrusive Suite (see particularly, Hooper 1962) had densities near this mean. These rocks apparently underlie much of the section of the piedmont under investigation, so a value of 2.67 g/cm^3 was assumed for the rock density. From the standard deviation Dewart expected an error of 5 percent due to density variation. Furthermore, some of the area may be underlain by Cape Monaco Granite, the few

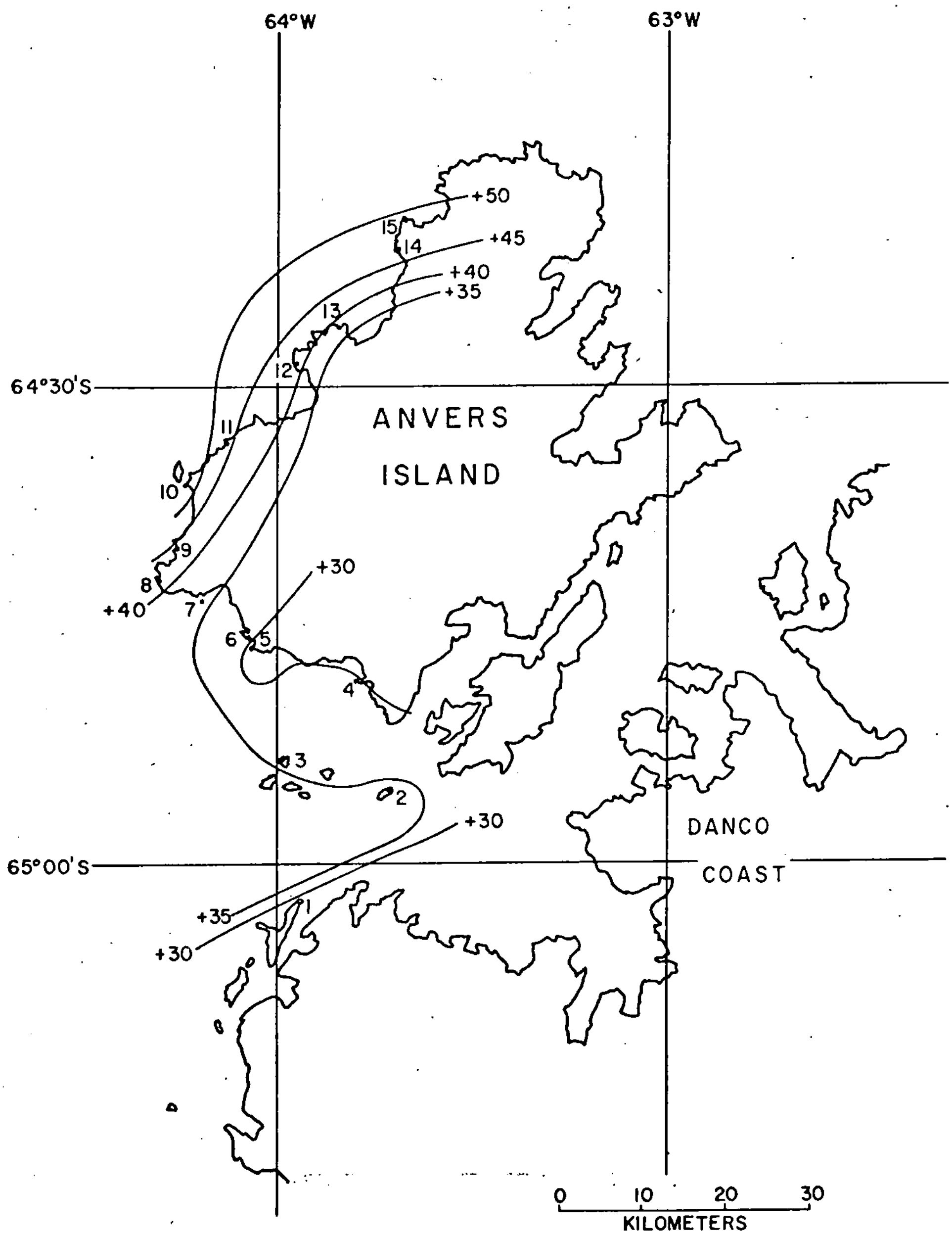


Fig. 11. Bouguer anamalies in the vicinity of Anvers Island and Bismarck Strait. Contour interval is 5 mgal. Numbered points refer to regional-survey gravimeter stations. (After Dewart 1971).

specimens of which were available having densities in the range 2.4 to 2.6 g/cm^3.

As the gravimeter drift became more significant during the work on the piedmont, it was necessary to frequently re-occupy stations to make corrections. For the main period of the field investigation, January 30 to February 22 1967, the mean drift rate was 0.02 mgal/hr. and double this rate was experienced for periods of one day or less.

Overall, Dewart estimated that the ice thickness values he reported could be in error by as much as ± 20 percent, though he did concede that the accuracy in the flat region between the coastal slope and the proximity of the mountains was probably better than this. Dewart's assessment of accuracy could however, be slightly pessimistic. Swithinbank's radio-echo analysis obtained a series of ice depths in the coastal region around stations H1 to H6, shown in Figure 10. His assessment of accuracy (personal communication) was ± 24 meters in ice averaging about 250 m. thick, or about 10 percent. Dewart's results are quite similar; a difference of 30 m. in ice almost 300 m. thick. Unless the echo-sounder was functioning very badly, it would seem that the gravimetric error could be appreciably below 20 percent. The gravimetrically derived thicknesses do therefore, appear to be reliable.

The estimated ice thickness over the study area is shown in Table I and in Figure 12. Figure 13 is a map of the underlying topography.

From these data the subglacial surface of the study area appears to consist largely of a narrow coastal plain and a low inland plateau. The overall slope is gentle, being 2 to 3 percent, but the slope incr-

TABLE I

VALUES OF ICE THICKNESS (h)

Station	h (m)	Station	h (m)
Alpha	150	Pi1	471
Big X	174	Pi2	453
D	187	Pi3	536
Delta	153	Pi4	423
		Pi5	177
E	255	N1	452
H	312	N2	425
K	348	N3	382
L	364	N4	366
M	348	N5	401
N	392	N6	322
P	448	N7	306
R	496	0.5 Km beyond	292
S	522		
T	550	K1	317
U	581	K2	332
V	589	K3	356
W	557	K4	402
Little x	510	K5	440
1.0 km beyond	303	Flag	358
		H1	260
C1	361	H2	245
C2	306	H3	205
C3	372	H4	364
C4	416	H5	-
C5	447	H6	378
C6	461		
		G	274
T1	506	G1	287
T2	535	G2	380
T3	558	G3	339
T4	598		
T5	612	Mu1	404
TS**	567		

TS** is pole #21 on Top Line

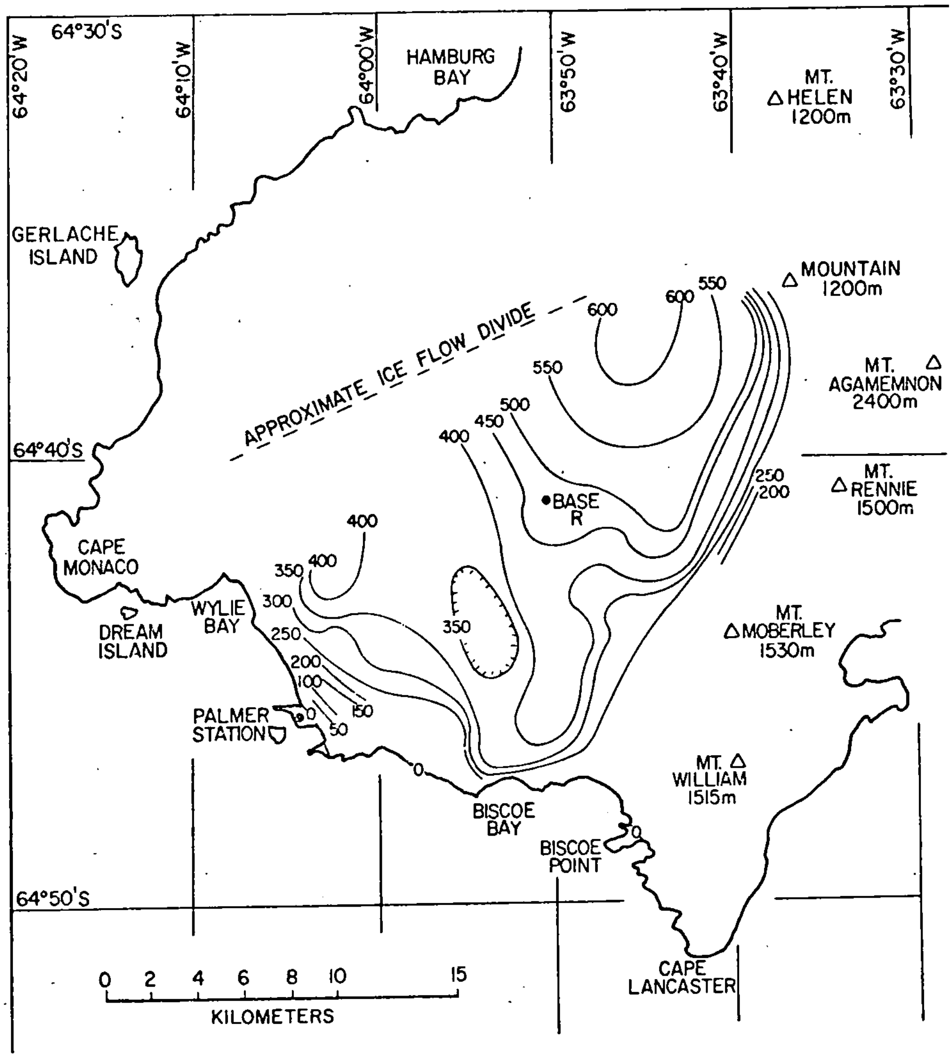


Fig. 12. Ice thickness within the study area. Contour interval is 50 meters. (After Dewart 1971)

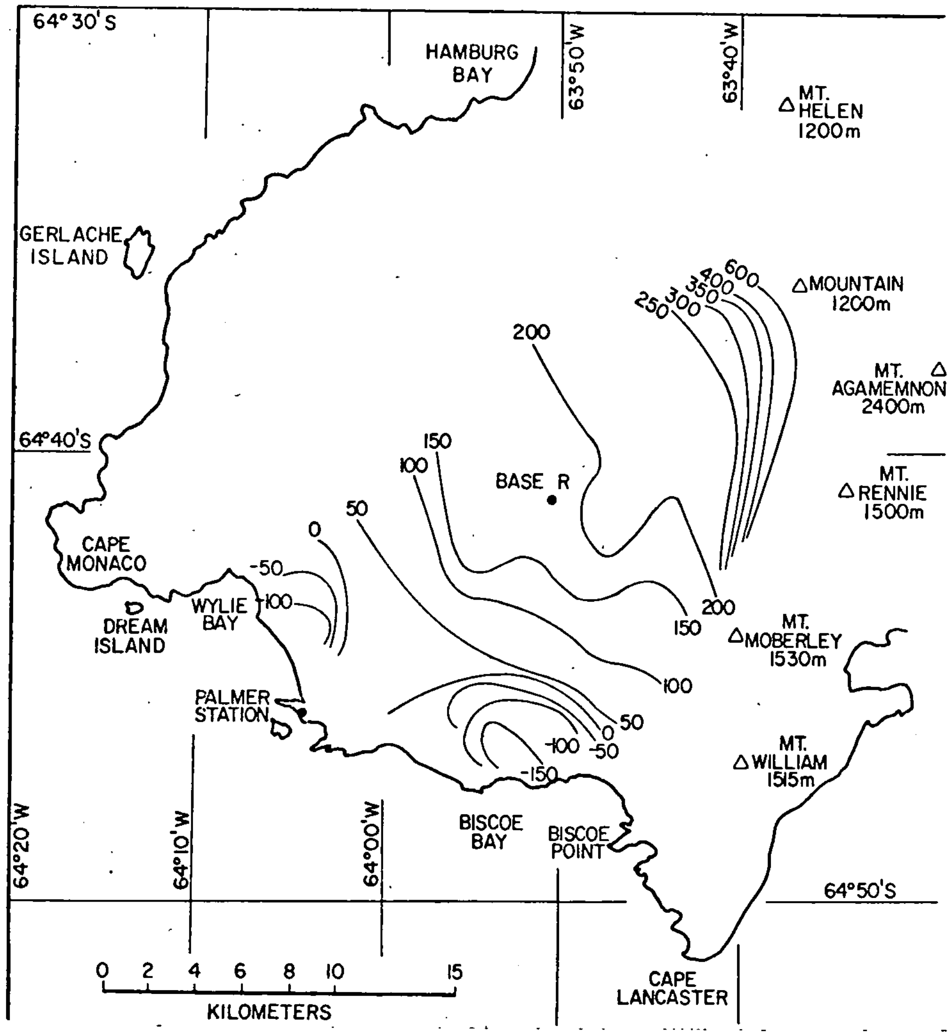


Fig. 13. Bedrock elevation in meters within the study area. (After Dewart 1971)

eases on the approach to the plateau between 4 and 10 km from the coast and again at the foot of the mountains. The coastal plain is indented by two hollows which represent the heads of Wylie Bay and Biscoe Bay. In these bays the bedrock drops below sea level about 3 to 4 km in from the ice front which marks the apparent coastline. However, the ice front appears to be grounded and along the rest of the southwest coast, bedrock can be seen at the foot of the ice front. These bays are separated by a low saddle which reaches the sea in the rocky promontories of Norsel Point and Bonapart Point near Palmer Station. Bedrock is visible below the ice front in the northern part of Arthur Harbor but not in the southern part but it seems probable that inland of the ice front, much of the subglacial surface inland for about 1.0 to 1.5 km, is below sea level forming a smaller but similar hollow to that of Wylie Bay and Biscoe Bay.

The bedrock terrain has an ice cover which is largely between 300 and 600 m. in thickness. Greatest thickness is reached just below the foot of the mountains, then thins abruptly as the elevation rises. Toward the coast the ice thickness decreases over the plateau, then remains almost constant and even increases slightly on the slope down to the coastal plain. On the coastal plain the ice thickness decreases rapidly toward to the coast.

The features described above are more clearly illustrated in the cross sections shown in Figure 14. The position of the ice movement stations along the profiles are also shown.

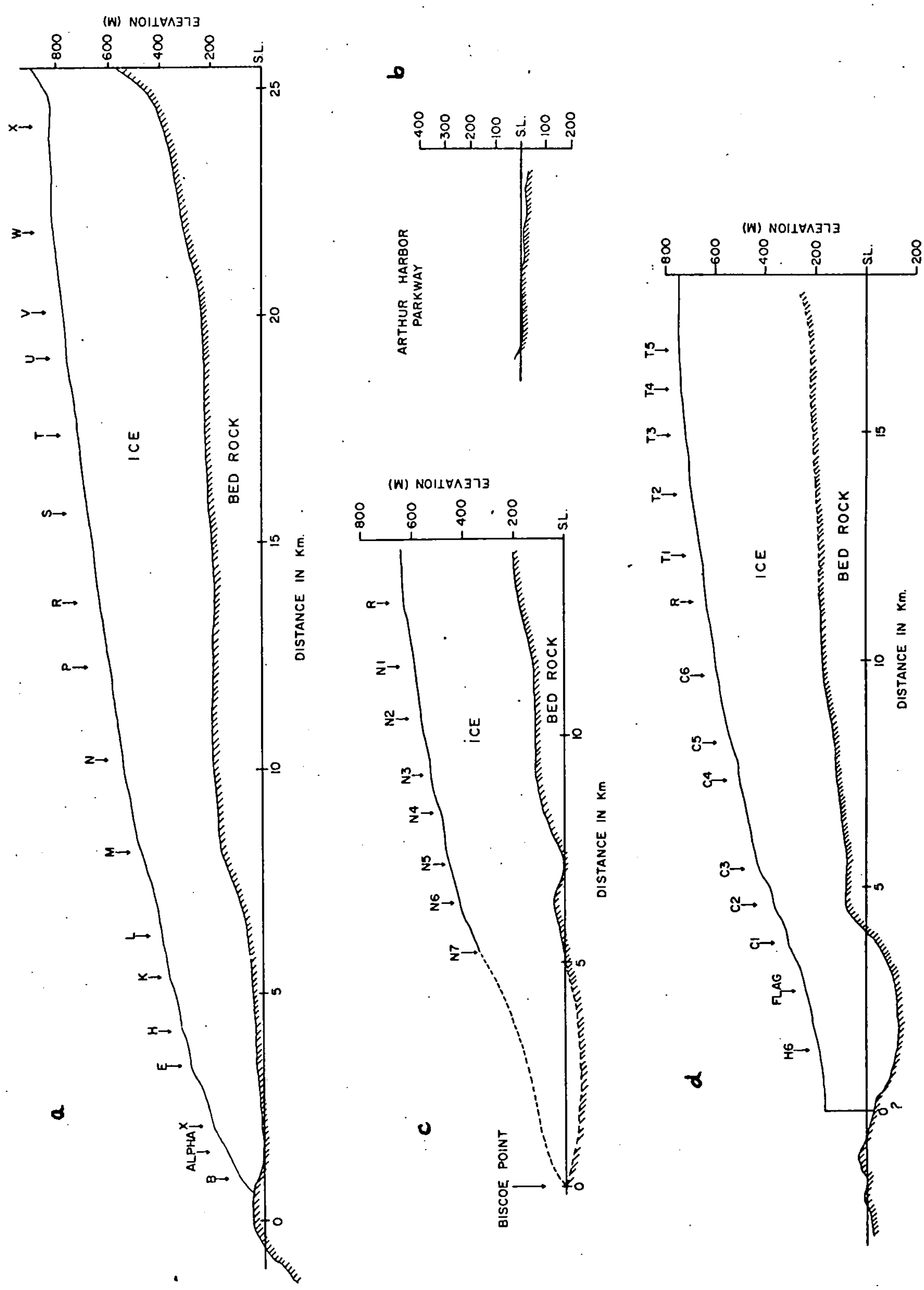


Fig. 14. Longitudinal profiles of bedrock and ice surface.

Along the Main profile (figure 14a) there is no decisive topographic trend until movement station L is reached. Here there is an abrupt change of slope and the subglacial surface rises about 150 m. between points L and N. Beyond point N the surface levels off, forming the plateau which here has an elevation of about 200 m. above sea level. The plateau then slopes gradually upwards to the northeast, but at a lesser rate than the ice surface, which in this region reaches to 700 to 800 m. above sea level. Consequently, ice thickness gradually increases toward the foot of the mountains. In the vicinity of station V, another break of slope occurs and it is clear that in this region the Achaean Range begins to emerge from the plateau. The greatest thickness of ice, approximately 600 m, probably occurs in the vicinity of Station V.

Figure 14b is a "transverse" section plotted from gravity readings taken at each of the accumulation poles on the "Parkway" (see figure 9) around Arthur Harbor. Considering the possible error in the ice thickness values, the bedrock surface appears to be at or very close to sea level below this transect. The Parkway is situated on the periphery of the crevasse field of Arthur Harbor and the shape and lineation of the crevasses together with a sharp break of slope at the surface, suggests to this author that the ice is slumping into the harbor and that the head of the harbor is in the form of a hollow, smaller but similar to that at the head of Wylie Bay. The southern section of this profile appears to be distinctly lower than the northern part and represents positions on the flanks of Bonapart Point. On the surface in this area there is a distinct depression which may reflect a small valley on this side of the harbor head (figure 15). Unlike the northern side of the harbor at the ice front, bedrock is not visible below the ice cliff in this southern section.

Fig. 15. Surface depression in the ice behind Arthur Harbor

On the Neumayer line (figure 14c) the plateau also appears rather abruptly, here between stations N3 and N5. The exact configuration of the coastal plain in this section is not known because the heavily crevassed surface prevented access for gravity readings. However, in the area below station N7, the surface falls away rapidly, causing heavy crevassing, to form a well defined depression. On the basis of these considerations it seems probable that bedrock in this area is slightly below sea level but rises again to form the Biscoe Point promontory.

Access in the coastal region in the vicinity of stations H6 and Flag was good and a distinct bottom profile was obtained (figure 14d). Here a very pronounced subglacial basin occurs and suggests a former glacial overdeepening action. Whether or not the ice front is grounded at the end of this profile could not be precisely determined as access was prevented by the calving cliffs. However, bedrock does rise to form small rock outcrops immediately in front of the ice cliff. The ice front itself therefore, must rest on bedrock at or very close to sea level. The emergence of the inland plateau is more difficult to recognise in this profile. It seems more likely, considering the place of emergence in other areas, that the plateau begins very abruptly in the vicinity of stations C1 and C2, in which case there is no real coastal plain here. However, the break of slope in the area of station C5 may indicate the beginning of the plateau. In this case the coastal plain has been heavily eroded below stations Flag and H6 and the transition from the plain to the plateau is very gradual. The gravity survey did not extend into the T profile but on the basis of surface observations, bedrock can reasonably be expected to be as indicated in Figure 14d.

To the north the surface above the plateau culminates in a divide more or less coincident with the Skyline profile (see figure 9). Beyond this point the surface falls away gradually. As the gravity survey did not extend into this area, it can only be speculated that the bedrock also falls away slightly.

Figure 16 shows transverse sections moving progressively inland. Though speculative in some parts they illustrate the low-lying character of the subglacial surface and the general form of the valley indentations. The bedrock profile below station Linda however, does not strongly reflect the dominant ice stream in which that station was placed (figure 17). The gravity survey in that area was weak and a more pronounced "valley" might be detected through a more detailed gravity survey.

Discussion

The results of the bedrock survey do not describe a "strandflat" according to the classical definition, which, from Embleton and King's (1968) summary, is essentially a horizontal cut into the land mass and which encompasses only low elevations up to about 40 m. Holtedahl (1960) describes the strandflat as often being made up of low islands and skerries which often continue below sea level before falling abruptly to deeper water. Maximum height is usually 40 to 50 meters only and inland, the strandflat ends sharply against cliffs

To the contrary, the subglacial surface of Anvers Island emerges as a relatively massive platform, generally inclined seaward with a general elevation of 200 meters a.s.l. but reaching to over 400 m. at its inland terminus. If any part conforms to the classical description of the strandflat, it is the low coastal plain, seaward of the

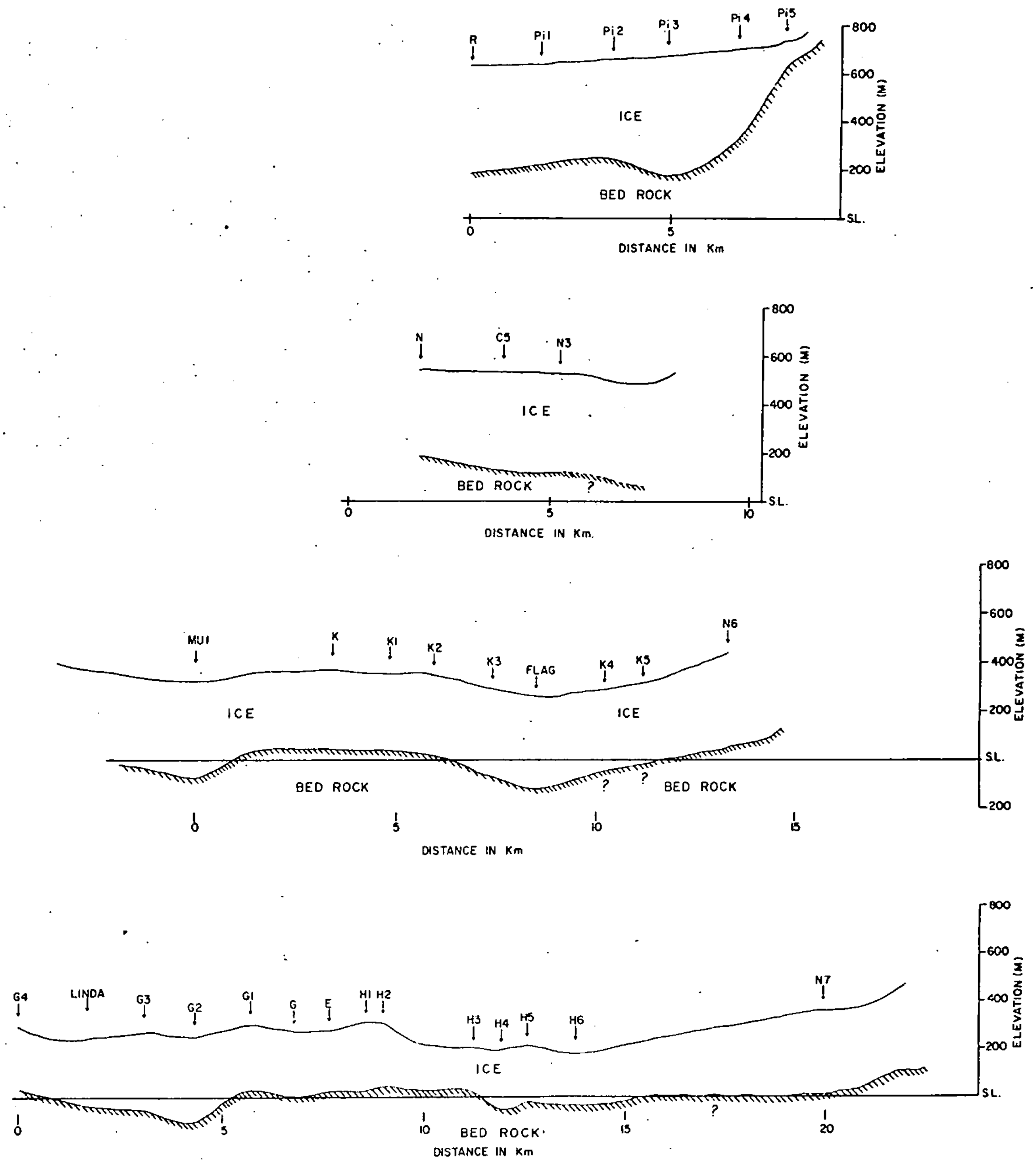


Fig. 16. Transverse profiles of bedrock and ice surface.

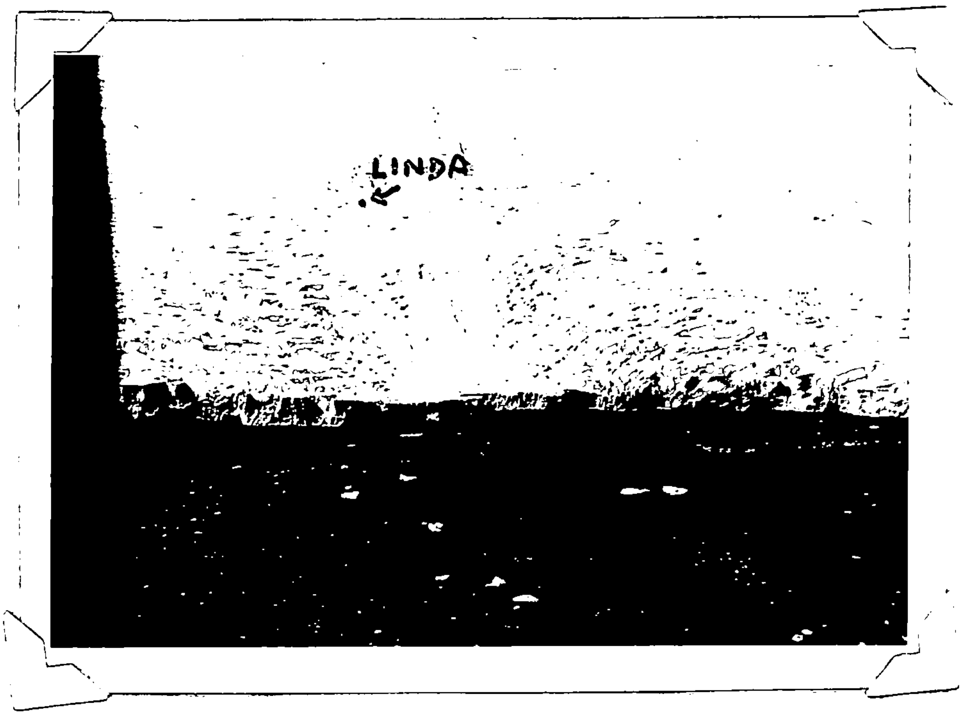


Fig. 17. View looking east of the ice stream in which movement station LINDA was placed.

50 m. contour, which continues out to sea, like a skjaergard, to include Litchfield Island and the small islands of Arthur Harbor (see figure 2), the rock promontories extending outward from the ice ramps and the scattered skerry islands and shallowly-submerged rocks offshore (figure 18).

Recent Changes in Surface Level and Dimensions

In the absence of detailed topographic maps, it is not possible to compare the present distribution of surface elevations with previous surface levels. Absolute elevations on the piedmont were determined at most velocity stations on at least two occasions and the configuration of surface slope was also measured. If the elevation of the ice surface is stationary in space for long periods, that is, in an equilibrium condition, the elevation of a velocity station at a second determination (E_2) should be:

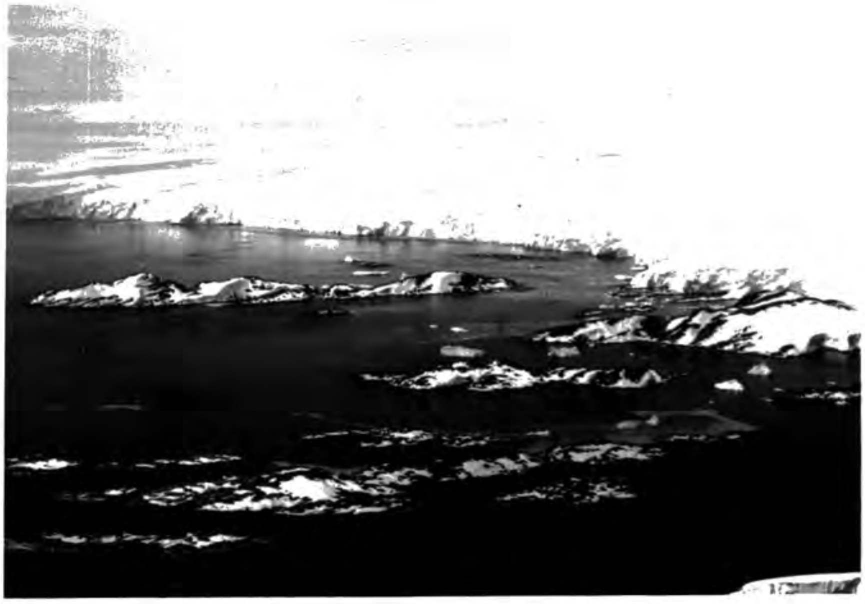
$$E_2 = E_1 - d \sin \alpha$$

where d is the distance moved down slope and α is the angle of the surface slope. E_1 is the first determination of absolute surface elevation. If the absolute elevation of the surface is increasing by a factor equal to the accumulation then:

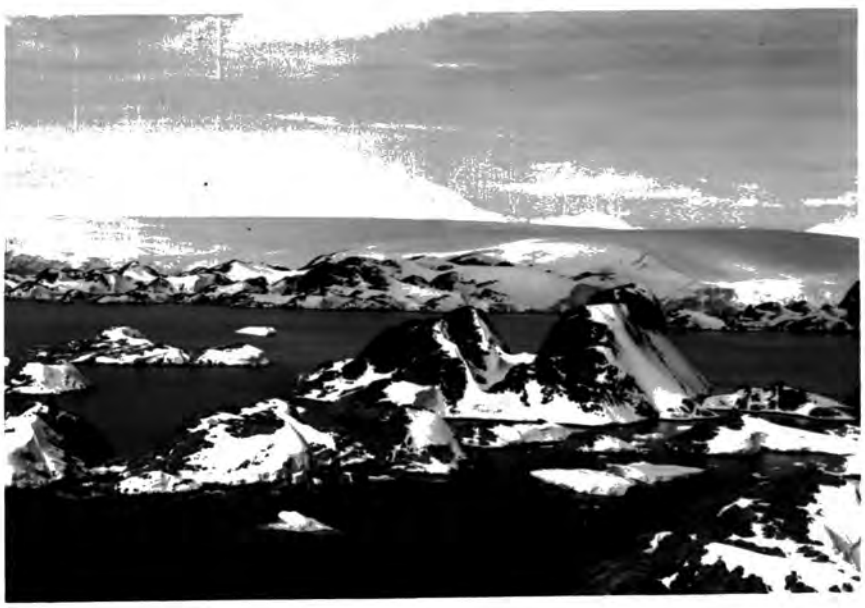
$$E_2 = E_1 - d \sin\alpha + \text{Accumulation}$$

Evaluation of these parameters from the piedmont data gives inconclusive results and it is assumed that surface levels are approximately stationary with regard to a fixed coordinate system in space.

Comparison of the British Antarctic Survey aerial photographs



a) Area around Biscoe Point



b) Offshore of Cape Monaco.

Fig. 18. Skaerjard-type topography offshore of Anvers Island.

taken in 1956 and subsequent photographs taken by US military personnel and the glaciology working group show virtually no change in the shape or position of the terminal ice cliffs and indicates that these are stable features. Thus the dimensions of the piedmont seem unchanged over the past ten years at least. However, the ice boundary is probably governed by sea level. It is uncertain as to how much of the peripheral ice is afloat. The British maps show grounded ice at a few scattered locations but the oblique aerial photographs taken by the US Navy show bedrock exposures along most of the coastline. Dewart, during his gravity survey, made similar observations (personal communication). It seems that the heavily crevassed peripheral ice is not sufficiently cohesive to float like an ice shelf or ice tongue (any consideration of ice temperature notwithstanding) and consequently, calves on reaching the sea. Any advance of the ice front would therefore, call for a lowering of sea level and conversely, a recession would result from a rise in sea level. The overall dimensions of the piedmont, both horizontally and vertically, seem therefore, to have been stable for a short time at least.

Similar conclusions can be drawn from the geological evidence. The physiographic features of Litchfield Island, less than one kilometer from the present ice front, bear evidence of a former ice cover but there is no evidence of very recent glaciation. Subaerial erosion has advanced to the stage of talus and scree formation which is already fixed by a luxuriant vegetation of mosses, lichens and scattered tufts of the grass _Deschampsia antarctica_. A prolific insect fauna is associated with this. Nor do the smaller, Humble and Torgersen Islands in Arthur Harbor, bear evidence of a recent ice cover.

On the Bonapart Point promontory, extending outward from the present ice edge for 10-20 meters, there is evidence of a very recent glacial recession, but the area is very small. Striated boulders and chatter marks are abundant and fresh rock flour is visible on many of the larger boulders. The localisation of this retreat is remarkable and suggests a local meteorological phenomenon as its possible cause. No such evidence exists at the edge of the Norsel Point ramp.

On both promontories, extending outward for 100 to 150 meters, is old evidence of glaciation. Most of this is faint and though the degree of erosion has not reached that on Litchfield Island, plant colonisation is well advanced. K.R. Everett, in a personal communication, has suggested that these rock surfaces have been exposed for at least 100 years and possibly for 150 years. There is no evidence of a former ice cover beyond about 150 meters from the present ice edge. This paucity of evidence was remarked upon by Hooper (1962) who came to similar conclusions.

Glacial History

From these observations, little can be said about the glacial history of the Marr Ice Piedmont. The evidence in this area is not as widespread as in the vicinity of Adelaide Island where the evidence reaches six to eleven miles from the present ice front (Dewar 1967). What little evidence there is, is difficult to correlate with that from other parts of the Antarctic Peninsula because the relatively small amount of information currently available, for example, Nichols (1960), his summary of Antarctic glacial geology (1964), Adie's (1964a; 1964b) summaries of the work of the British Antarctic Survey and Hobbs (1968) observations, together with the more recent work of Clapperton (1971), Everett (1971) and John & Sugden (1971), deals more so with sea level changes as recorded by raised beaches than with chronology. From field observations on Livingston Island, South Shetland Islands, Everett (1971) has suggested that at least three glacial events are recorded and John & Sugden (1971) have presented a tentative glacial chronology with three phases and Clapperton (1971) recognized four glacial stages in South Georgia.

However, none of this evidence is helpful in formulating any kind of glacial history for the Marr Ice Piedmont. It must simply be concluded that within the general framework of present knowledge on the glacial history of the Antarctic Peninsula, there is little reason to doubt that the Marr Ice Piedmont could have been thicker and more extensive in the past. The evidence however, as is the case throughout much of the peninsula, is now below sea level.

The obvious conclusions to be drawn at this point therefore, are that investigations relating to the glacial history of the Marr Ice Piedmont should form a major part of future work in this area and that the glacial history of the Antarctic Peninsula as a whole is far from being adequately understood.

Climate and Meteorology

The meteorological facility at Palmer Station was operated in conjunction with the glaciological program and formed an integral part of the overall scientific investigation. However, though it is recognized that the relationship between glaciers and climate is intimate, the present investigation was not designed as a glacio-meteorological study. The meteorological record itself is of insufficient duration to warrant any detailed analysis and discussion and it is inappropriate and beyond the scope of the main text of this thesis to do so. Therefore, the salient features of meteorological conditions at Palmer Station from 1965 to 1967, together with the data obtained, are described separately in Appendix I, Volume 2.

Pertinent meteorological parameters and their implications relative to the main study are referred to in the discussions to follow, and the main text of this thesis will restrict itself simply to a detailed discussion of precipitation over the Marr Ice Piedmont.

CHAPTER FOUR

SNOW ACCUMULATION

INTRODUCTION

After an extensive search of the literature, it appears that the first measurements of accumulation on the Marr Ice Piedmont are those reported on in this thesis. The program of measurement was begun on February 3 1965 with the establishment of the Main-Mountain and Neumayer-Top 21 lines (figure 19). As pointed out on page 21 these are regarded as "longitudinal" profiles aimed at obtaining information relative to elevation and distance from the coast. The expansion of the stake network was begun in November 1965 and was completed by the end of the year. All poles were visited 3 to 4 times each year until January 1968 when the program was terminated.

Generally, accumulation is of a high order and is comparable to that of nearby areas (Bryan 1965; Sadler 1968). The high rate of accumulation caused maintenance of some of the lines to be extremely difficult and occasionally poles were buried before they could be visited. The most inaccessible from this aspect were the Monaco and Therese lines, due to the high incidence of fog and low cloud in this area. Another hindrance and source of loss over most of the piedmont was the massive buildup of rime and ice on the poles which, coupled with high wind, caused extensive breakages. Even though such losses were replaced there are several poles for which an uninterupted record is not available for any one complete year.

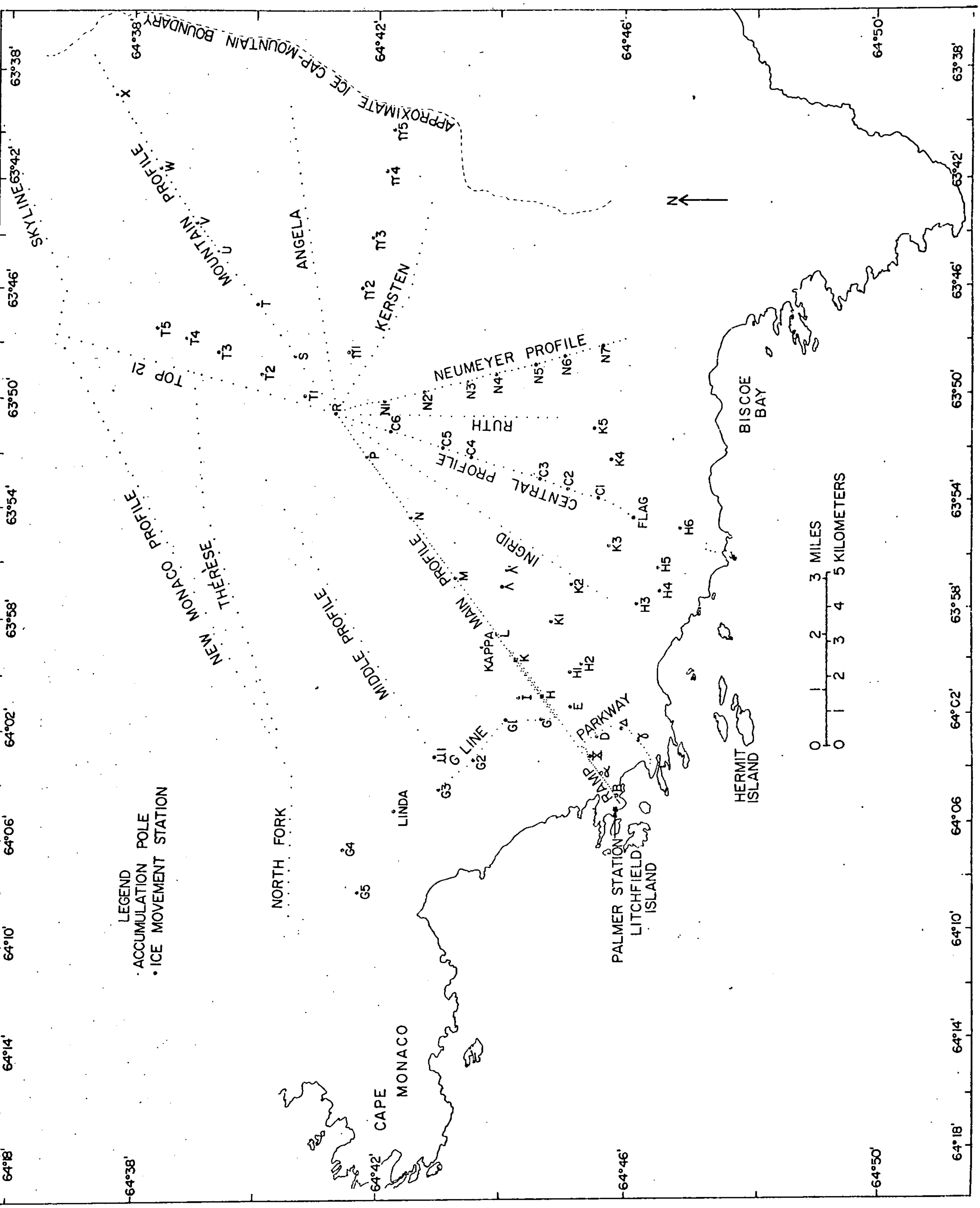


Fig. 19. Network of accumulation and ice movement stations.

Identification of individual poles was sometimes difficult even though distance between them was recorded and the poles were set at varying heights. Occasionally accumulation values were recorded which certainly seemed abnormal. Sinking of the poles is not regarded as the cause of these anamolous readings because resetting was always difficult and often impossible due to their freezing into the firn. Improper identification is the most probable cause. In such cases the records have been rejected and do not enter into the survey of accumulation.

SNOW DENSITY

The densities used to convert snow accumulation to water equivalent values have been derived from snow pit measurements and core samples taken at each of the 14 ice movement stations along the Main and Mountain profiles during the winter of 1965; (these studies are discussed in Chapter 9 below). The density values represent the average density from the surface to a depth equivalent to the average accumulation as was ultimately recorded by the stakes for the period 1965-1968. Thus the average density of the snow cover from about February 1965 and about 1.0 to 1.5 meters of the previous year's accumulation has been used.

In the highest parts of the accumulation zone, between about 650 m. and 850 m. elevation, the average density for the year's accumulation (assumed to be about August to August) was 0.453 g/cm^3. From about 450 m. to 650 m. elevation the density was 0.445 g/cm^3. At lower elevations the density was 0.429 g/cm^3 (300 m. to 450 m.), 0.518 g/cm^3 (60 m. to 300 m.), and 0.850 g/cm^3 (20 m. to 60 m.).

The variation in these values probably results from the degree of percolation and saturation of the firn cover and the amount of compaction resulting from the overburden of accumulated snow. Saturation above about 300 m. elevation is not severe but annual accumulation of snow increases markedly with increase in elevation, ranging from about 170 cm at 300 m, 230 cm at 450 m, 450 cm at 650 m and over 550 cm at 850 m elevation. The slight increase with elevation of the average densities above 300 m is probably due to the increase of the overburden. Below 300 m elevation the degree of percolation and saturation increases and is probably the cause of the increase in the average density. The value of 0.85 g/cm^3 between 20 and 60 meters is used in the superimposed ice accumulation and ice ablation zones.

MAIN LINE: STATIONS E TO R

Detailed data are available from the Main line for almost three complete years from 102 stakes set at intervals of approximately 100 meters. This was a heavily travelled trail and was well maintained throughout. The results have been broken down into periods of more or less one year each and are shown as g/cm^2 in Table II and in Figure 20. Despite irregularities in the lower part of the profile, a distinct increase of accumulation with elevation is evident.

The marked irregularity of accumulation in the lower parts of the profile is possibly related to local surface topography. Each of the

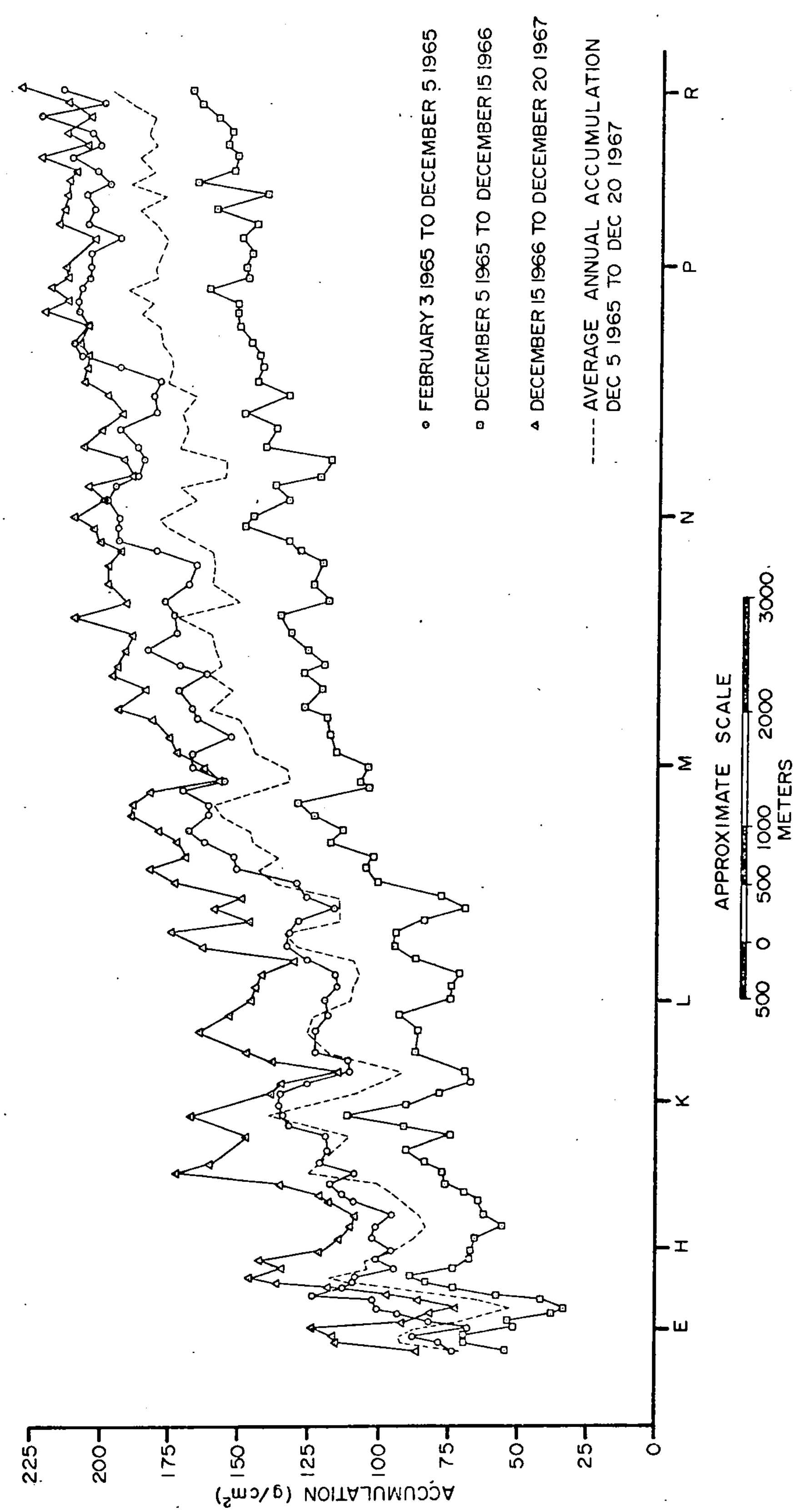


Fig. 20. Total and average annual accumulation (g/cm^2)
Main line: stations E to R.

TABLE II

TOTAL AND AVERAGE ANNUAL ACCUMULATION
(gm/cm^2) MAIN LINE

(1) February 3 1965 to December 5 1965)
(2) December 5 1965 to December 15 1966) (4) Average*) (5) Average
(3) December 15 1966 to December 20 1967))

Stake	(1)	(2)	(3)	(4)	(5)
1	73.4	54.5	86.2	70.3	71.4
2	78.1	69.1	115.8	92.4	87.7
3	88.8	69.9	116.7	93.3	91.8
4 (E)	68.2	51.9	124.0	87.9	81.5
5	82.4	53.6	91.8	72.7	75.9
6	93.1	37.8	81.8	59.8	70.9
7	100.8	33.9	72.1	53.0	68.9
8	102.5	41.6	85.4	63.5	76.5
9	123.6	57.1	97.0	77.5	92.6
10	112.8	73.4	118.4	95.9	101.5
11	109.4	83.2	136.9	110.0	109.8
12	108.5	89.2	146.8	118.0	114.8
13	94.8	73.8	134.3	104.0	101.0
14	101.7	67.8	142.9	105.3	104.1
15 (H)	95.2	67.4	121.8	94.6	94.8
16	103.8	60.9	114.1	87.5	92.9
17	102.1	55.3	110.7	83.0	89.4
18	95.2	62.2	108.5	85.3	88.6
19	109.4	64.8	118.0	91.4	97.4
20	113.7	69.5	121.8	95.6	101.7
21	117.5	76.4	136.0	101.2	110.0
22	109.0	77.6	172.9	125.2	119.8
23	121.8	84.0	160.9	122.4	122.2
24	118.4	90.9	—	—	—
25	119.3	74.2	147.6	110.9	113.7
26	132.6	91.4	—	—	—
27	134.3	112.0	167.7	139.8	138.0
28 (K)	136.4	90.1	—	—	—
29	130.8	78.1	139.0	108.5	116.0
30	125.3	67.4	135.1	101.2	109.3
31	110.3	69.5	114.5	92.0	98.1
32	111.5	—	138.1	—	—
33	123.5	87.1	148.0	117.5	119.5
34	123.1	86.2	164.3	125.2	124.5

Continues

TABLE II (cont)

Stake		(1)	(2)	(3)	(4)	(5)
35		118.8	93.5	153.6	123.5	122.0
36	(L)	119.7	74.6	146.3	110.4	113.5
37		115.4	74.2	114.6	109.4	111.4
38		116.2	71.6	142.8	107.2	110.2
39		126.1	87.5	131.0	109.2	114.9
40		133.8	95.0	163.0	129.0	130.3
41		132.6	94.8	174.6	134.7	134.0
42		129.5	84.1	145.4	114.7	119.7
43		116.7	69.5	159.6	114.5	115.3
44		126.1	78.1	149.7	114.9	118.0
45		130.0	101.2	173.3	137.2	134.8
46		151.4	105.5	182.3	143.9	146.4
47		152.3	103.0	169.4	136.2	141.6
48		163.0	118.8	172.9	145.8	151.6
49		168.6	113.2	179.3	146.2	153.7
50		161.3	124.0	189.6	156.8	158.3
51		161.3	130.0	188.8	159.4	160.0
52		170.7	104.7	182.7	143.7	152.7
53	(M)	156.1	107.7	157.0	132.3	140.3
54		167.3	104.7	163.0	133.8	145.0
55		167.3	116.7	173.3	145.0	152.4
56		153.6	118.4	175.5	146.9	149.2
57		166.0	119.3	182.0	150.6	155.8
58		167.8	127.3	194.9	161.1	163.3
59		173.1	121.9	184.2	153.0	159.7
60		162.9	127.7	196.2	161.9	162.3
61		172.7	120.2	194.5	157.3	162.5
62		183.8	126.4	191.3	158.8	167.2
63		173.6	132.2	189.6	160.9	165.1
64		174.0	136.6	210.5	173.5	173.7
65		178.0	118.8	191.8	155.3	162.9
66		169.5	124.6	198.5	161.6	164.2
67		166.4	121.0	198.5	159.7	162.0
68		181.1	129.0	193.6	161.3	167.9
69	(N)	194.5	133.5	201.6	167.5	176.5
70		194.9	149.1	203.8	176.4	182.6
71		194.5	146.4	212.7	179.5	184.5
72		198.9	133.5	199.4	166.4	177.3
73		195.4	138.8	206.0	172.4	180.1
74		187.8	122.4	189.6	156.0	166.6
75		185.1	118.8	192.7	155.7	165.5

Continues

TABLE II (cont)

Stake	(1)	(2)	(3)	(4)	(5)
76	187.3	142.0	202.5	172.2	177.3
77	195.4	138.0	201.6	169.8	178.3
78	181.1	149.1	193.3	171.2	174.5
79	182.0	133.9	198.9	166.4	171.6
80	179.8	145.5	207.8	176.6	177.7
81	194.5	143.7	206.5	175.1	181.6
82	208.3	144.6	206.0	175.3	186.3
83	211.4	147.7	209.2	178.4	189.4
84	206.5	151.3	206.5	178.9	188.1
85	209.2	152.2	221.2	186.7	194.2
86	209.6	151.7	213.2	182.4	191.5
87 (P)	208.3	162.9	219.4	191.1	196.9
88	205.6	148.6	213.2	180.9	189.1
89	205.1	149.5	214.0	181.7	189.5
90	205.1	147.7	—	—	—
91	194.5	150.9	203.4	177.1	182.9
92	206.9	145.5	216.7	181.1	189.7
93	204.3	160.2	214.9	187.5	193.1
94	207.4	142.0	213.6	177.8	187.7
95	198.5	167.3	213.6	190.4	193.1
96	203.4	154.0	210.5	182.1	189.2
97	212.3	152.2	223.4	187.8	196.0
98	202.0	156.6	206.5	181.0	188.4
99	205.1	154.9	214.5	184.7	191.5
100	223.8	158.9	205.1	182.0	195.9
101 (R)	200.7	165.5	213.6	189.5	193.3
102	215.8	168.7	225.2	196.9	203.2
Average	153.7	109.4	169.8		

These figures represent:

29% decrease during the second year over the first year
10% increase during the third year over the first year
55% increase during the third year over the second year

* Isolined accumulation map (Figure 28)

movement stations was placed slightly up-profile from a noticable break of slope and from the curves in Figure 20 it appears that the peaks of accumulation are just below the crest of the breaks of slope and that the troughs are slightly above it. The very marked consistancy of this pattern over the study period leads to the conclusion that the effects of the high snow-bearing wind is to deposit larger amounts of snow in the lee of the slope than at the crest. There is no evidence to indicate scouring at the crest and redeposition of accumulation in the lee but this possibly does occur. Above station M, where surface topographic irregularities are virtually absent, the accumulation pattern is quite regular.

The curves in Figure 20 show a pronounced variation in total accumulation from year to year. During the first "year" from February 3 to December 5 1965, a period of only 10 months, the average accumulation was 153.7 g/cm^2. The second year, early December 1965 to mid-December 1966, recorded an average for the profile of only 109.4 g/cm^2 which represents a decrease over the first "year" of 29 percent. The average of 169.8 g/cm^2, recorded during the third year from mid-December 1966 to mid-December 1967, represents an increase of 55 percent over the preceding year and ten percent over the first year. The difference of ten percent between the first and third years' values can probably be accounted for by the shorter length of the first period but the much lower second period value is significantly real.

The differences in these values are not readily explainable from the contemporary meteorological records from Palmer Station (Appendix I

Vol 2) which show no greatly significant anamalies from year to year but subtle variations in the degree of winter climatic stability or "continentality" seems a logical explanation.

The principal snow-bearing winds are storm winds, associated with low pressure systems moving through the area and are from the NE quadrant (Appendix I). Comparison of the 1965 and 1966 records shows that the mean speed of all winds from this quadrant was 6.0 m/s (11.7 kts) in 1965 and 4.8 m/s (9.4 kts) in 1966. This might be taken to imply that 1965 was stormier than 1966 and therefore, that the higher 1965 accumulation resulted from a greater incidence of storm activity. However, the frequency of all winds from the NE quadrant was about the same, differing by only 0.3 percent; 7.6% in 1965 and 7.3% in 1966. Therefore, the incidence itself of the snow-bearing winds does not account for the accumulation differences. The most likely explanation is probably similar to that suggested by Schytt (1964) for Spitsbergen and is related to local and widespread variations in sea ice conditions which, in 1966 may have been more intense and imparted a more continental and therefore, cooler and dryer, character to the regional climate in that year. Markedly lower winter temperatures and a lower mean annual temperature were recorded at Palmer Station in 1966, which seems to support this suggestion (Appendix I).

The cumulative curves of gross accumulation for selected stations (figure24 p79 forward) indicate that the rate of snow accumulation in the early part of 1966, i.e. about March to June, was not as great as during the same period in the previous and following years. This implies that this period of relative deficiency accounts for the lower total accumulation for 1966. Again the temperature record appears to be supporting evidence; the 1966 autumn temperatures were markedly lower than in the previous and following years, suggesting an earlier development of regional sea ice and consequently a more rapid development of dry winter stability.

From Figure 20, it is also evident that the rate of increase of accumulation with elevation was more rapid in 1965 than in 1967; in spite of its being two months shorter, the 1965-period curve approaches that for 1967 with increasing elevation and at the highest elevations, as can be seen in Figure 21a below, becomes the dominant accumulation year.

MOUNTAIN LINE: STATIONS R TO LITTLE x

Completion of the Mountain line was not accomplished until March 15 1965. The last measurements were made on December 20 1967 by R.A. Honkala's group. On an annual basis, the record cannot be divided as readily as that for the Main line as the last measurements made by this author's party were on September 9 1966 and the records donated by Honkala cover the period September 9 1966 to December 20 1967 with a single, total value.

The first annual period has therefore, been defined as March 15 1965 to February 6 1966. The remaining record, February 6 1966 to December 20 1967, regarded as two years, has been averaged to give annual values. The total record (three years) has also been averaged. The results are shown in Tables III and IV and form Figure 21b.

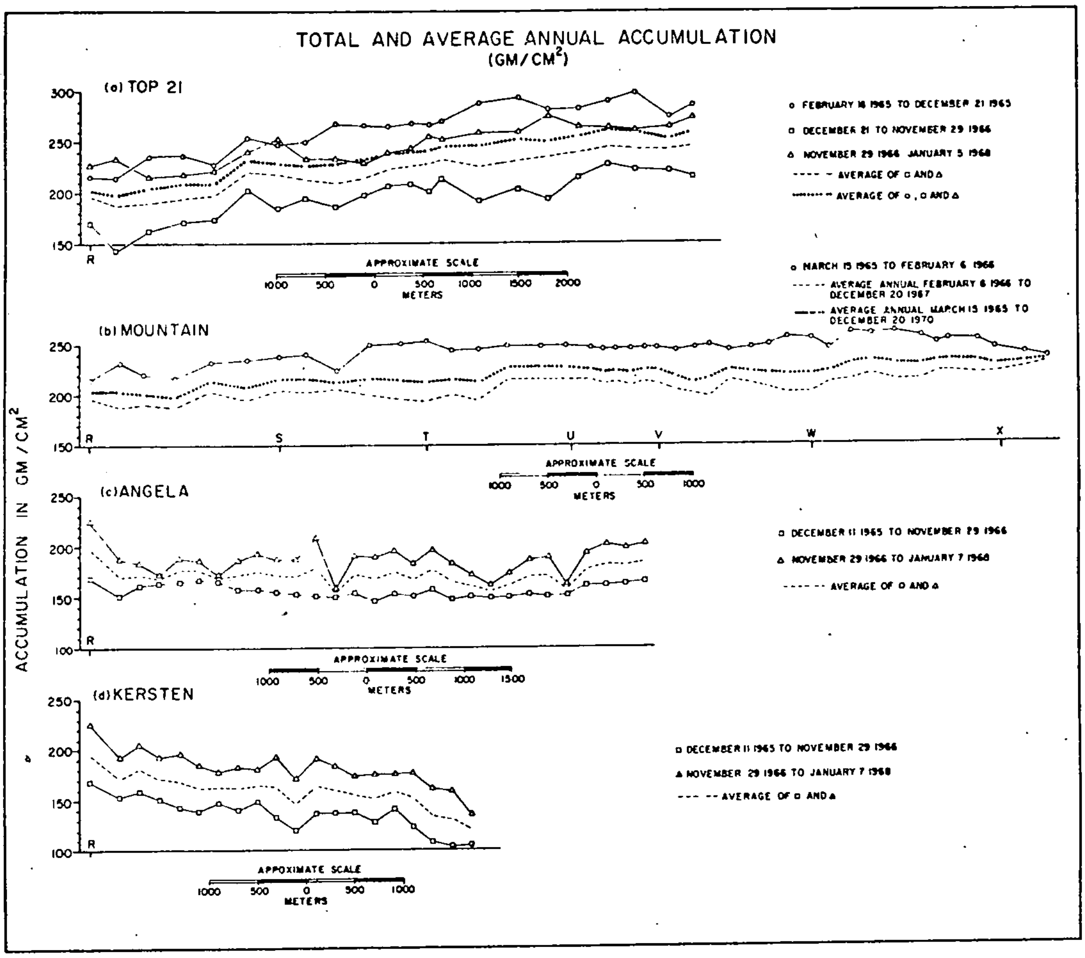

TOTAL AND AVERAGE ANNUAL ACCUMULATION
(GM/CM²)
(a) TOP 21
FEBRUARY 16 1965 TO DECEMBER 21 1965
DECEMBER 21 TO NOVEMBER 29 1966
NOVEMBER 29 1966 JANUARY 5 1968
AVERAGE OF □ AND △
AVERAGE OF ○, □ AND △
APPROXIMATE SCALE
METERS
(b) MOUNTAIN
MARCH 15 1965 TO FEBRUARY 6 1966
AVERAGE ANNUAL FEBRUARY 6 1966 TO DECEMBER 20 1967
AVERAGE ANNUAL MARCH 15 1965 TO DECEMBER 20 1970
(c) ANGELA
DECEMBER 11 1965 TO NOVEMBER 29 1966
NOVEMBER 29 1966 TO JANUARY 7 1968
AVERAGE OF □ AND △
(d) KERSTEN
DECEMBER 11 1965 TO NOVEMBER 29 1966
NOVEMBER 29 1966 TO JANUARY 7 1968
AVERAGE OF □ AND △
APPOXIMATE SCALE
ACCUMULATION IN GM/CM²

Fig. 21

TABLE III

TOTAL ACCUMULATION (gm/cm^2) MOUNTAIN LINE
March 15 1965 to February 6 1966

Stake	gm/cm^2	Stake	gm/cm^2	Stake	gm/cm^2
1	232.4	15(U)	248.2	29	256.4
2	221.1	16	248.7	30	253.7
3	217.4	17	247.3	31	244.6
4	233.3	18	241.9	32(W)	260.0
5(S)	235.6	19	244.6	33	258.7
6	238.3	20	245.1	34	261.4
7	240.1	21	246.4	35	255.9
8	224.2	22	246.4	36	250.0
9	249.1	23	243.3	37	253.2
10(T)	251.0	24	245.9	38(x)	253.2
11	253.2	25(V)	248.7	39	243.8
12	243.8	26	243.3	40	239.2
13	245.5	27	245.9	41	236.5
14	247.8	28	248.7		

TABLE IV

AVERAGE ANNUAL ACCUMULATION (gm/cm^2) MOUNTAIN LINE

(1) February 6 1966 to December 20 1967
(2) March 15 1965 to December 20 1967

Stake	(1)	(2)	Stake	(1)	(2)	Stake	(1)	(2)
1	188.7	203.2	15	214.4	225.7	29	199.7	218.6
2	191.9	201.6	16	214.5	225.9	30	200.9	218.5
3	188.0	197.8	17	214.3	225.3	31	211.3	222.4
4	204.2	213.9	18	209.9	220.6	32	213.3	228.9
5	195.1	208.6	19	212.0	222.9	33	218.8	232.1
6	205.4	216.4	20	208.3	220.6	34	213.1	229.2
7	203.0	215.4	21	213.4	224.4	35	214.0	228.0
8	206.5	212.4	22	211.1	222.9	36	220.9	230.6
9	200.5	216.7	23	203.4	216.7	37	220.8	231.6
10	195.9	214.3	24	199.8	211.5	38	219.3	230.6
11	193.5	213.4	25	197.2	214.4	39	219.2	227.4
12	200.0	214.6	26	214.5	224.1	40	224.5	229.4
13	194.8	211.7	27	210.5	222.3	41	230.5	232.5
14	214.6	225.7	28	206.4	220.5			

The curves in Figure 21b are a direct continuation of those from the Main line and the trends noted in the upper parts of the Main line are evident. There is but a very slight positive gradient to the curves in spite of a steady increase in elevation of about 200 meters between stations R and Little x. It appears possible also that the total accumulation during the first period was greater, in parts at least, than during the third year even though the first was somewhat less than a calendar year. This can be deduced from consideration of the upper Main line curves (also the Top-21 stakes) where the values for the two periods run close together with the first period values eventually becoming the greater. The average for the final two years ranged from 180 to 230 g/cm^2, thus, unless the second year's value decreased considerably inland, say to 150 g/cm^2 or less in places, the final year's value could not have reached as high as that for the first. This is confirmed by the three-year average curve.

As with the Main line curves, it is notable that in spite of the much lower accumulation in the second year, the three-year and the two-year averages are not widely separated though on the Mountain line the separation is generally the greater. Average accumulation along the Mountain line in the first year was 245.1 g/cm^2 and the average for the remaining two years was 207.7 g/cm^2. Average annual accumulation, based on three year's data, was 220.2 g/cm^2.

TOP 21 STAKES

On February 16 1965, 100 bamboo poles were emplaced on a line running northward for a distance of 35 km. from station R. It was 5 to 6 km from, but generally parallel to the Achaean Mountain Range. The line terminated just north of Perrier Bay and Mt. Nestor (see figure 1) about 15 km. from the northern coast. The objective was to obtain accumulation data relative to latitude and aspect along the entire length of the piedmont as well as relative to increasing elevation from station R to the local drainage basin divide. The divide, as well as could be determined in the field, was marked by the 21st pole from station R. The entire line was called Top Line Trail.

Maintenance of the entire line overtaxed the ability of the field parties and all but the first 21 stakes were abandoned or lost after September 2 1965. Thus, a record for the entire line is available for only part of the first year (Table XVII). A complete record is available from the first 21 stakes and total and average accumulation is shown in Table V and Figure 21a.

Immediately evident in the curves in Figure 21a is the steady increase in the rate of accumulation (with increasing elevation), in notable contrast to the trend along the Mountain line where the elevation increase is about the same (750 m. at station U and 744 m. at T5 (see figure 19). The much lower total values during the second year are also strongly evident. For the first and final year's values however, the trend seen in the upper Main line is more emphasised and the sequence has generally reversed with the higher rate having been recorded in the first year in much of this area. During the first year the average accumulation was 265.3 g/cm^2; in the second, 196.4 g/cm^2 and 247.0 g/cm^2 during the third year. Based on the three years' data, average annual accumulation was 236.2 g/cm^2. In terms of percentage

TABLE V

TOTAL AND AVERAGE ANNUAL ACCUMULATION
(gm/cm^2) TOP 21 STAKES

(1) February 16 1965 to December 21 1965)
(2) December 21 1965 to November 29 1966) (4) Average*) (5) Average
(3) November 29 1966 to January 5 1968))

Stake	(1)	(2)	(3)	(4)	(5)
1	214.7	141.8	233.3	187.5	196.6
2	235.6	162.0	215.2	188.6	204.3
3	236.5	170.3	217.9	194.1	208.2
4	228.7	173.0	222.0	197.5	207.9
5	254.1	200.2	240.5	220.3	231.6
6	246.9	184.2	252.8	218.5	228.0
7	250.5	194.2	232.8	213.5	225.8
8	267.3	185.2	232.8	209.0	228.4
9	265.9	197.5	228.3	212.9	230.6
10	265.4	206.0	239.6	222.8	237.0
11	268.2	208.3	243.7	226.0	240.1
12	266.8	200.6	254.6	227.6	240.7
13	270.4	213.7	251.9	232.8	245.3
14	289.0	191.5	258.7	225.1	246.4
15	294.4	203.2	259.6	231.4	252.4
16	281.8	193.3	275.9	234.6	250.3
17	282.2	214.3	264.5	239.4	253.7
18	291.3	227.4	263.6	245.5	260.8
19	299.4	222.3	262.3	242.3	261.3
20	275.4	220.9	264.1	242.5	253.5
21	286.7	215.2	274.1	244.6	258.7

* Isolined accumulation map (Figure 28)

variation from year to year this represents 26% less accumulation in the second year than in the first and 7% less in the third than the first. The third year showed a 26% increase over the second. Overall, when compared with the values from the entire Main line, these latter two values show a total difference between the two lines of 17% between the third and first years and 19% between the third and the second.

The reason for the reversal at these higher elevations is obscure and the available meteorological record is not helpful toward an explanation. Air temperatures in 1967 were generally higher than in the two previous years which could lead to the conclusion of decreased "continentality", lower sea ice concentrations and consequently, increased precipitation. This apparently was not the case. Mere speculation suggests that the slightly cooler summer of 1967 together with residual cold from the relatively severe winter of 1966 (see Appendix I) helped to maintain a degree of winter "continentality" somewhat out of proportion to the air temperatures recorded at Palmer Station. The basic conclusion is drawn in Chapter 13 (below); local precipitation patterns cannot be satisfactorily explained on the basis of local meteorological data alone. The problem requires a regional approach, with a study of the interaction of sea ice conditions, air temperature distribution, local wind patterns and the changing position of depression tracks, all of which affect the relative distribution of precipitation throughout the peninsula.

ANGELA AND KERSTEN LINES

The Angela and Kersten lines were established on December 11 1965 and were last measured by Honkala's group on January 7 1968. The record is divided into two years at November 29 1966 (Tables VI & VII).

There is virtually no gradient to the curves from the Angela line in spite of an elevation increase of over 100 m. over its length. The Kersten line however, exhibits a pronounced negative accumulation gradient over the same increase of elevation. From both lines the same pattern of accumulation rates emerges with the 1966 values being significantly below those for 1967 (figures 21c & 21d). Average annual accumulation in 1966 was 155.9 g/cm^2 at the Angela stations and 134.4 g/cm^2 along the Kersten line. The 1967 rate was 185.3 g/cm^2 and 179.5 g/cm^2 for the two lines respectively. The 1967 values represent an 18% increase over 1966 on the Angela line and 33% on the Kersten line. Average annual accumulation for the two-year period was 170.6 g/cm^2 (Angela) and 156.9 g/cm^2 (Kersten).

NEUMAYER LINE

The Neumayer line was established on April 11 1965 and the last measurements by Honkala were made on December 26 1967. Thirty nine stakes were originally emplaced and the line terminated at that point only because access beyond was prevented by the crevasse fields. Movement of the profile ultimately carried the lower stakes into the crevassed area and the last three poles were abandoned after the first year. As the last measurements by this author's party were on September 19 1966 and Honkala's first were on April 17 1967, the record cannot be divided into annual units similar to most of the other lines. Total accumulation from April 11 to December 21 1965 and average annual accumulation from December 21 1965 to December 26 1967 are shown in Tables VIII and IX and form Figure 22a.

There is a steady decline in accumulation rates with decreasing altitude from station R. There is also a correlation between the pattern of accumulation and surface topography. The surface slope between sta-

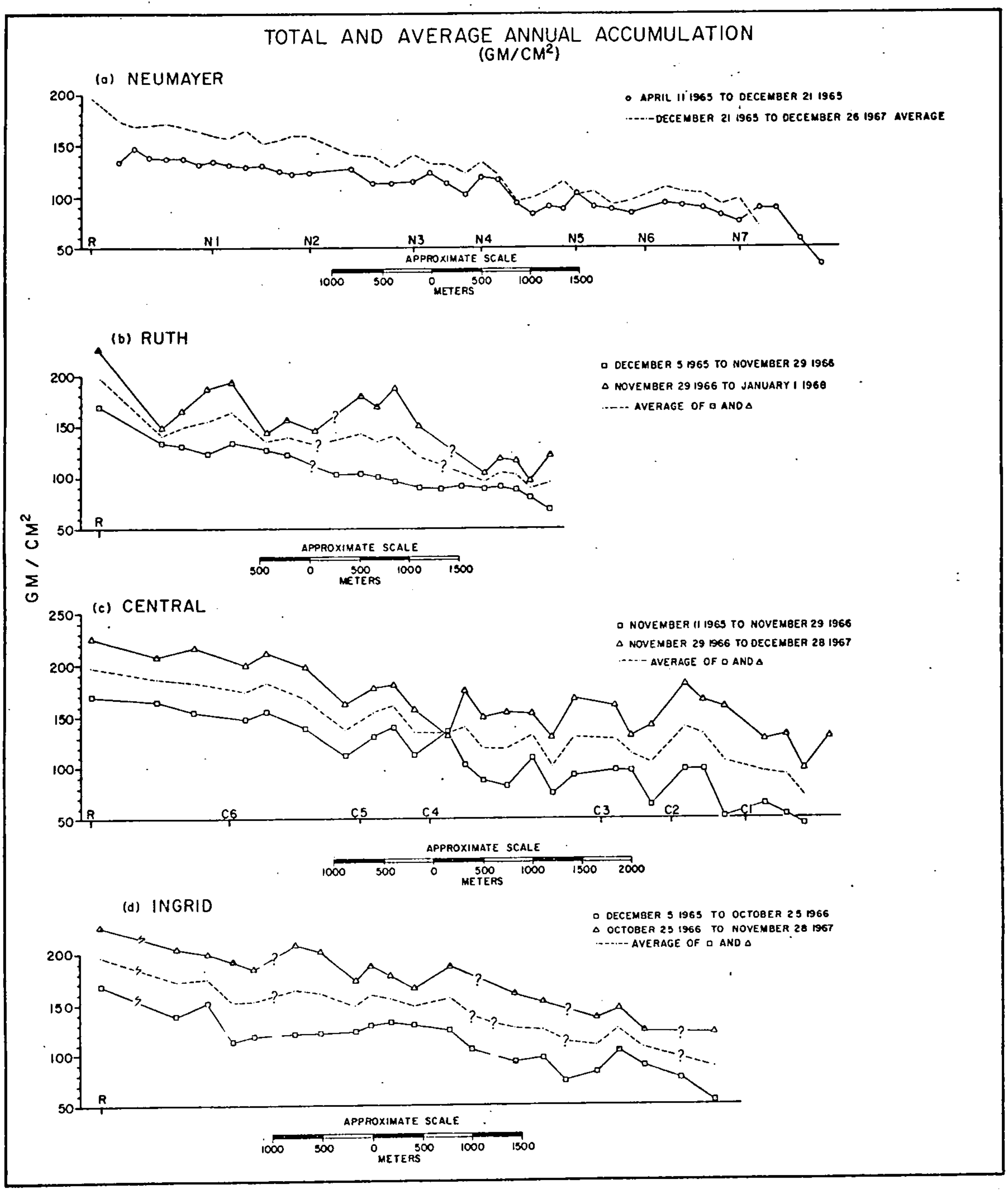

TOTAL AND AVERAGE ANNUAL ACCUMULATION
(GM/CM²)
GM / CM²
(a) NEUMAYER
o APRIL 11 1965 TO DECEMBER 21 1965
DECEMBER 21 1965 TO DECEMBER 26 1967 AVERAGE
R
N1
N2
N3
N4
N5
N6
N7
APPROXIMATE SCALE
1000 500 0 500 1000 1500
METERS
(b) RUTH
DECEMBER 5 1965 TO NOVEMBER 29 1966
NOVEMBER 29 1966 TO JANUARY 1 1968
AVERAGE OF AND
APPROXIMATE SCALE
500 0 500 1000 1500
METERS
(c) CENTRAL
NOVEMBER 11 1965 TO NOVEMBER 29 1966
NOVEMBER 29 1966 TO DECEMBER 28 1967
AVERAGE OF AND
C6
C5
C4
C3
C2
C1
APPROXIMATE SCALE
1000 500 0 500 1000 1500 2000
METERS
(d) INGRID
DECEMBER 5 1965 TO OCTOBER 25 1966
OCTOBER 25 1966 TO NOVEMBER 28 1967
AVERAGE OF AND
APPROXIMATE SCALE
1000 500 0 500 1000 1500
METERS

Fig. 22

TABLE VI

TOTAL AND
AVERAGE ANNUAL ACCUMULATION (gm/cm^2) ANGELA LINE

(1) December 11 1965 to November 29 1966
(2) November 29 1966 to January 7 1968
(3) December 11 1965 to January 7 1968

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	151.7	187.1	169.4	15	154.0	196.2	175.1
2	160.4	183.4	171.9	16	150.8	182.6	166.7
3	164.0	171.6	167.8	17	157.6	197.0	177.3
4	164.9	188.9	176.9	18	147.7	182.9	165.3
5	167.6	186.2	176.9	19	151.3	171.7	161.5
6	164.9	171.7	168.3	20	149.0	162.2	155.6
7	157.2	186.2	171.7	21	150.4	172.6	161.5
8	157.6	192.6	175.1	22	153.1	187.5	170.3
9	154.9	187.1	171.0	23	150.8	189.2	170.0
10	152.7	187.1	169.9	24	152.2	160.8	156.5
11	150.8	208.0	179.4	25	160.8	193.0	176.9
12	149.5	158.5	154.0	26	161.7	202.5	182.1
13	154.5	190.3	172.4	27	163.5	198.3	180.9
14	145.9	189.3	167.6	28	165.8	203.2	184.5

TABLE VII

TOTAL AND
AVERAGE ANNUAL ACCUMULATION (gm/cm^2) KERSTEN LINE

(1) November 11 1965 to November 29 1966
(2) November 29 1966 to January 7 1968
(3) November 11 1965 to January 7 1968

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	151.7	192.1	171.9	11	136.8	190.8	163.8
2	158.1	205.1	181.6	12	136.8	183.8	160.3
3	150.4	193.4	171.9	13	137.7	172.7	155.2
4	142.2	196.6	169.4	14	128.2	175.6	151.9
5	139.5	184.3	161.9	15	140.9	176.1	158.5
6	147.7	178.1	162.9	16	123.7	177.5	150.6
7	140.4	183.4	161.9	17	107.8	161.2	134.5
8	149.0	181.2	165.1	18	103.3	159.1	131.2
9	132.7	193.1	162.9	19	105.1	135.5	120.3
10	120.9	170.9	145.9				

TABLE VIII

AVERAGE ANNUAL ACCUMULATION (gm/cm^2) NEUMAYER LINE
December 21 1965 to December 26 1967

Stake	gm/cm^2	Stake	gm/cm^2
1	173.3	19	131.0
2	168.2	20	121.7
3	169.0	21 (N4)	134.0
4	170.6	22	120.5
5	167.3	23	96.3
6	163.0	24	99.3
7 (N1)	158.8	25	106.1
8	157.5	26	114.7
9	163.9	27 (N5)	101.0
10	150.9	28	104.6
11	155.3	29	91.5
12	159.3	30	95.0
13 (N2)	158.1	31 (N6)	108.5
14	140.8	32	104.0
15	139.2	33	101.8
16	127.3	34	92.0
17	140.1	35 (N7)	99.7
18 (N3)	130.8	36	70.9

TABLE IX

TOTAL ACCUMULATION (gm/cm^2) NEUMAYER LINE
April 11 1965 to December 21 1965

Stake	gm/cm^2	Stake	gm/cm^2	Stake	gm/cm^2
1	133.9	14	127.3	27	103.4
2	146.8	15	113.5	28	90.5
3	138.4	16	113.5	29	87.5
4	136.6	17	114.4	30	83.2
5	137.1	18	122.8	31	93.9
6	131.3	19	113.0	32	91.4
7	134.4	20	101.9	33	90.1
8	130.4	21	119.3	34	81.1
9	129.5	22	117.5	35	73.8
10	129.9	23	95.7	36	87.9
11	123.7	24	85.2	37	88.4
12	122.4	25	90.5	38	58.3
13	122.8	26	87.9	39	34.3

tions R and N3 was, for the most part, constant with no significant topographic irregularities; in fact N3 was visible (by theodolite telescope) from station R. Down-profile from N3 however, the surface was undulating, more in the form of steps than ridges and troughs and stations N3 through N7 were set near the edge of successive steps. In Figure 22a the evidence is for slightly greater accumulation in the lee of the steps and less up-profile (up-wind) of the break of slope. The cause seems similar to that in the lower Main line where the greater accumulation in the lee is attributed to slackening wind speed and eddy effects and to possible scouring of the crest. Such evidence appears even more pronounced along the Central line (figure 22c), where the movement stations were similarly placed relative to even more pronounced surface topography.

Total accumulation along the Neumayer line between April 11 1965 and December 21 1965 was 107.2 g/cm^2. The two-year period, December 21 1965 to December 26 1967 yields an average annual value of 130.2 g/cm^2

RUTH, CENTRAL AND INGRID LINES

The record from the Central line dates from November 11 1965 and was kept until December 28 1967. That from Ruth and Ingrid began on December 5 1965 and ended on January 1 1968 and November 28 1967 respectively. The record is documented in Figure 22b, c & d and Tables X, XI and XII.

Along all three profiles there is a close correlation between the rate of accumulation and the rate of elevation change. As noted above

TABLE X

TOTAL AND

AVERAGE ANNUAL ACCUMULATION (gm/cm^2) RUTH LINE

(1) December 5 1965 to November 29 1966
(2) November 29 1966 to January 1 1968
(3) December 5 1965 to January 1 1968

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	132.9	147.3	140.1	11	93.8	185.6	139.7
2	131.6	163.8	147.7	12	87.6	148.2	117.9
3	122.4	186.0	154.2	13	87.2	—	—
4	133.0	192.2	162.6	14	89.4	—	—
5	124.6	142.4	133.5	15	86.2	100.8	93.5
6	120.6	156.6	138.6	16	88.0	116.2	102.1
7	—	—	—	17	85.3	114.5	99.9
8	100.6	—	—	18	77.6	94.4	86.0
9	102.8	178.4	140.6	19	65.5	119.3	92.4
10	99.7	167.3	133.5				

TABLE XI

TOTAL AND

AVERAGE ANNUAL ACCUMULATION (gm/cm^2) INGRID LINE

(1) December 5 1965 to October 25 1966
(2) October 25 1966 to November 28 1967
(3) December 5 1965 to November 28 1967

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	52.8	121.2	87.0	12	127.8	165.6	146.7
2	76.8	—	—	13	132.6	178.4	155.5
3	89.2	122.6	105.9	14	129.5	188.5	159.0
4	104.2	145.0	124.6	15	123.3	173.5	148.4
5	82.8	136.0	109.4	16	121.0	201.6	161.3
6	73.3	—	—	17	120.6	207.8	164.2
7	97.4	151.8	124.6	18	—	—	—
8	92.7	159.9	126.3	19	118.4	185.4	151.9
9	—	—	—	20	112.6	191.8	152.2
10	105.1	—	—	21	150.4	199.8	175.1
11	125.7	186.1	155.9	22	139.3	204.3	171.8

TABLE XII

TOTAL AND

AVERAGE ANNUAL ACCUMULATION (gm/cm^2) CENTRAL LINE

(1) November 11 1965 to November 29 1966
(2) November 29 1966 to December 28 1967
(3) November 11 1965 to December 28 1967

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	----	130.0	----	15	109.4	152.2	130.8
2	44.4	97.4	70.9	16	82.3	153.5	117.9
3	53.1	131.3	92.2	17	86.7	147.7	117.2
4	63.9	126.5	95.2	18	----	----	----
C1				19	102.8	174.4	138.6
6	51.1	158.3	104.7	20(C4)	133.9	131.3	132.6
7	98.4	165.2	131.8	21	112.1	155.7	133.9
8 (C2)	98.2	180.0	139.1	22	139.6	180.2	159.9
9	63.5	140.3	101.9	23(C5)	130.3	177.1	153.7
10	96.0	130.0	113.0	24	111.2	160.6	135.9
11	96.1	158.7	127.4	25	138.2	196.2	167.2
12 (C3)				26	153.9	210.5	182.2
13	92.4	166.4	129.4	27	147.3	198.9	173.1
14	74.2	128.6	101.4	28(C6)	----	----	----

the relationship between surface topography and the pattern of accumulation is markedly evident in the lower parts of the Central line, a relationship which may explain the variations on the Ruth and Ingrid lines (though no notes are available on topographic detail on these lines except that undulations were pronounced). The lengths of the individual annual periods for these profiles vary slightly but generally are comparable and average accumulation during the first year was 101.6 g/cm^2 (Ruth), 103.9 g/cm^2 (Central) and 108.8 g/cm^2 (Ingrid). During the second year, 147.3 g/cm^2, 160.6 g/cm^2 and 171.7 g/cm^2 respectively, and average annual accumulation over the two-year period was 125.5 g/cm^2 (Ruth), 132.8 g/cm^2 (Central) and 142.3 g/cm^2 (Ingrid). Considered in terms of percentage increase in the second year over the first, these figures represent 45% on the Ruth line, 55% on the Central line and 58% on the Ingrid line.

These figures themselves are interesting in that they indicate a trend toward greater accumulation rates outward from the mountains (see Figure 19) and also an increasing percentage variation between the two years. In the absence of detailed topographic information, it can only be speculated that the increase in rate is topographically determined but the possibility that this results from a wide-ranging effect of the mountains and a resultant precipitation shadow cannot be overlooked. The increasing difference between the two year's values, if indeed significant, is difficult to explain. Possibly it results from differences in local climatological and meteorological patterns.

MIDDLE LINE

The record from the Middle line is sparse and of an original 20 stakes emplaced on December 23 1965, only 10 were recovered which provide an adequate record of accumulation. Heavy losses resulted in this area from ice buildup on the poles and subsequent breakage in high winds. Also, many poles seem to have been buried in the early parts of 1967. The record cannot be divided into two years. The last measurement by this author were on September 7 1966 and the first by Honkala's group was on April 11 1967. The final measurements were on December 19 1967. The data are insufficient for graphical discussion and are tabulated below in Table XIII. The stakes are numbered from station Mu1 and elevation increases with increasing number.

TABLE XIII

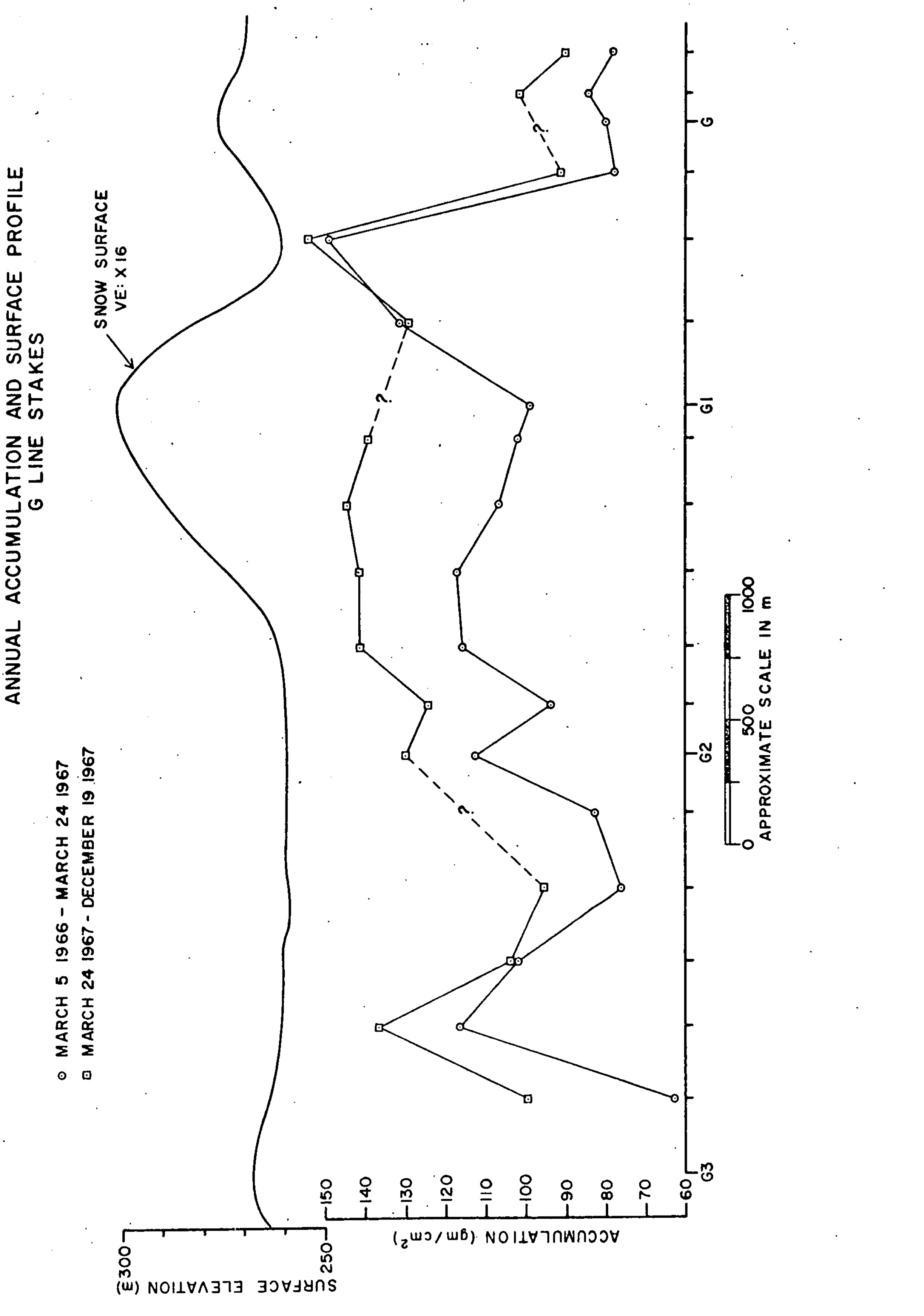
AVERAGE ANNUAL ACCUMULATION (g/cm^2) MIDDLE LINE
December 23 1965 to December 19 1967

Stake	g/cm^2	Stake	g/cm^2	Stake	g/cm^2
1	131.2	8	---	15	---
2	125.2	9	---	16	198.0
3	126.3	10	191.7	17	---
4	157.0	11	172.9	18	---
5	148.0	12	---	19	---
6	150.2	13	164.9	20	---
7	---	14	---		

An increase in accumulation rates with increasing altitude is evident and is similar in magnitude to other longitudinal profiles. Though the data are sparse there are marked irregularities which might be related to topographic irregularites which in this area are pronounced. Possibly accumulation rates vary more in this area than in the region of the Main profile. Were more values available for graphical inspection, an accumulation pattern similar to that in the lower Central line might emerge.

G LINE STAKES

Data from 15 of an original 18 stakes forming the G line are available from March 5 1966 to December 19 1967. The record has been divided into two periods at March 24 1967 and is shown in Figure 23 and Table XIV.



ANNUAL ACCUMULATION AND SURFACE PROFILE
G LINE STAKES
⊙ MARCH 5 1966 - MARCH 24 1967
⊡ MARCH 24 1967 - DECEMBER 19 1967
SNOW SURFACE
VE: X16
?
?
?
?
G3
G2
G1
G
0
500
1000
APPROXIMATE SCALE IN m
60
70
80
90
100
110
120
130
140
150
ACCUMULATION (gm/cm²)
250
300
SURFACE ELEVATION (m)

Fig. 23

TABLE XIV

TOTAL AND AVERAGE ANNUAL ACCUMULATION (g/cm^2) G LINE

(1) March 5 1966 to March 24 1967
(2) March 24 1967 to December 19 1967
(3) March 5 1966 to December 19 1967

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	78.2	90.1	84.1	10	117.1	141.4	129.2
2	84.4	101.5	92.9	11	116.0	141.4	128.7
3(G)	79.8	---	---	12	93.7	124.3	109.0
4	77.7	91.2	84.4	13(G2)	112.9	130.0	121.4
5	149.2	154.4	151.8	14	82.9	---	---
6	131.6	129.5	130.5	15	76.1	95.3	85.7
7(G1)	97.9	---	---	16	102.0	103.6	102.8
8	102.0	139.3	120.6	17	116.5	136.7	126.6
9	106.7	144.5	126.1	18 G3	62.7	99.4	81.0

This was a "transverse" profile and was set mainly to investigate the effects of pronounced local topography on accumulation rates. An original plan to survey the surface profile with level and staff was not realised due mainly to shortage of time. However, notations and sketches from field notebooks provide a qualitative schematic of the profile shape which is included in Figure 23.

On the basis of these data there appears to be a strong influence on the pattern of accumulation rates by the surface topography. The pattern is similar to that on the lower Main line and Central line with the rate of accumulation being greater in the troughs than on the higher, more exposed surfaces.

During the first year the average accumulation from the eighteen stakes was 99.3 g/cm^2. The 15 stakes remaining in the second years recorded an average of 121.5 g/cm^2 and for the two-year period an average annual accumulation of 111.6 g/cm^2. Even though the second period was considerably shorter than the first, the 15 stakes with unbroken records recorded 20% more accumulation during the shorter period.

THERESE LINE AND NORTH FORK

On December 21 1965, 29 stakes were emplaced to form the Therese line and 13 stakes were set to form North Fork. Both lines were successfully maintained until January 5 1967 when this author's party made its final measurements. The results from the Therese line are tabulated in column one in Table XV.

TABLE XV

TOTAL AND AVERAGE ANNUAL ACCUMULATION (g/cm^2) THERESE LINE

(1) December 21 1965 to January 5 1967
(2) January 5 1967 to January 6 1968
(3) December 21 1965 to January 6 1968

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	195.3	278.4	236.8	15	197.1	---	---
2	193.1	271.5	232.3	16	200.7	---	---
3	190.9	279.7	235.3	17	208.3	255.7	232.0
4	202.4	290.0	246.2	18	195.3	265.5	230.4
5	199.8	---	---	19	195.3	252.2	223.7
6	193.6	---	---	20	---	258.7	---
7	203.8	292.6	248.2	21	209.3	233.4	221.3
8	208.3	306.3	257.3	22	186.2	263.4	224.8
9	214.0	---	---	23	175.9	234.7	205.3
10	207.4	251.4	229.4	24	179.7	231.7	205.7
11	205.1	258.7	231.9	25	172.9	218.4	195.6
12	202.9	270.3	236.6	26	168.2	216.2	192.2
13	203.4	---	---	27	157.4	222.2	189.8
14	197.0	---	---				

On April 11 1967, Honkala's group made its first measurement of the lines and snow accumulation amounting to 140-170 cm. was recorded for the three-month period. The poles were reset and were high, one standing 290 cm. On August 24 1967 the entire line was missing. It was replaced the following day and was maintained until January 6 1968.

If it is assumed that the original poles were simply buried between April and August and that t he snowfall was just equal to the pole heights, the total snowfall recorded by the poles from January 5 1967 to January 6 1968 yields the water equivalent values shown in column two of Table XV. Though it is not impossible that almost 3 m. of snow fell in this area between April and August it is this author's opinion that it is improbable; the cumulative curves of snowfall for other areas (figure 24), though reflecting a sharp increase in accumulation rates between mid-July and early September, do not bear out a snowfall of such magnitude. Stake number 14 of the Top 21 line, closest to the first on Therese, recorded 203 cm. of snowfall in the same period. The fate of the stakes between April and August is the more likely explained by ice buildup, high winds and consequent breakage - perhaps not the entire line but enough to mislead the field parties. In any case, should any have been located, identification would have been a monumental, if not impossible task. The values in Table XV (2) and (3) are therefore, suspect.

The history of North Fork is little better. Of the original 13 stakes set in December 1965, eight were recovered with complete records by January 5 1967. The line was reset with a full complement of

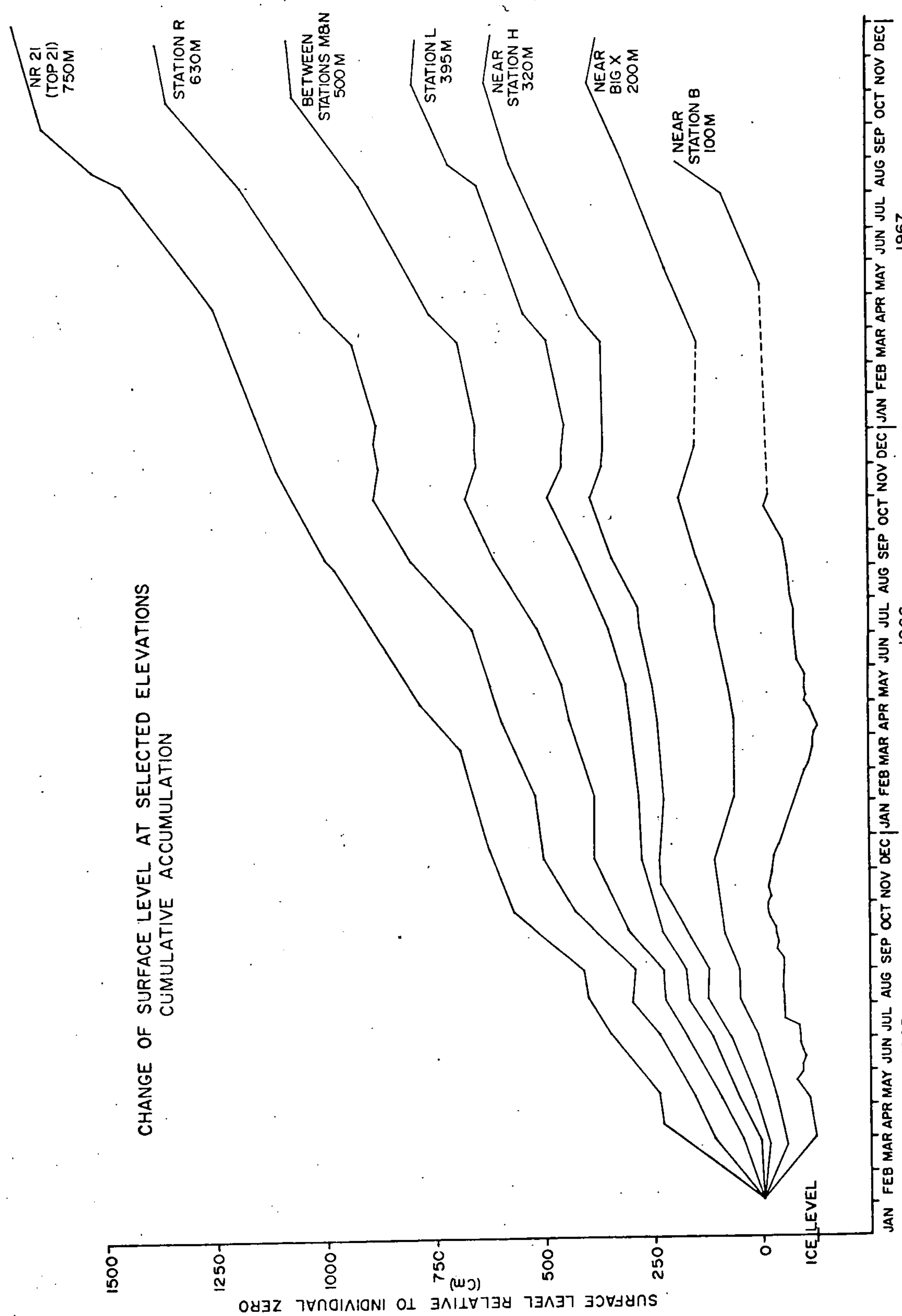

CHANGE OF SURFACE LEVEL AT SELECTED ELEVATIONS
CUMULATIVE ACCUMULATION
NR 21 (TOP 21) 750M
STATION R 630M
BETWEEN STATIONS M&N 500M
STATION L 395M
NEAR STATION H 320M
NEAR BIG X 200M
NEAR STATION B 100M
ICE LEVEL
SURFACE LEVEL RELATIVE TO INDIVIDUAL ZERO (cm)
1500
1250
1000
750
500
250
0
JAN FEB MAR APR MAY JUN JUL AUG SEP OCT NOV DEC
1965
JAN FEB MAR APR MAY JUN JUL AUG SEP OCT NOV DEC
1966
JAN FEB MAR APR MAY JUN JUL AUG SEP OCT NOV DEC
1967

Fig. 24

stakes but when they were next serviced on April 11 1967, Honkala found only six and reset them, some of them high. Two of these six were lost by May 12 but were replaced. On August 11 five remained and by august 24 only four remained. An unbroken record is therefore, available for only 3 stakes. The other one has a "lost" period of one month. Total accumulation from these stakes for the two years is listed below: (1) December 21 1965 to January 5 1967, (2) January 5 1967 to January 6 1968, (3) two-year average where available.

North Fork Stakes

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	140.0	---	---	8	---	142.0	---
2	137.7	---	---	9	137.7	173.3	155.5
3	---	---	---	10	137.7	131.3	134.5
4	---	---	---	11	146.7	143.7	145.2
5	---	---	---	12	119.3	---	---
6	140.3	---	---	13	106.4	---	---
7	---	---	---				

It is notable that the three stakes with an unbroken record, only one, number nine, shows an increase in 1967. This may be the result of mis-identification in the field at some time.

The data are interesting in another respect. Neither profile extended to particularly low elevation (the first pole on North Fork is estimated to be at about 300 meters a.s.l. or about the same as station H on the Main line -(G5 is at 281 m. a.s.l.). At about 300 m. elevation on the Main line the annual accumulation rate was about 70 g/cm^2 in 1966 and about 110 g/cm^2 in 1967, considerably lower than the North Fork values. This may be due to the effect of aspect and exposure to the snow-bearing wind in that the lower Main profile lies in the "lee" of Anvers Island and receives less snowfall as a result. North Fork on the other hand, stands in full exposure to the wind. Bryan (1965) proposed a similar argument to explain the lower accumulation rates in the southern part of Adelaide Island (the lee of the island) as compared with the northern, windward end.

SKYLINE AND MONACO LINE

The Skyline and Monaco line follow, as closely as could be determined in the field, the drainage basin divide and defined the northern geographic boundary of the study area. The record from the Skyline is extremely poor and regretably, no sensible record was recovered from the Monaco line.

The Monaco line was first set with 48 poles on April 15 and 16 1965 and was visible on September 2 when the Top Line Trail was serviced. The line was lost in a storm from September 3-7 (which in the vicinity of Lapatrere Bay in the north, deposited 97 cm. of fresh snow). It was eventually reset with 32 stakes on December 21 1965. By January 3 1967 only 5 stakes remained. On that date it was again rebuilt with 28 stakes but was lost again, by Honkala's party, by August 11 1967 (a similar fate to the Therese line). After being rebuilt, again with 28 stakes, between August 16 and 18, Honkala's party maintained it until January 6 1968. By that date only 17 remained to give a record of accumulation from mid-August 1967 to early January 1968.

The Skyline was established with 15 stakes on December 21 1965 and by November 29 1966, 12 remained with unbroken records. The line was then maintained until January 5 1968, but only 7 stakes were recovered with complete records. The results are shown in Table XVI and reflect an average accumulation of 205.4 g/cm^2 at the 12 stakes during 1966 and 281.9 g/cm^2 at the 7 stakes in 1967. The average annual accumulation at the 7 stakes with a continuous record was 242.6 g/cm^2. There was a 31% increase in 1967 over 1966.

TABLE XVI

TOTAL AND AVERAGE ANNUAL ACCUMULATION (g/cm^2) SKYLINE

(1) December 21 1965 to November 29 1966
(2) November 29 1966 to January 5 1968
(3) December 21 1965 to January 5 1968

Stake	(1)	(2)	(3)	Stake	(1)	(2)	(3)
1	204.7	---	---	9	202.0	---	---
2	212.0	---	---	10	210.6	288.1	249.3
3	---	---	---	11	216.5	---	---
4	205.2	279.4	242.3	12	208.8	286.3	247.5
5	---	---	---	13	205.2	280.7	242.9
6	197.9	283.2	240.5	14	205.7	---	---
7	---	---	---	15	197.5	274.1	235.8
8	198.4	281.3	239.8				

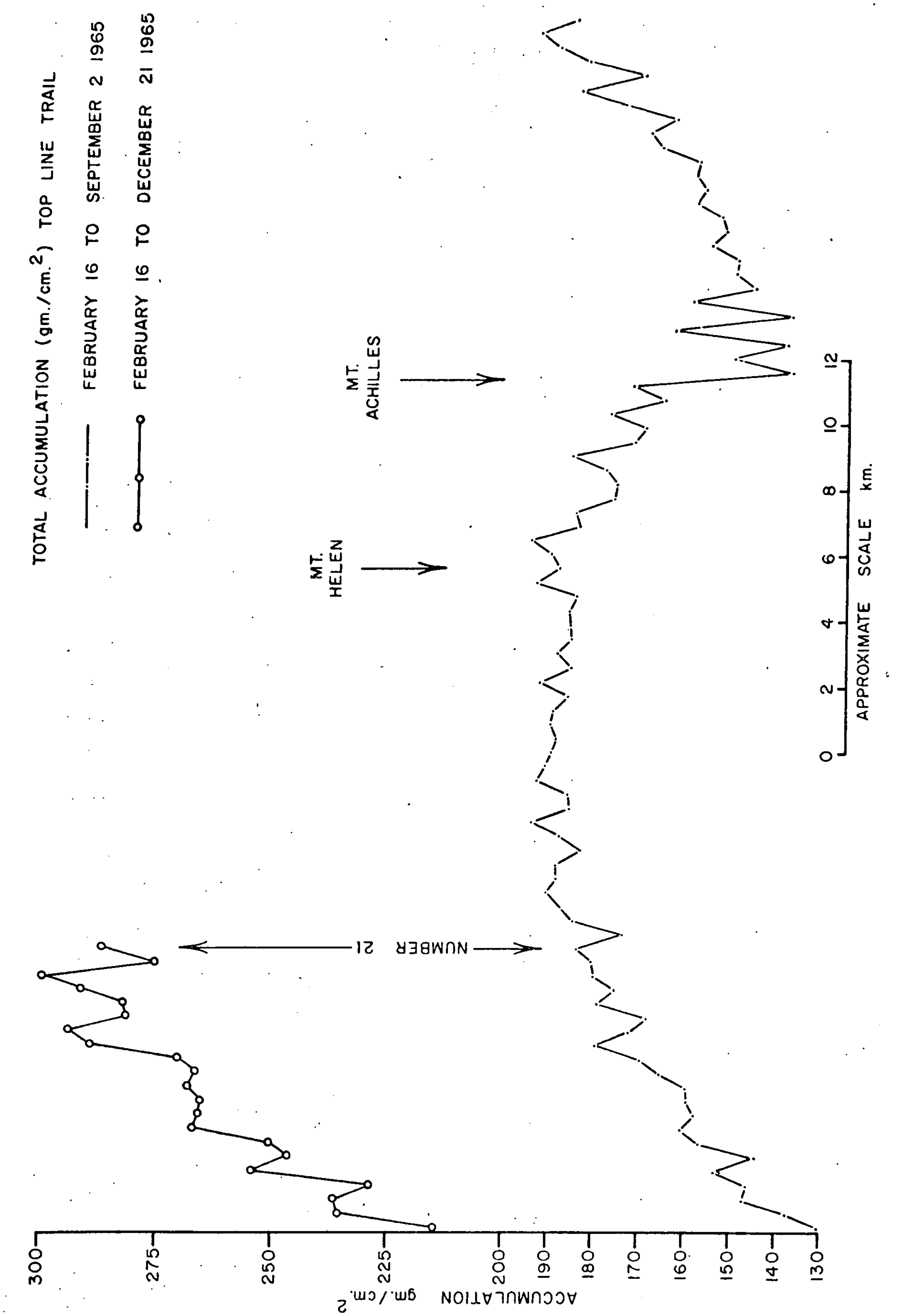
TOP LINE TRAIL

The Top Line Trail discussed here existed for less than one year. The line was established on February 16 1965 with 100 stakes and all 100 were remeasured on April 14. On September 2, 88 were measured before a heavy storm halted work and buried the remaining 12 stakes. Most of the stakes were lost by mid-December and identification of those remaining was impossible. The line was then abandoned.

The 21st stake marked the drainage divide and from this point to the region between Mt. Nestor and Mt. Achilles (see figure 1) the surface was generally quite level. The line then gradually entered a deep depression in the Perrier Bay-Mt. Achilles area then, after turning at stake number 69, climbed again to a level of about 750 m. a.s.l. From the end of the line northward, the surface, as far as could be judged in the field, appeared quite level.

Total accumulation along the line for the 198-day period is shown in Figure 25 and in Table XVII. For comparison, the first "year" total values (308 days) from the first 21 stakes, are also shown in Figure 25.

There is a close correlation between accumulation rate and elev ation with the rate gradually increasing toward the drainage divide and then becoming remarkably constant for about 12 km. across the high, level plateau. The depression in the Perrier Bay area apparently has a marked effect on the accumulation rate which here may be additionally influenced by wind erosion as south and southeasterly winds (not heavily snow-laden) blow through the low saddle between Mt. Helen



TOTAL ACCUMULATION (gm./cm.2) TOP LINE TRAIL
FEBRUARY 16 TO SEPTEMBER 2 1965
FEBRUARY 16 TO DECEMBER 21 1965
MT. ACHILLES
MT. HELEN
NUMBER 21
APPROXIMATE SCALE km.
0
2
4
6
8
10
12
ACCUMULATION gm./cm.2
300
275
250
225
200
190
180
170
160
150
140
130

Fig. 25

and Mt. Achilles. North of Perrier Bay the rate of accumulation again increases with increasing elevation and reaches values similar to those to the south. These results appear to be a departure from conditions on the Fuchs Ice Piedmont on Adelaide Island, where Bryan (1965) reported considerably higher rates of accumulation in the north of the island than in the south, and may reflect the more northerly and possibly more maritime position of Anvers Island. However, the trend of accumulation rates from the southern part of Anvers Island to the most northerly point (the Neumayer line and Top Line Trail) do indicate that the lee slope of the island receives lower precipitation than the higher windward plateau.

STAKE FARMS

Stake farms record accumulation rates at one location rather than at each of several stakes set close together, and detect areal variations of accumulation. This yields an average value for that particular location and provides a more realistic and reliable assessment of accumulation than would a single stake. This form of measurement is mandatory when measuring stations are widely separated (as may be the case on the Antarctic Mainland), or where accumulation rates are very low and where wind action produces a pronounced micro-relief (sastrugi).

Five stake farms were established on the piedmont in early 1965. Their locations are shown in Figure 8 above. The largest was R with 30 stakes and the four others were originally set with 25 stakes each.

TABLE XVII

TOTAL ACCUMULATION (gm/cm^2) TOP LINE TRAIL
February 16 1965 to September 2 1965

Stake	gm/cm^2	Stake	gm/cm^2
1	130.4	46	184.2
2	137.5	47	193.1
3	146.8	48	188.2
4	146.4	49	190.0
5	153.1	50	194.5
6	143.7	51	183.8
7	156.6	52	183.8
8	160.6	53	184.7
9	157.9	54	176.2
10	159.3	55	175.8
11	159.7	56	178.0
12	165.1	57	186.9
13	169.5	58	171.8
14	179.3	59	169.1
15	172.2	60	177.1
16	168.2	61	165.1
17	178.9	62	172.7
18	175.3	63	137.1
19	179.8	64	150.0
20	180.2	65	138.4
21	183.8	66	163.3
		67	137.9
22	173.5	68	159.3
23	184.7	69	145.5
24	187.3	70	150.0
25	190.9	71	149.5
26	188.2	72	155.7
27	188.7	73	152.2
28	182.9	74	153.1
29	187.8	75	158.9
30	194.0	76	156.6
31	185.6	77	158.9
32	186.0	78	158.0
33	193.1	79	166.9
34	191.3	80	169.1
35	189.6	81	163.3
36	188.7	82	174.9
37	190.0	83	184.7
38	189.6	84	170.0
39	186.0	85	182.9
40	192.7	86	189.1
41	185.1	87	193.6
42	188.7	88	185.1
43	185.1		
44	185.6		
45	186.0		

Farms B and C were lost in the storm of early September 1965, while A, D, and R were cannibalised to extinction by late 1965. Available data are plotted as cm. of snow accumulation in Figures 26 and 27.

The variation about the mean is generally constant at about 7 to 10 cm with no relation to the magnitude of the snowfall, which leads to smaller percentage variation for larger snowfall. The value for each stake has been plotted in the same sequence throughout and indicate no systematic areal variation within the arrays. The magnitude of the variation is a fair reflection of the amplitude of the prevailing surface micro-relief, which over the entire piedmont is remarkably small.

The scatter of values cannot be extended as a direct possible or probable error in a single stake measurement because a single measurement is real at the time it is taken. It does indicate that if a single stake had been placed at a different position in the same general area, or if it had been measured at a slightly different time, it would have recorded a different value simply as a result of redisribution of accumulation over the surface. The variation therefore, allows an assessment of the validity of a final value as being representative of the general location. Over most of the piedmont the stakes were set close enough together to overcome the problem raised by large-area variations but the farm results are useful in assessing the worth of the Top Line Trail stakes and those in the Monaco, Therese lines and Skyline where limited data were obtained and measurement stations were more widely spaced.

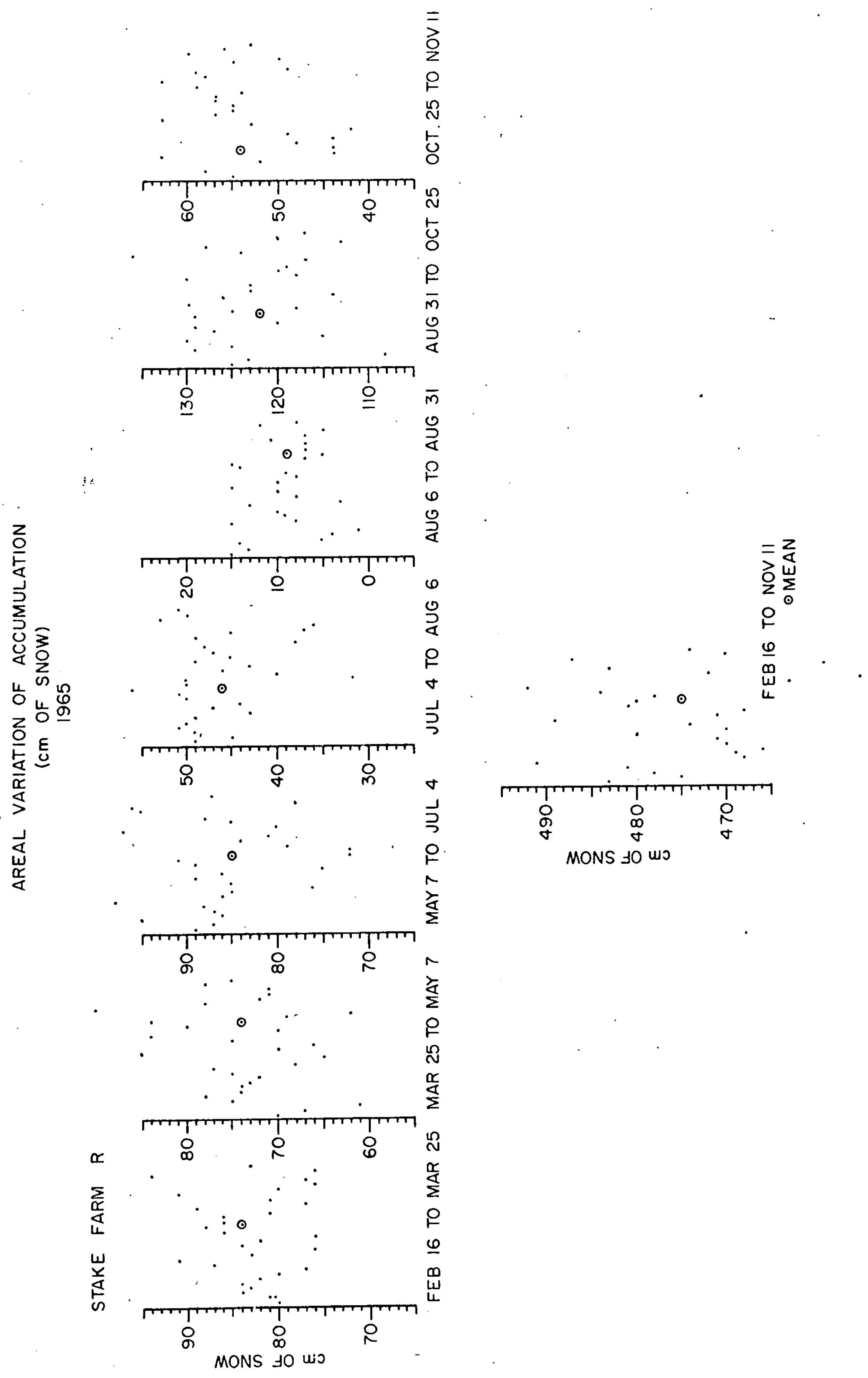

STAKE FARM R
AREAL VARIATION OF ACCUMULATION
(cm OF SNOW)
1965
FEB 16 TO MAR 25
MAR 25 TO MAY 7
MAY 7 TO JUL 4
JUL 4 TO AUG 6
AUG 6 TO AUG 31
AUG 31 TO OCT 25
OCT. 25 TO NOV 11
cm OF SNOW
FEB 16 TO NOV 11
⊙ MEAN
cm OF SNOW

Fig. 26

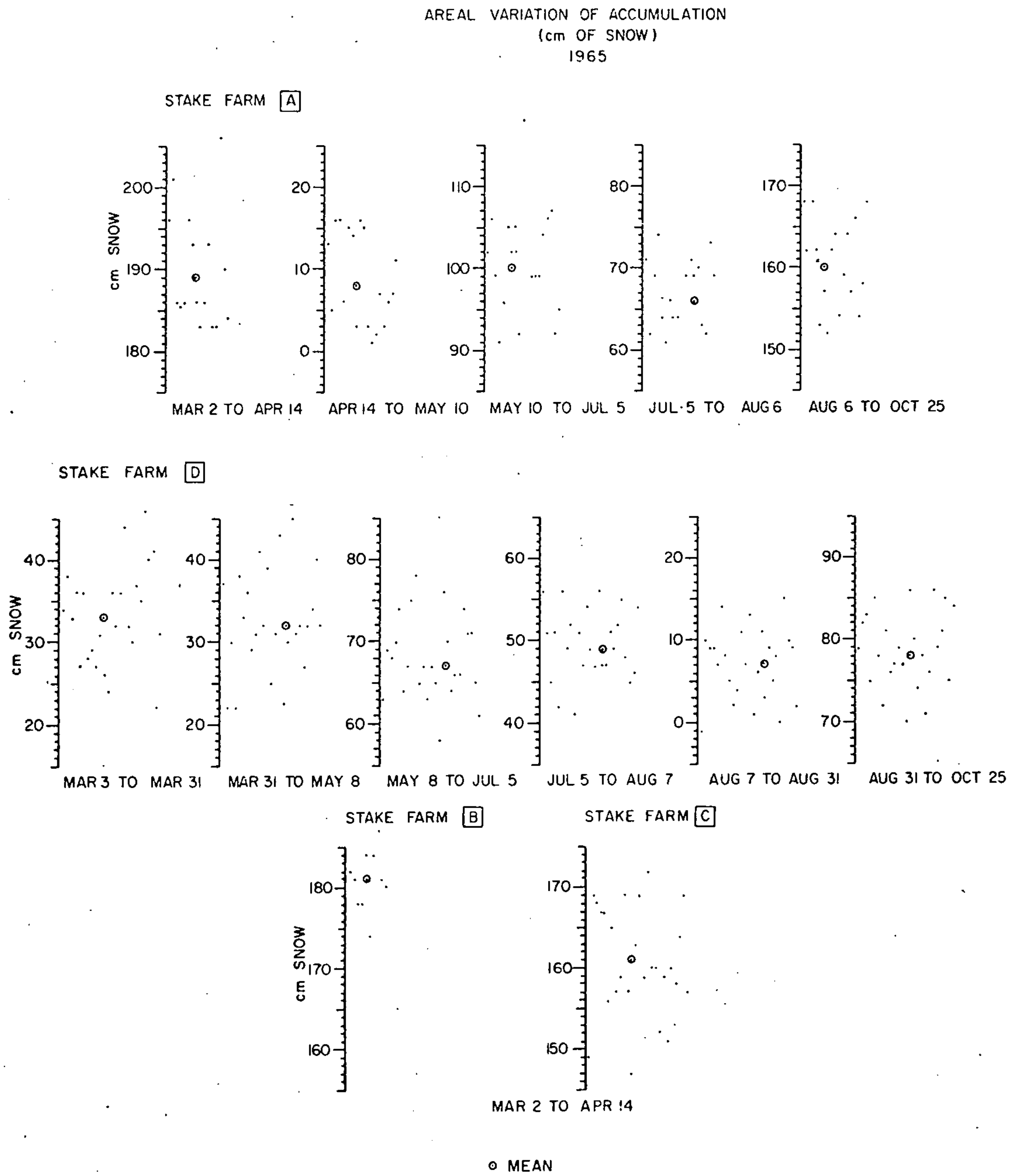

AREAL VARIATION OF ACCUMULATION
(cm OF SNOW)
1965
STAKE FARM A
cm SNOW
MAR 2 TO APR 14
APR 14 TO MAY 10
MAY 10 TO JUL 5
JUL 5 TO AUG 6
AUG 6 TO OCT 25
STAKE FARM D
MAR 3 TO MAR 31
MAR 31 TO MAY 8
MAY 8 TO JUL 5
JUL 5 TO AUG 7
AUG 7 TO AUG 31
AUG 31 TO OCT 25
STAKE FARM B
STAKE FARM C
MAR 2 TO APR 14
MEAN

Fig. 27

At stake farm R the total readings for the period of measurement show a maximum variation of 37 cm. or 17 and 20 cm. about the mean. On a percentage basis of the total accumulation recorded, this reduces to about 4% variation about the mean. On the basis of the two years' data which generally have been evaluated, the variation amounts to about 2%. The average accumulation recorded by stake farm R from February 16 to November 11 1965 was 211.5 g/cm^2. The final stake on the Main line recorded 215.8 g/cm^2 over a similar period (February 3 to December 5 1965) indicating that an individual stake record is representative of accumulation rates in its general vicinity and that few if any anamolous results have been incorporated into the overall survey of accumulation on the ice piedmont.

SUMMARY AND CONCLUSIONS

Total annual accumulation values from February 1965 to January 1968 are available for the two longitudinal lines only. Values for the entire stake network are available from November 1965 to January 1968 and provide a comprehensive survey of annual accumulation rates over the southwestern lobe of the Marr Ice Piedmont.

Figure 28 is a summary of all data obtained from the stake measurements during the latter two years and presents the areal distribution of average annual accumulation for the two-year period. The isoline interval is 10 g/cm^2/yr. In the area north of the dotted line on this map, the isolines are likely to be less accurate than those to the south because they are based on more scattered absolute values.

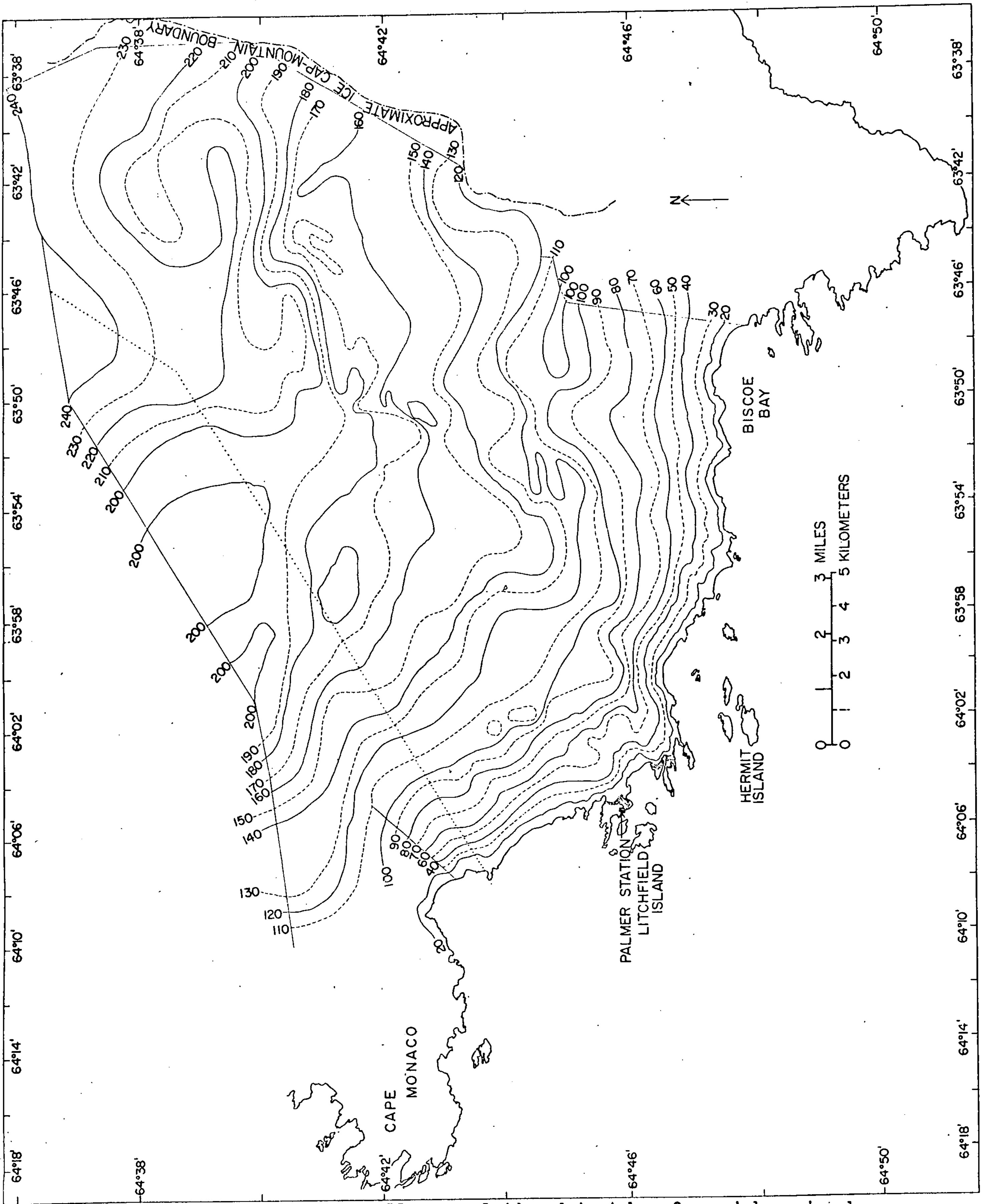


Fig. 28. Summary of all accumulation data taken from stake network between November 1965 and January 1968. Isolines are average annual accumulation in g/cm^2 /yr. Interval 10 g/cm/yr

In the south, the isolines have been drawn on the basis of a generalised curve for each profile, obtained from the three-point running means of the absolute values.

The survey shows that accumulation rates are high, among the highest on record for the Antarctic Peninsula, and are typical of the western coast of the peninsula. There is a very pronounced relationship between elevation and rates of accumulation but over the small latitude range of the piedmont there appears to be little change in the rate of accumulation due to differences in latitude alone. All data, particularly that from the Top Line Trail and the Neumayer line, show a marked influence of aspect on accumulation rates, with less total accumulation being received in the lee slope of the piedmont.

The four sets of curves shown in Figure 21 show an interesting pattern and are strongly suggestive of a precipitation shadow in the eastern part of the study area resulting from the prevailing direction of the principal snow-bearing winds. Along all the lines in question there is a distinctly positive elevation gradient outward from station R, but only along the Top 21 line is the rate of accumulation ~~also~~ markedly positive. The Mountain line exhibits only a slight positive accumulation gradient while the Angela values have virtually no gradient at all. Along the Kersten line, the accumulation gradient is decidedly negative. This pattern can probably be explained in large part by position and aspect relative to the mountains and the snow-bearing winds.

This area of the piedmont has a cirque-like form with the mountains forming an arc around the four profiles. The major snow-bearing winds are from the NE quadrant, particularly from NNE and E (see Appendix I, Volume 2) and approach the profiles from the mountains. Thus the profiles are situated in an increasingly protected position, moving easterly from the Top 21 stakes which are clearly exposed to the winds, to the Kersten line which is most protected. The varying pattern of accumulation in this area appears to result therefore, in the fact that the accumulation lines lie, to a varying degree in a precipitation shadow.

The effects of this shadow may in fact be more widespread, extending westward toward the coast and may explain the variations in accumulation rates along the Ruth, Central and Ingrid lines. The rate of accumulation along the Skyline however, is quite constant with no indication of the precipitation shadow effect. This may be due to increased elevation toward the mountains, as suggested for the Angela line or it may simply be that the Skyline is sufficiently exposed and lies outside the shadow.

During the period of the survey, there was a marked variation of average annual accumulation rates from one year to the next which most probably is related to variations in local and widespread sea ice conditions and to differences in the climatological pattern effecting the Antarctic Peninsula as a whole. The record of past accumulation which is discussed in Chapter 9 below, does not emphasise this variation because of the difficulty of stratigraphic interpretation and because the record itself is relatively short.

At higher elevations on the piedmont where topographic irregularities are small or absent, there are no widespread areal variations of accumulation. The surface of the piedmont is generally smooth, indicating that this is a deposition surface more so than an erosion surface.

CHAPTER FIVE

SURFACE VELOCITY OBSERVATIONS

INTRODUCTION

A network of stations was established on the piedmont for the determination of absolute surface velocities and to investigate the possibility of seasonal variations in surface velocity. Velocity stations generally consisted of 12' x 2" OD aluminum pipe pushed firmly into the snow. 12' x 2" x 4" timbers were used at some locations and in three cases the station was marked by three 12' x 1" bamboo poles tied together. Sighting was always made to the center of the bottom of each stake where it entered the snow surface. Stakes were periodically reset to prevent burial by snow accumulation.

Originally 67 stations were set and surveyed but only 54 were adequately recovered for velocity determination. Station LINDA was lost in the crevasse field and dangerous vehicular access prevented the maintenance and recovery of stations K4, K5, H5 and H6. The survey was incomplete on FLAG and recovery of KAPPA, I and LAMBDA 1 and 2 was not attempted as the first survey of these stations was regarded as weak. Station V was buried during the winter of 1966 and because of the survey technique employed, so prevented the recovery of stations W and Little x.

The nomenclature and location of each velocity station are shown in Figure 29. This network was set to obtain velocity data from a wide range of locations in the study area. Particularly, the Main-Mountain

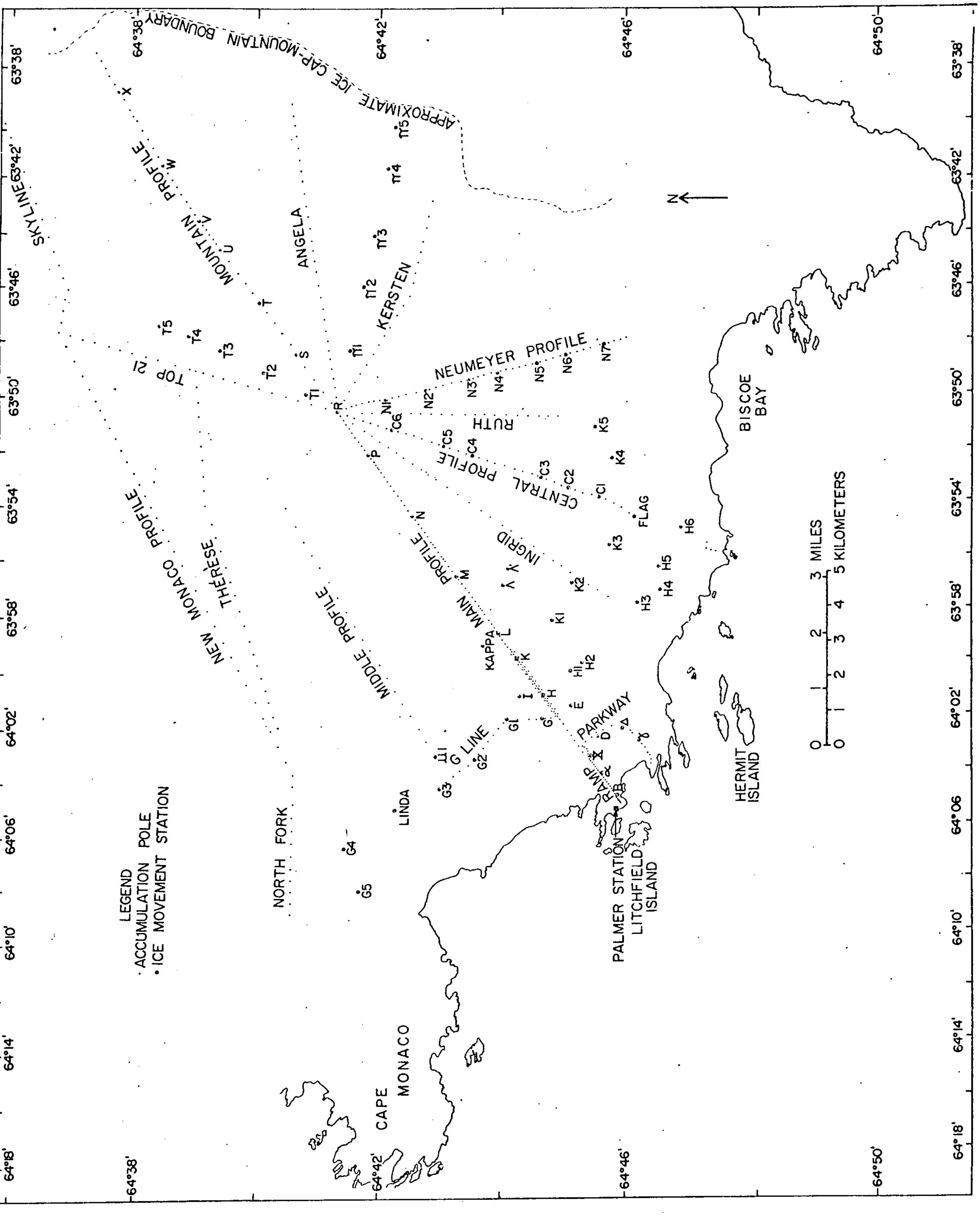


Fig. 29. Location and nomenclature of ice velocity stations.

line, Neumayer line and Central-T line were designated "longitudinal" profiles, while the G and H lines were peripheral, for mass discharge information.

Normally an ice movement survey would be made using triangulation techniques from the ends of fixed, measured baselines preferably parallel to the general direction of ice flow. This system is typical of studies on valley glaciers. This system was attempted for the velocity stations around Arthur Harbor and was tied to a hydrographic triangulation carried out for the US Navy Hydrographic Office during 1965. This triangulation was not extended to the stations further inland because local topographic features made intervisibility of stations difficult to achieve over any appreciable distance and the number of intermediate points would have become too large for occupation by the the small working group. The triangulation would also have extended from a relatively weak setup where the original baseline (Litchfield Island to Norsel Point) was short, perpendicular to the general ice flow direction and slightly offset from the main axis of the survey.

The only bedrock in the area on which baselines could be established are the small islands around the periphery of the piedmont. Any baseline established on these islands would be perpendicular to the ice flow direction and the desired accuracy could not be obtained. For the remaining stations therefore, other methods were employed.

The inland velocity stations were established using a system of distance and angle measurement in open traverse lines. Distances were measured with a Tellurometer system and the angle between the lines of measurement was determined with a Wild T2 theodolite.

This system was an adaptation of techniques which had proven successful in previous glacier surveys (Hoffman 1960; Hoffman and others 1964) and the main body of this text will not dwell upon the technical details of the system but will restrict itself to the discussion of the results. The technicalities together with pertinent data tabulations are presented in Appendix II. Coordinates and velocity values are shown in Table XVIII.

SURFACE VELOCITY DATA

The pattern of surface velocity, shown in Figure 30, exhibits remarkable differences from one part of the study area to another but there is a distinct sensibility to the overall distribution and this clearly reflects the subglacial topography (Chapter 3). In magnitude the velocities fall into two distinct ranges.

1) The Pi line, Neumayer line and Central line have velocities which are closely comparable with those of some valley glaciers. The velocity at stations G2, G3 and Mu1 is of similar magnitude.

2) The remaining stations all have velocities which are lower than might be expected under these conditions of ice thickness, surface slope and activity of regime.

The subglacial topography as revealed by Dewart's gravity survey (figures 13, 14 & 16) can be deduced from the velocity vectors and there is clearly a high degree if ice channeling which greatly enhances the rate of movement in certain areas. The stations along the Neumayer and Central lines apparently lie somewhat on the walls of the submerged valley and are creating a convergent flow resulting in

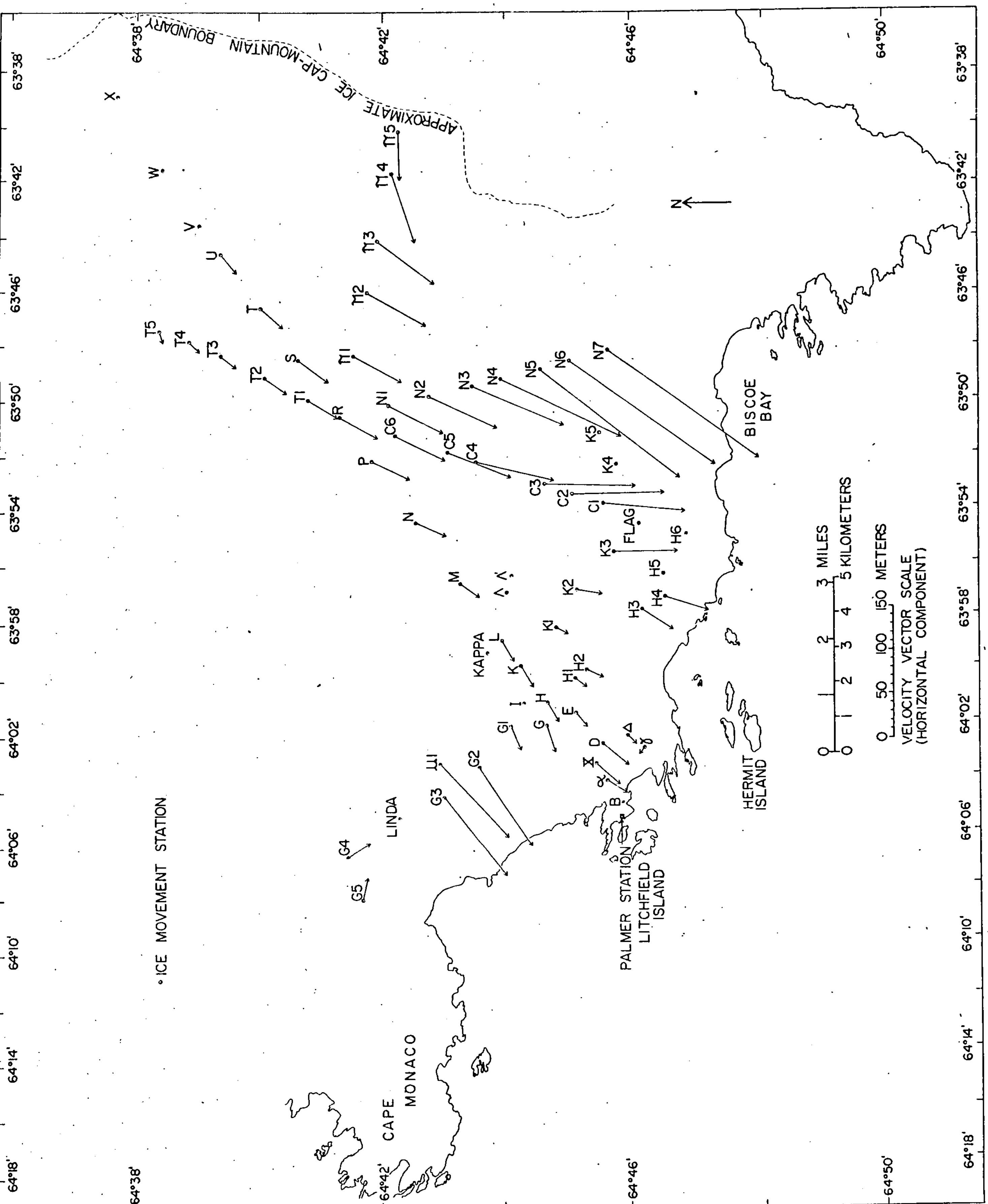


Fig. 30. Horizontal component of surface velocity vectors. Vector scale in meters per year.

TABLE XVIII

COORDINATES OF ICE VELOCITY STATIONS

	FIRST EPOCH			SECOND EPOCH			MOVEMENT(m/y)		
Station	X	Y	Z	X	Y	Z	Horizontal	Direction	Vertical
E	14125.69	10643.03	284.11	14106.28	10636.55	281.73	20.46	251.6	-1.38
H	14715.32	11325.99	312.97	14689.61	11322.44	311.64	25.96	262.1	-1.33
K	16010.18	11657.99	369.99	15982.22	11653.90	368.75	28.26	261.7	-1.24
L	16826.77	11863.07	394.64	16800.07	11859.51	396.22	26.94	262.4	-1.59
M	18867.96	12374.87	474.35	18844.71	12361.13	475.53	27.01	239.4	-1.18
N	21099.72	12933.27	540.72	21070.63	12906.62	541.44	39.45	227.5	-0.72
P	23266.91	13471.13	593.54	23229.45	13439.34	594.21	49.13	229.7	-0.66
R	24797.82	13878.85	631.11	24757.95	13848.86	632.71	49.89	233.0	-1.60
S	26885.59	14419.04	676.47	26847.71	14397.93	677.12	43.37	240.9	-0.65
T	28730.70	14897.16	716.39	28698.49	14883.30	717.20	35.06	246.7	-0.80
U	30645.75	15392.79	750.42	30617.96	15384.85	751.66	28.90	254.0	-1.23
K1	16433.26	10174.05	360.20	16422.22	10165.04	358.60	14.25	230.7	-1.60
K2	17385.50	9171.54	349.83	17368.45	9145.15	347.55	31.42	212.9	-2.28
K3	17969.34	7641 98	295.55	17939.03	7573.73	291.81	74.68	203.9	-3.73
C1	19371.40	7580.33	305.02	19323.35	7495.69	300.11	97.33	209.6	-4.91
C2	20017.85	8320.86	364.97	19974.37	8220.93	357.01	108.98	203.5	-7.96

Continues

TABLE XVIII (cont)

Station	FIRST EPOCH X	Y	Z	SECOND EPOCH X	Y	Z	MOVEMENT (m/y) Horizontal	Direction	Vertical
C3	20662.56	9085.25	422.91	20616.14	8987.18	414.13	108.51	205.3	-8.78
C4	22016.66	10658.05	501.52	21958.80	10582.37	493.11	95.27	217.4	-8.41
C5	22650.03	11393.14	536.29	22592.80	11335.61	528.60	81.15	224.9	-7.70
C6	23693.18	12605.28	587.72	23643.10	12562.98	580.54	65.55	229.8	-7.17
RC*	24794.95	13882.73	633.97	24753.36	13856.06	627.48	49.40	237.3	-6.48
G	14018.83	11782.57	279.04	13987.57	11784.89	277.24	31.34	274.2	-1.81
G1	14471.44	12724.69	304.00	14442.07	12725.41	302.42	29.37	271.4	-1.58
G2	13871.99	14490.07	264.01	13764.00	14469.38	260.50	109.95	259.2	-3.50
G3	13419.71	15497.34	273.94	13305.78	15465.35	268.08	118.34	254.3	-5.86
G4	12998.58	18942.88	301.69	13003.32	18911.07	300.32	32.16	171.5	-1.37
G5	11722.71	18973.42	281.58	11743.61	18958.39	282.16	25.74	125.7	-0.58
H1	15002.65	10264.49	314.90	14988.17	10256.03	253.23	16.77	239.7	-61.68
H2	15126.53	9867.42	306.06	15111.18	9853.28	244.87	20.87	227.3	-61.19
H3	16058.95	7597.84	207.52	16023.74	7573.82	142.85	42.63	235.7	-64.67
H4	16119.04	6852.08	197.86	16084.21	6812.35	136.88	52.83	221.2	-60.98
N1	24577.93	12387.50	595.97	24522.74	12343.30	595.43	70.70	230.0	-0.54
N2	24370.30	11213.32	566.90	24303.05	11155.14	564.85	88.92	228.1	-2.05
N3	24135.80	9906.63	527.97	24048.69	9826.70	524.46	118.22	226.7	-3.51
N4	23953.88	8945.36	487.49	23834.45	8844.01	482.41	156.64	229.1	-5.08
N5	23737.79	7823.36	454.80	23557.06	7721.73	436.25	207.35	240.2	-18.55

* RC is station R computed via the K and C traverse

Continues

TABLE XVIII (cont)

Station	FIRST EPOCH X	Y	Z	SECOND EPOCH X	Y	Z	MOVEMENT (m/y) Horizontal	Direction	Vertical
N6	23599.16	6940.41	416.00	23418.65	6832.16	355.93	210.47	238.6	-60.07
N7	23399.38	5772.43	360.65	23208.62	5665.38	302.04	218.74	240.2	-58.61
Pi1	26361.88	12831.86	644.80	26309.64	12794.68	642.90	64.13	233.1	-1.90
Pi2	27996.12	11721.98	661.63	27931.08	11676.93	655.54	79.13	234.1	-6.09
Pi3	29305.71	10856.21	672.09	29231.91	10817.13	671.58	83.50	240.9	-0.51
Pi4	30972.35	9718.16	700.54	30889.10	9727.64	700.14	83.79	275.2	-0.40
Pi5	32027.05	9043.45	737.29	31976.64	9066.64	737.92	55.49	293.0	-0.63
T1	25586.03	14516.24	656.97	25550.05	14493.36	655.46	42.65	235.2	-1.51
T2	26793.24	15540.10	691.42	26764.32	15525.31	690.32	32.49	239.8	-1.09
T3	27890.38	16449.75	717.48	27868.66	16440.65	716.04	23.55	242.9	-1.44
T4	28758.10	17197.24	736.60	28741.34	17193.13	735.72	17.25	249.9	-0.88
T5	29432.39	17836.99	744.76	29418.84	17840.04	742.48	13.89	275.0	-2.29
Mul	14334.86	15338.81	314.33	14225.88	15297.42	307.54	116.58	249.2	-6.79

a rapid velocity increase down-valley. The stations at Pi2 and Pi3 lie over the local valley bottom while Pi4 and Pi5 are being channeled into the valley head. The gradual southward swing of the vectors along the T line and the Mountain line emphasise this channeling effect and points up the significance of the configuration of the piedmont-mountain boundary. The slightly curved configuration of the line through Mt. Rennie, Mt. Agamemnon and "Big Mountain", together with the steeply rising subglacial bedrock, is strongly suggestive of a cirque, which is forming the initial accumulation area for an ice stream or submerged valley glacier.

The direction of movement of stations M, N, P, K2 and K3 closely follow the direction of maximum surface slope and suggest that the "platform" in this area is gently dipping southeastwards. The vectors at stations C6 and C5, which indicate a movement direction westward across the main axis of the Central line, clearly reflect the small secondary valley outlined by the 150 meter subglacial contour on Dewart's map (figure 13). The extension of this feature northeast of station R is reflected by the ice movement at stations S, T1, R and P. The foot of this valley, trending northwestwards leads to the sudden change in direction of movement of the stations beginning at C3, and for the more southerly direction at K3 compared with K2. In this part of the study area there can therefore, be discerned two streams of flow; one is bringing ice from the end of the Mountain line, through the eastern part of the Pi line and then through the lower parts of the Neumayer line into the deep valley basin in the vicinity of FLAG and H6. The other stream brings ice from the middle and lower portions of the T line through stations R, P, C5, C6 and N1 with station N lying on its western edge. This stream then swings southwards into the valley basin, with station K3 on its western edge, to meet the easterly ice stream.

The marked change in the direction of movement along the Main line at station L is also explained by the bedrock configuration. Station L lies a little below the 50 meter contour on Dewart's map (figure 13). Here there is a gently curving stream passing through stations L, K and G with station H on its eastern edge. In this area the "platform" is gently dipping to the northeast and forms a shoulder to the valley foot around Wylie Bay. The high velocities at the three stations G2, G3 and Mu1 result from the channeling effect of the deeper parts of this valley and form part of the heavily crevassed ice stream seen in the aerial photograph (figure 17). No surface slope measurements were accomplished at stations G4 and G5 though the vectors do appear to be following the general direction of maximum surface slope as it is recorded by this author. A very sharply southward curving ice stream can be anticipated in this area.

Stations E, H1 and H2 together with K1 lie over the central part of the "platform" and reflect another ice-flow system. Station E, diverging slightly away from the axis of the Main line, obviously forms part of the main stream which discharges through Arthur Harbor. The ice flowing through stations H1, H2 and K1 forms a subsidiary stream and probably discharges into the bay to the east of the Arthur

Harbor promontories. K2 more probably forms part of the larger ice stream flowing into Biscoe Bay.

Discussion

There is then a highly complex configuration to the surface velocity field which is controlled almost wholly by the configuration of the subglacial topography. The vector distribution explains the variation in magnitude of the measured surface movement and permits initial speculation on the mechanisms of flow operating in the ice piedmont.

In an idealised situation, as might be found on a model valley glacier, whose bedrock is formed by a gently sloping surface, the velocity would gradually increase from the head of the accumulation zone until it reached a maximum at the equilibrium line. This would result from the gradual increase of mass which has to pass through the cross section in unit time. Below the equilibrium line, where mass is gradually lost by ablation processes, the surface velocity would gradually decrease and be at or close to zero at the terminus. Variations of velocity along the longitudinal profile of such a glacier would be caused by changes in the surface slope or increase or decrease in ice thickness.

This situation however, as can be seen from Figure 31, is approached only along the Tline-Neumayer line and T line-Central line. The lowest rate of movement recorded in the study area, 13.9 meters per year at T5, gradually increases through all the T stations through R and then into the Central line stations. This acceleration is main-

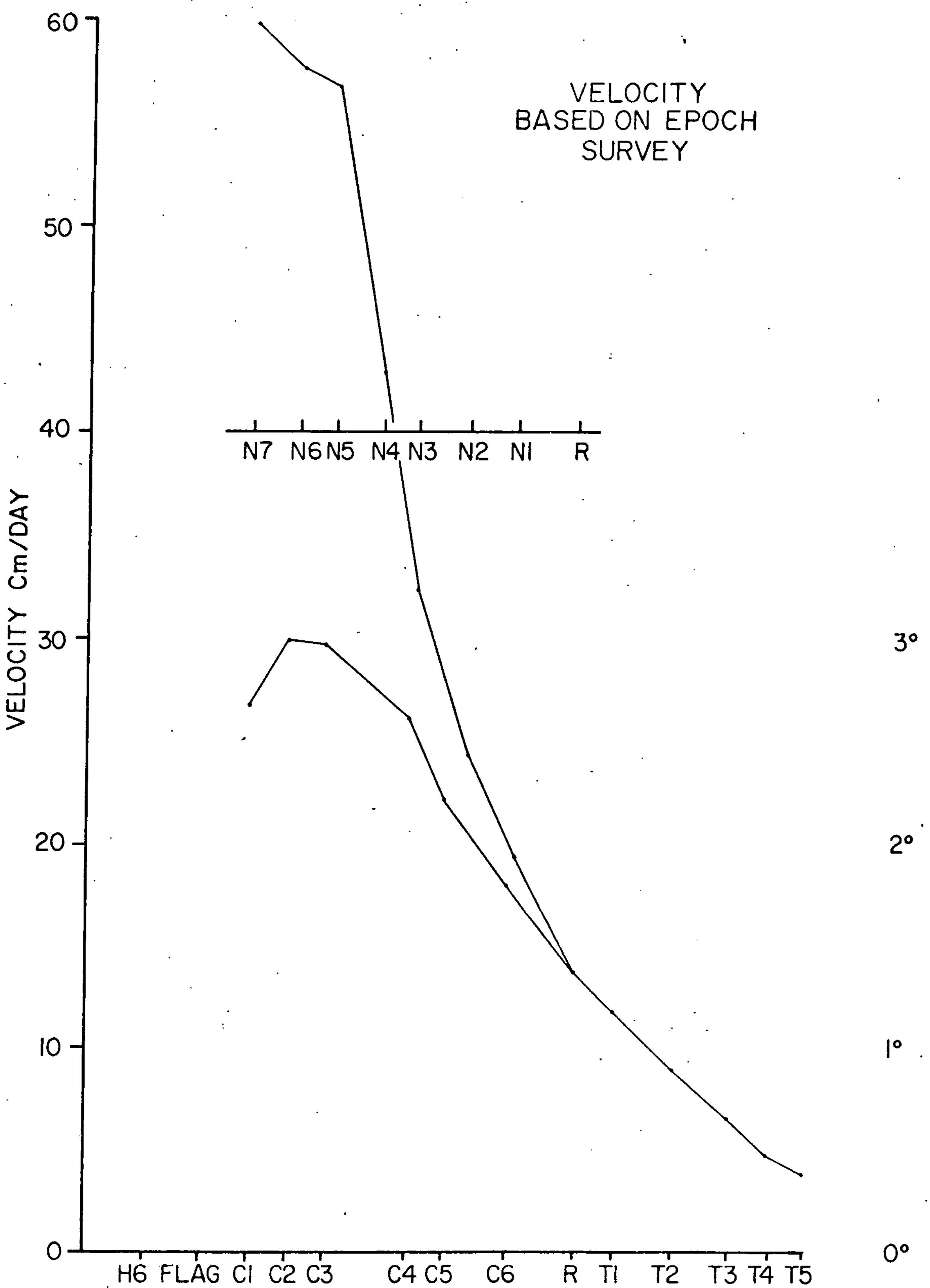


Fig. 31. Surface velocity profiles, T line-Neumayer line and T line-Central line.

tained until it drops off at C1.

The sudden deceleration at C1 can be related to the increase in ice thickness over the deep basin below this station, FLAG and H6. A similar pattern is evident on the Neumayer line. Almost all the remaining stations in the study area lie on profiles which cut, at varying angles, the principal flow lines. Their velocities therefore, do not reflect the idealised situation, as can be seen from Figure 32. Station E for example, from the point of view of its position within the study area, lies near the terminus of the ice piedmont but well above the equilibrium line. Its rate of movement is a mere 20 meters per year. This is compared with some inland stations, for example, station N whose velocity is approximately twice this value. Station E is in fact at or near to the head of its own accumulation area; it is not receiving, and therefore does not have to handle a massive volume of ice from the hinterland. This has already been discharged in other directions through stations L and K. The accumulation passing through station M does not reach to E because it is swept southwards and probably flows closer to stations K1 and K2 and discharges eventually into the bay to the east of Arthur Harbor.

The variation of velocity along the main line as well as the relatively low values recorded, probably also results from divergent flow upstream. The Main line, as can be seen from the velocity vector pattern (figure 30), is tangential to the major direction of ice streaming. There are several shallow ridge-and-trough features (though these are not reflected in the bedrock map prepared by Dewart) in the large area occupied by the Middle profile, and the valley whose

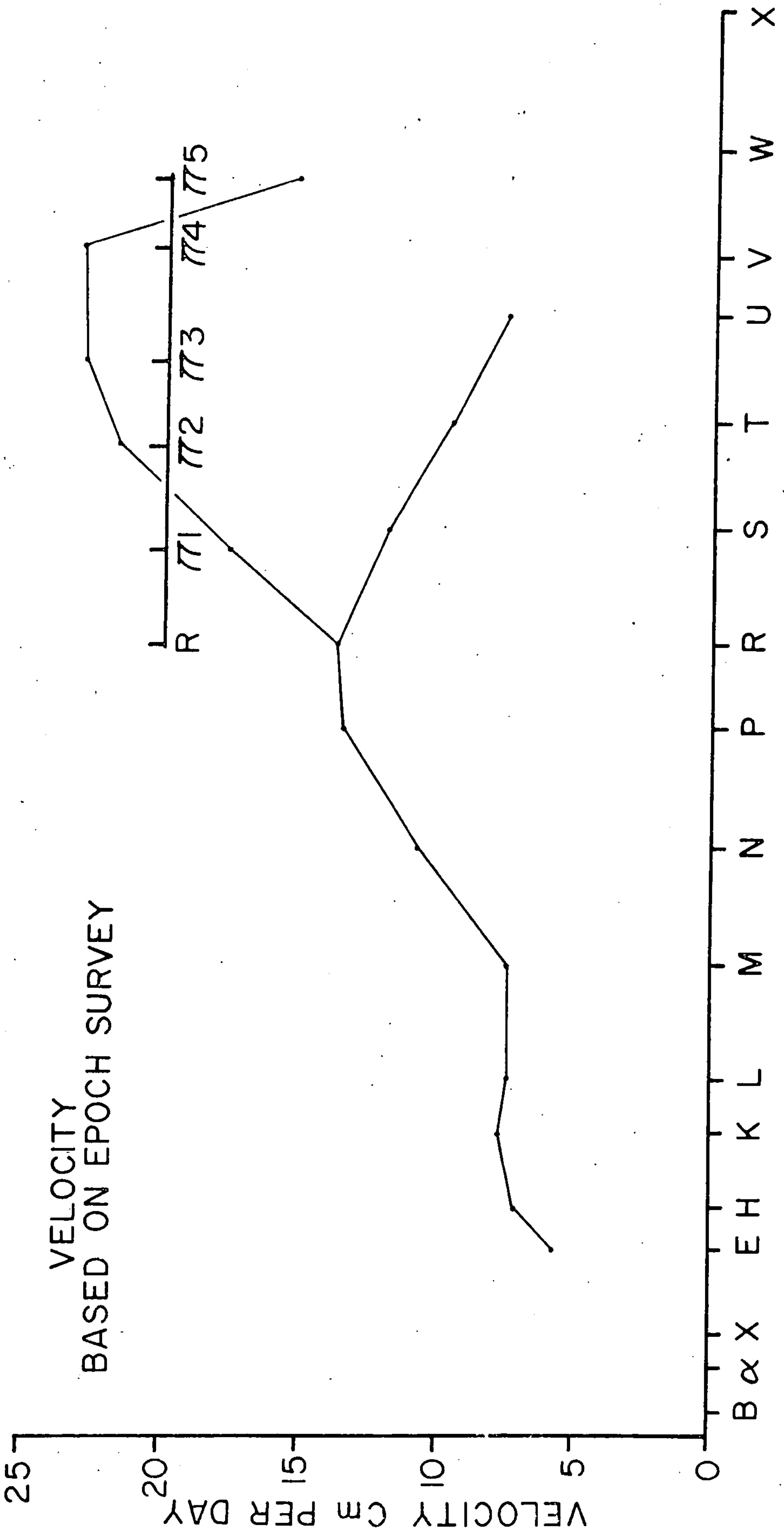


Fig. 32. Surface velocity profiles, Main-Mountain line and Pi line.

foot lies in the region of stations G2 and G3, may extend futher into this area. Unfortunately, no velocity stations were established in this part of the study area, but from the relatively low velocities along the Main line, it seems probable that the ice in the vicinity of the Middle profile is streaming towards the Wylie Bay area rather than crossing the Main line. Compare for instance, the velocity at N (39.4 m./yr) and C5 (81.1 m./yr) or C6 (65.5 m./yr), which lie at approximately the same distance from the ice divide of the Monaco-Skyline profiles. It seems obvious that the ice accumulating to the north of station N does not flow through this station. On the other hand, the ice in the vicinity of the T-line does pass through C6 and C5 and greatly enhances the the velocity of those stations.

ARTHUR HARBOR SURVEY

Triangulation was used to obtain velocity data from the ice immediately behind Arthur Harbor but because the terrain prevented the establishment of ideal baselines, the network setup was weak. Consequently, to strengthen the survey, distances were also measured to some of the ice stations. Figure 33 is a schematic of the network.

Four surveys were made; April-May 1965, September 1965, May 1966 and December 1966-January 1967. As with the main traverse-surveys, the data obtained were reduced to daily rates, then extrapolated to common epochs. Three epochs were set; May 17 1965, May 2 1966 and January 6 1967, each representing approximately the mid-point of survey activity. Table XIX gives station coordinates at each epoch, total movement and

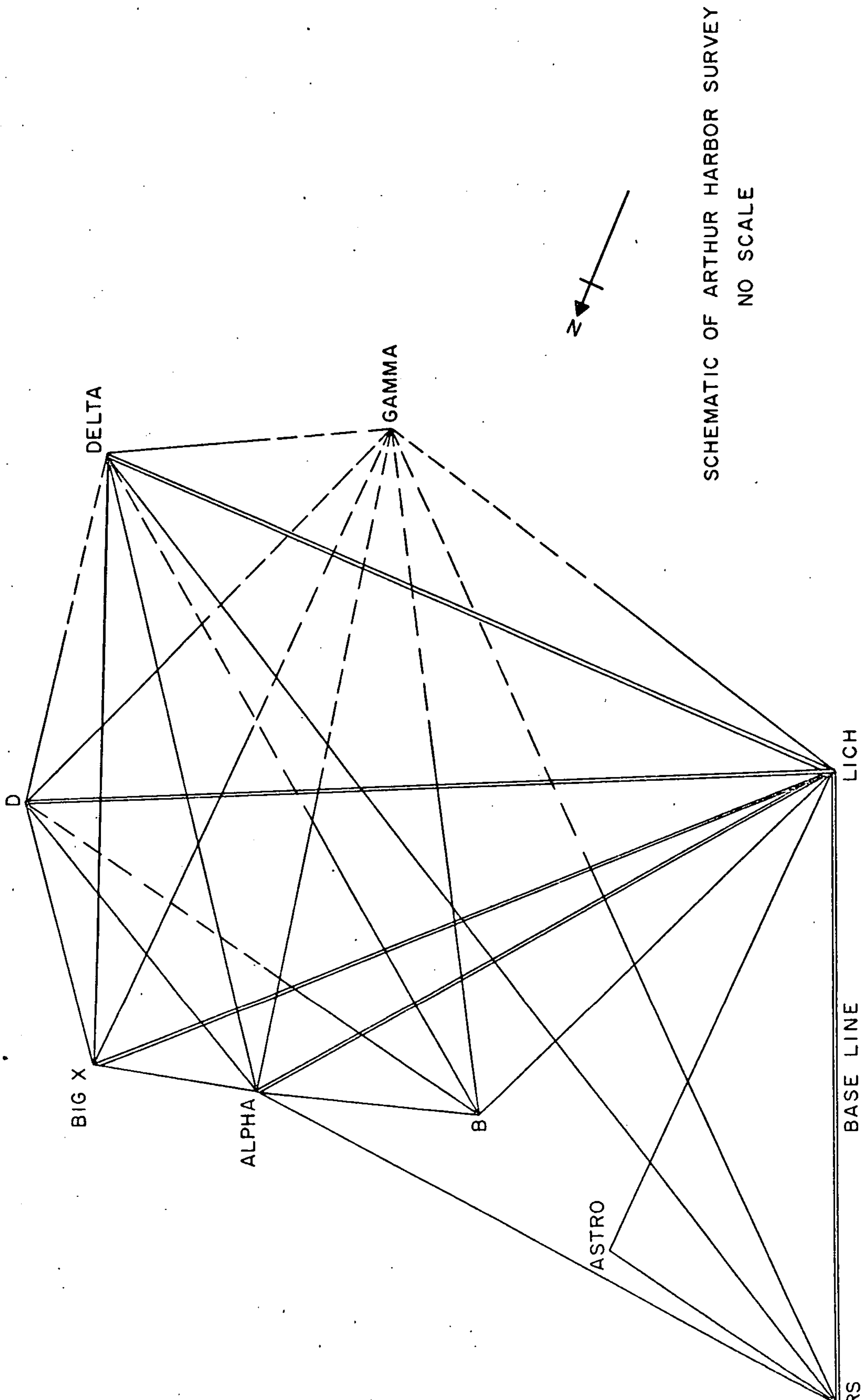

SCHEMATIC OF ARTHUR HARBOR SURVEY
NO SCALE
N
GAMMA
DELTA
D
LICH
BIG X
ALPHA
B
BASE LINE
ASTRO
NORS

Fig. 33.

extrapolated average annual velocity between each epoch and the average annual velocity based on the total epoch time of 599 days. Figure 34 shows the vectors plotted from these data.

With the exception of Gamma, the slight change in azimuth in these vectors can probably be accounted for by field errors in the survey. At Gamma, survey error may also account for the azimuth change but at this station, azimuth change could be real as the station flows into the mainstream of ice in the harbor.

All five stations show channeling of the ice into Arthur Harbor, probably reflecting a subglacial basin between the Norsel and Bonapart Point promontories. These "Parkway" stations were emplaced immediately inland of the crevasse field and it is probable that a sharp break of slope in the subglacial surface occurs in this area and that much of the basin floor is below sea level. The crevassing occurs at the break of slope.

Three of the stations, Alpha, Big X and D, have similar velocities which are generally consistant with the velocity pattern immediately inland. At Alpha and Big X the extrapolated annual velocities show little variation, suggesting consistant movement throughout. It is known that a crevasse formed immediately inland of station D and seaward of Big X sometime between May and September 1965. Part of the movement of D therefore, was probably relatively rapid with the development of the crevasse and may account for the slightly higher velocity between epochs 2 and 3. The very low velocity of Delta and the relatively high velocity of Gamma are not readily explainable and

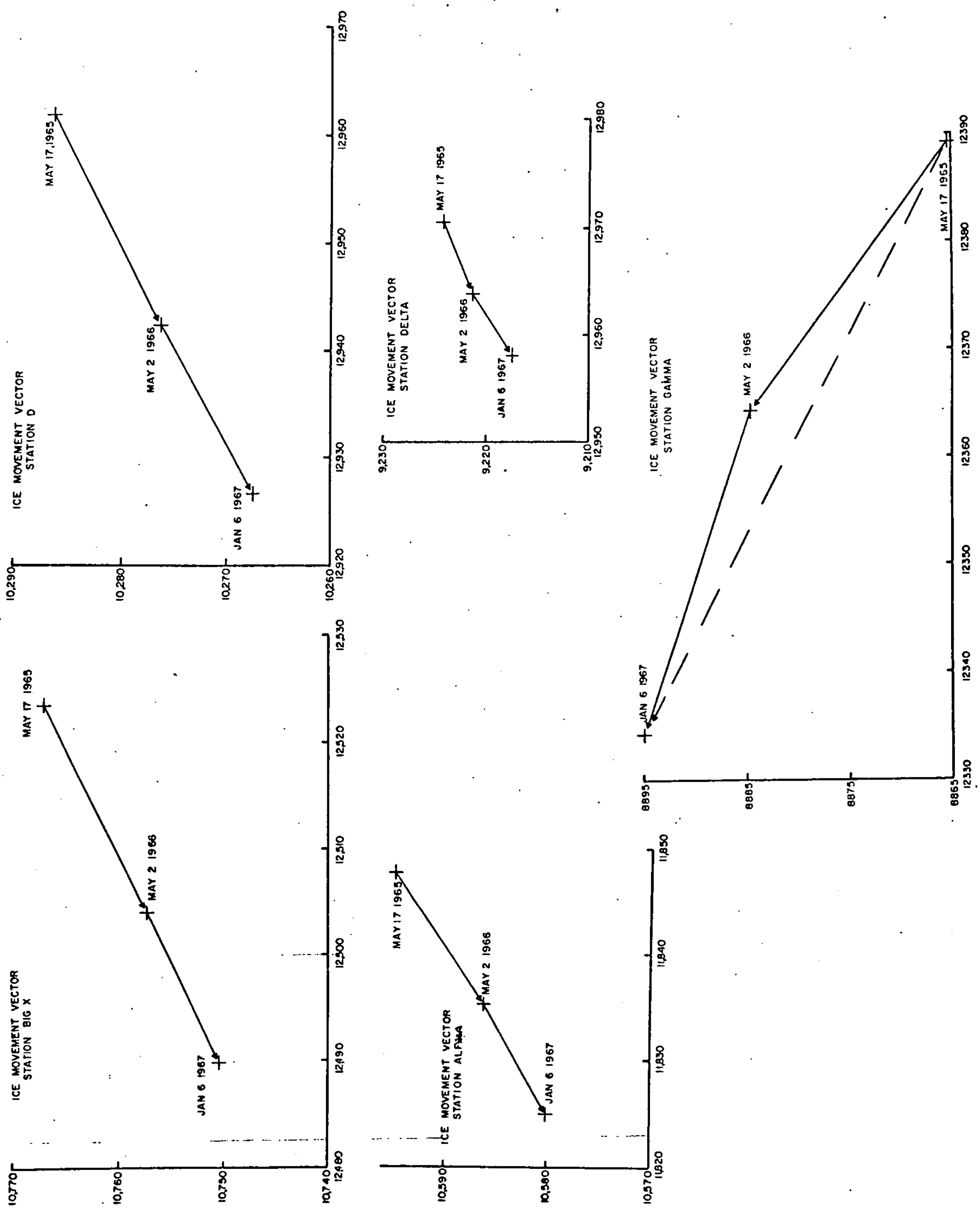


Fig. 34. Horizontal component of surface velocity vectors, Arthur Harbor stations. Coordinate scale in meters.

TABLE XIX

STATION COORDINATES
HARBOR SURVEY

May 17 1965

Station	X	Y	Z	TOT MOV	AVG MOV (m/yr)	
Alpha	11847.666	10595.033	129.558	-	-	
Big X	12523.623	10766.824	187.643	-	-	
D	12961.902	10286.527	191.142	-	-	
Delta	12970.485	9224.223	150.830	-	-	
Gamma	12389.209	8865.369	112.127	-	-	
May 2 1966						
Alpha	11835.129	10586.343	127.190	17.2 m	17.9	
Big X	12504.129	10757.159	185.178	21.6 m	22.5	
D	12942.300	10276.294	187.760	22.1 m	23.0	
Delta	12963.771	9221.274	148.473	7.3 m	7.6	
Gamma	12364.415	8884.522	108.370	31.3 m	32.6	
Jan 6 1967						AVG MOV (m/y) 599 Days
Alpha	11824.946	10580.142	126.651	12.1 m	17.7	17.8
Big X	12489.830	10750.473	184.044	15.9 m	23.3	22.8
D	12926.736	10267.400	186.512	18.0 m	26.4	24.4
Delta	12958.060	9217.434	146.956	6.9 m	10.1	8.6
Gamma	12334.338	8894.978	116.195	31.9 m	46.7	38.5

crevassing does not account for the differences. No crevasses are known to have formed near Delta. Its very low velocity may simply reflect insubstantial mass from its hinterland or, unlike the other stations, the subglacial surface there, is above sea level. Though Gamma was never occupied as a velocity station, it was frequently visited for the accumulation survey and no crevasses existed in the vicinity. The overall velocity of Gamma was much greater than the other stations and showed a pronounced increase between epochs 2 and 3. A markedly sub-sea-level bedrock, resulting in "lubrication" at the base, may account for the greater overall velocity at that station and the increased velocity during 1966 may have resulted from the station entering the main "stream" of ice flowing into the harbor, as suggested by the azimuth change in the vector.

Conclusions

The only ice entering Arthur Harbor is probably that from the vicinity of stations E and H1. All other ice from inland "shears" past the ice of the two promontories. The velocity field within the Harbor ice is probably complex and erratic as a result of crevassing. More information on the subglacial topography and frequent, short-term movement surveys are needed before this velocity pattern can be completely elucidated.

SEASONAL VARIATION OF VELOCITY

Introduction

There are several reports of seasonal and short-term variations in glacier movement and in some cases the results have been regarded as indicative of geophysical characteristics of the ice mass. In an attempt to detect any variations in the movement of the piedmont, stations E through N were surveyed on four occasions and stations P and R on three occasions. It would have been desireable to include more stations in these extra surveys but time and conditions prevented this.

The Surveys

The first survey of stations E through N was completed during April and May 1965. Distances were measured to P and R but the angle measurement was not accomplished. The second survey, which did include stations P and R was made in late September 1965, while surveys three and four were completed in May-June 1966 and October 1966 respectively, with the exception of tying E to datum (Litchfield Island), which was not possible (due to sea ice in the harbor) until January 1967. The movement between surveys 1 and 2 and surveys 3 and 4 is therefore regarded as winter movement while that between surveys 2 and 3 is summer movement.

To obtain the total movement and daily rate of movement between the surveys, three different time spreads are available. These are the time between the individual measurements of the traverse angle (called β time), the time between distance measurements (called LF time) and the average of these two times (AVG time). Calculation of coordinates for each survey (Table XX) and the movement values derived (Table XXI) are based on the average time spread. The remaining time spreads for each survey and the respective velocities derived are included in Table XXI. Figure 35 shows the total movement between surveys (based on average time) and Figure 36 is a plot of the velocities from all the time spreads.

Results

The results are not particularly convincing. From the plot of daily movement rates (figure 36), it appears that at E and H as well as at station L, winter movement is slightly in excess of summer movement. The magnitude is so small however, a matter of a few millimeters a day, that it can probably be discounted. The higher velocity at K is anamalous but clearly evident. The rest of the data, moving up-profile inland, show a steady decline in the rate of movement throughout the surveys.

The velocity of the other stations relative to E is plotted in Figure 37 and numerical values are tabulated in Table XXII. An almost identical pattern emerges. In all cases the slightly higher velocity of K relative to H and L is evident. The reason for this is not clear but it may reflect the position of K relative to the ice flow lines or it may be the result of a localised difference in the conditions at the base. The gravity survey did not reveal sufficient bedrock detail to explain the movement at station K.

However, the nature of the survey technique probably renders this study invalid. The derived velocities depend on two types of field measurement; the distances and angles in the traverse. Rarely was it possible to measure these at the same time or complete the traverse in a single day (which is true of course, for the epoch surveys). It is possible therefore, that in the time between the measurement of the two factors, a significant event ocurred, rendering the averaging of the time spread invalid.

Thus it is concluded that these observations provide no firm evidence of seasonal variation of surface velocity.

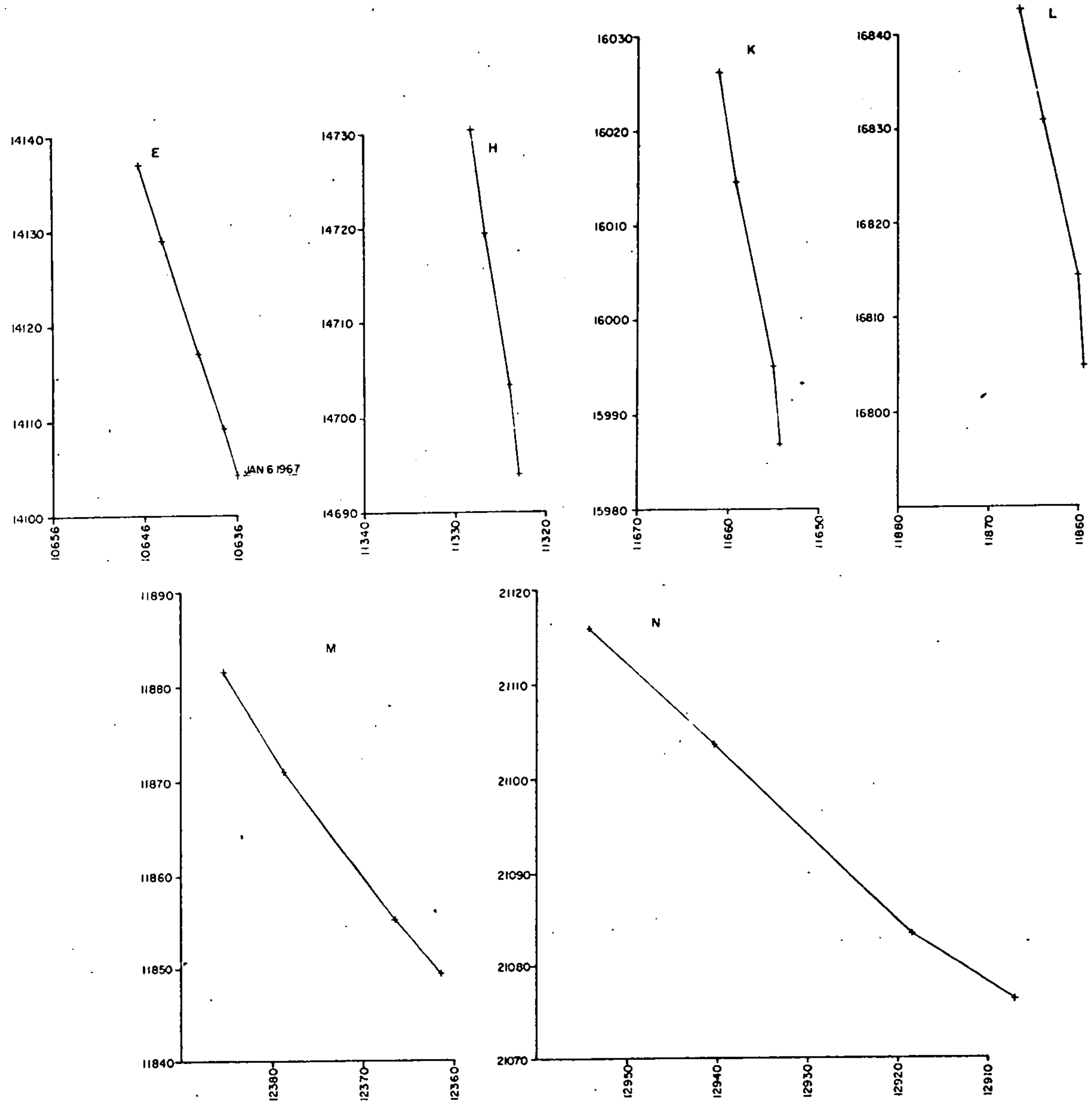


Fig. 35. Horizontal component of surface velocity vectors based on epoch and intermediate surveys. Coordinate scale in meters.

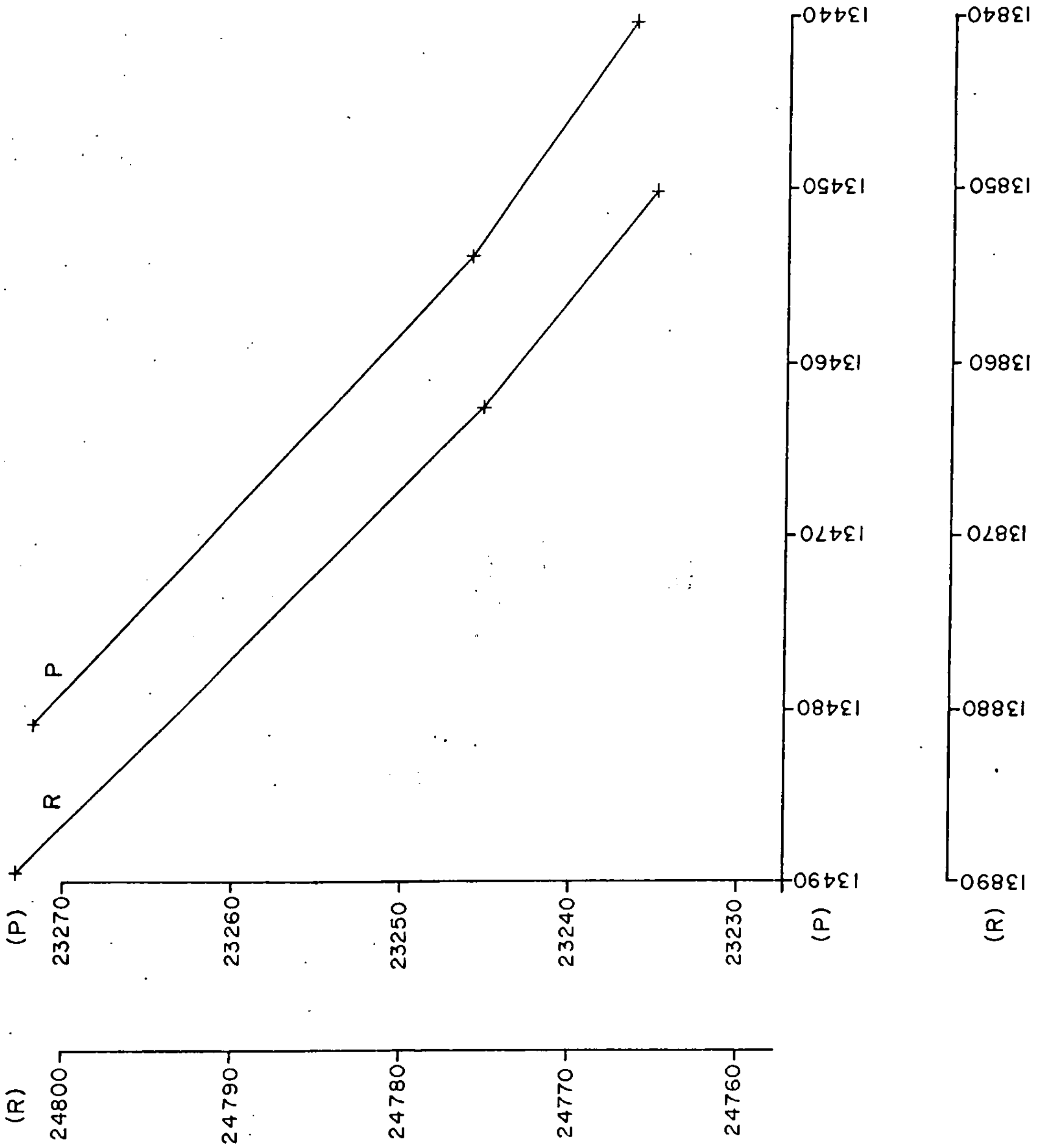


Fig. 35 (continued).

TABLE XX

COORDINATES FOR SHORT-TERM SURVEYS
STATIONS E-R

	E			M	
Survey	X	Y	Survey	X	Y
1	14137.0	10646.6	1	18881.6	12385.4
2	14129.0	10644.1	2	18871.1	12378.8
3	14117.0	10640.1	3	18857.5	12366.9
4 (Jan)	14104.2	10636.0	4	18849.2	12361.4
4 (Oct)	14109.2	10637.5			
	H			N	
1	14730.4	11328.3	1	21115.9	12954.2
2	14719.5	11326.7	2	21103.6	12940.4
3	14703.7	11324.1	3	21085.6	12918.8
4	14693.9	11323.0	4	21076.2	12907.1
	K			P	
1	16026.2	11661.2	1	-	-
2	16014.6	11659.2	2	23271.8	13480.6
3	15997.3	11655.4	3	23248.1	13454.3
4	15986.9	11654.2	4	23236.2	23440.4
	L			R	
1	16842.6	11866.6	1	-	-
2	16830.9	11864.5	2	24802.7	13889.5
3	16814.6	11860.5	3	24777.4	13863.0
4	16804.9	11859.5	4	24764.9	13850.2

TABLE XXI

VELOCITIES COMPUTED FROM SHORT-TERM SURVEYS
STATIONS E - R

	Survey	Time Spread in Days B	LF	AVG	Movm't (m)	Average Velocity (cm/d) AVG	B	LF	Epoch Vel cm/d
E	1-2	145	149	147	8.4	5.71	5.79	5.63	
	2-3	232	232	232	12.5	5.38	5.38	5.38	5.60
	3-4 (Jan)	233	233	233	13.4	5.75	5.75	5.75	
	3-4 (Oct)	141	141	141	8.2	5.81	5.81	5.81	
H	1-2	137	161	149	11.0	7.39	8.04	6.84	
	2-3	230	257	243	16.2	6.67	7.04	6.30	7.11
	3-4	142	115	128	9.5	7.46	6.72	8.30	
K	1-2	136	161	149	11.8	7.91	8.67	7.32	
	2-3	228	257	242	20.0	8.26	8.77	7.78	7.74
	3-4	144	117	130	10.5	8.05	7.27	8.95	
L	1-2	134	161	147	11.9	8.12	8.91	7.42	
	2-3	229	257	243	16.9	6.97	7.40	6.59	7.38
	3-4	143	117	130	9.5	7.35	6.68	8.16	
M	1-2	135	161	148	12.6	9.37	7.86	8.54	
	2-3	228	259	243	18.7	7.69	8.20	7.22	7.40
	3-4	170	115	142	9.8	6.94	5.79	8.56	
N	1-2	133	160	146	18.5	12.66	13.89	11.55	
	2-3	229	258	243	27.9	11.50	12.20	10.83	10.81
	3-4	169	115	142	15.0	10.57	8.88	13.05	
P	1-2	-	-	-	-	-	-	-	
	2-3	218	254	236	35.4	15.00	16.23	13.94	13.46
	3-4	156	115	135	18.3	13.55	11.73	15.91	
R	1-2	-	-	-	-	-	-	-	
	2-3	218	254	236	36.6	15.52	16.80	14.42	13.67
	3-4	171	115	143	17.9	12.52	10.47	15.56	

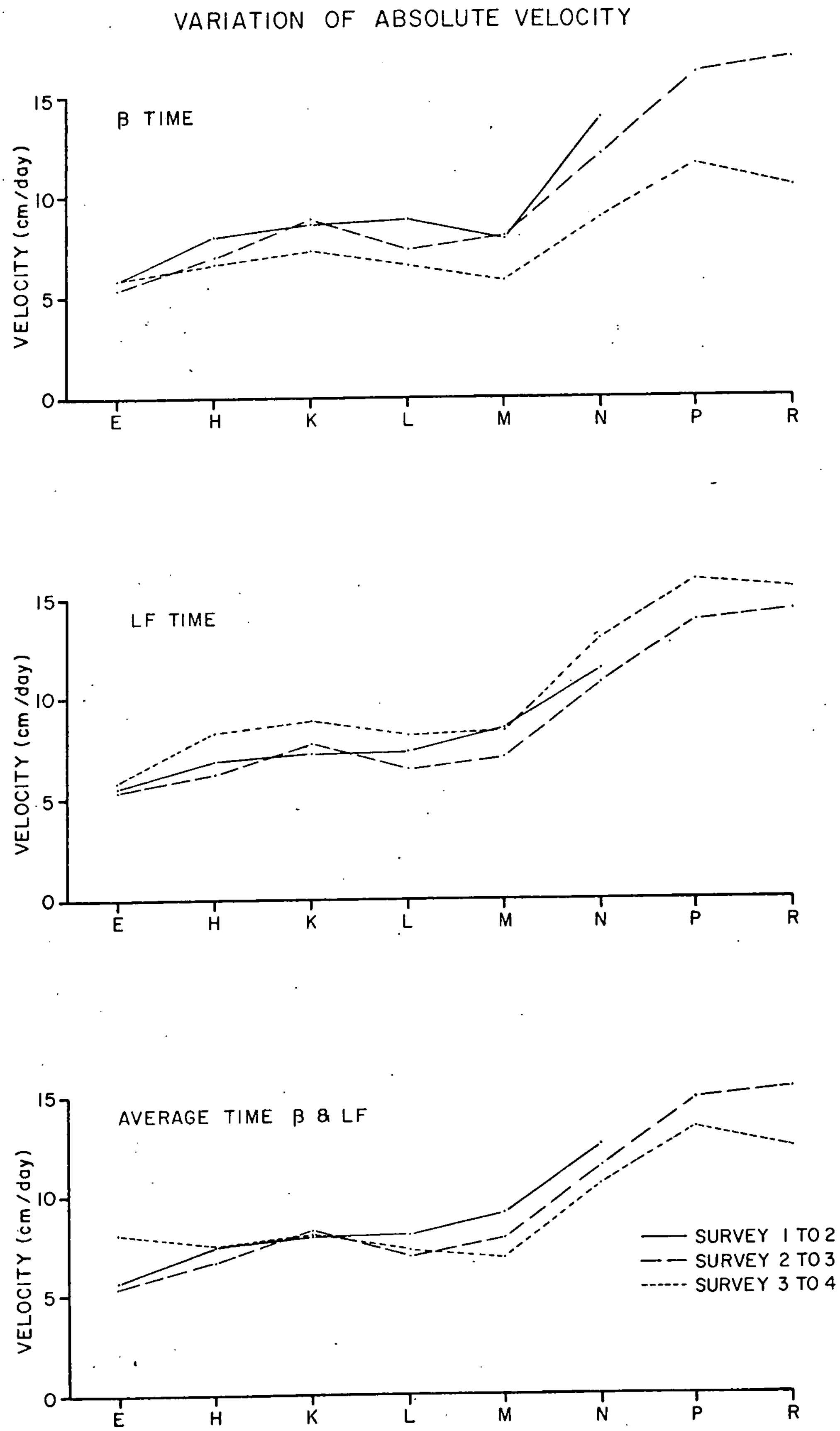


Fig. 36. Variation of absolute velocity between epoch and intermediate surveys based on times of angle measurement, distance measurement and average of both time intervals.

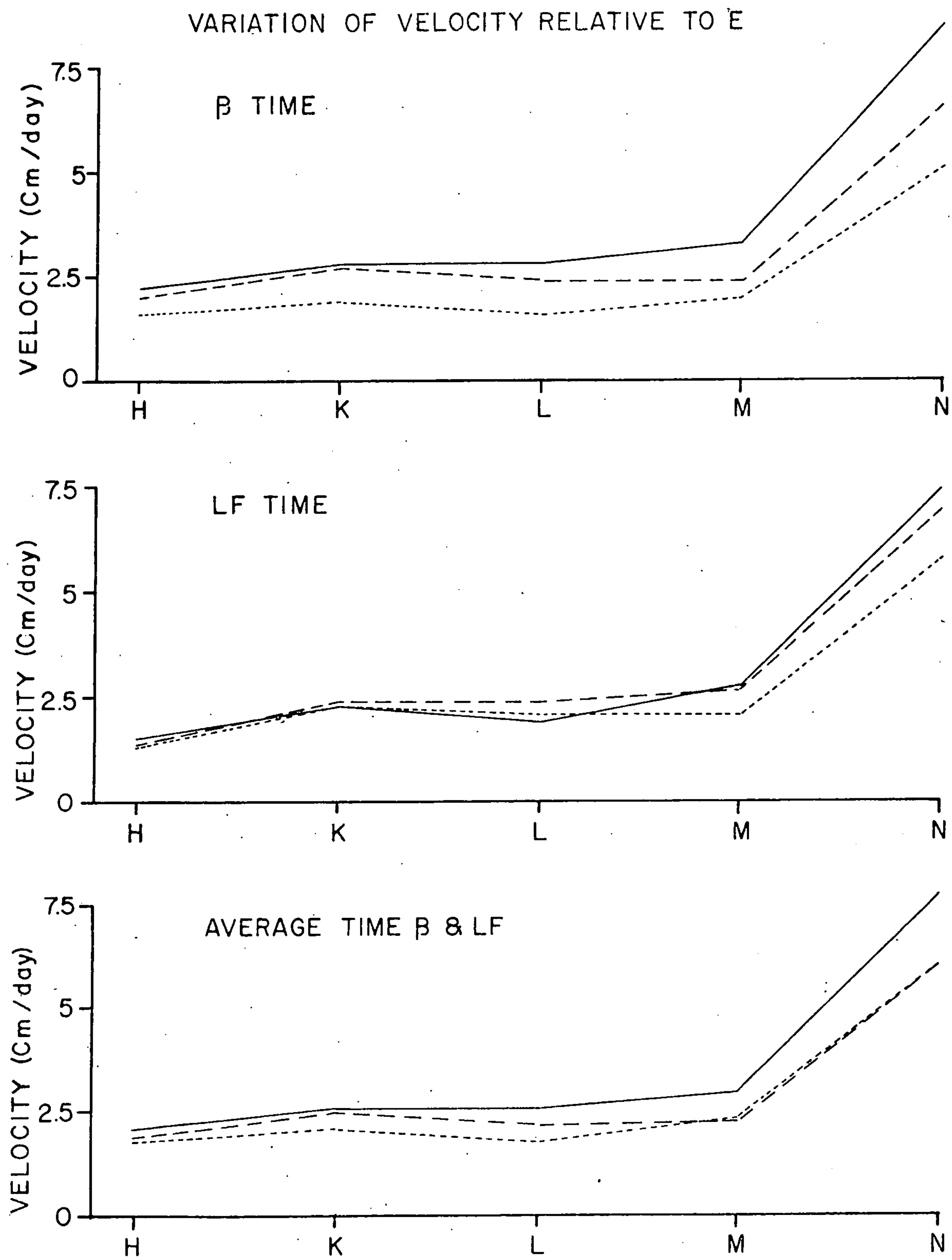


Fig. 37. Variation of velocity, relative to station E, between epoch and intermediate surveys based on times of angle measurement, distance measurement and average of both time intervals.

TABLE XXII

VELOCITIES RELATIVE TO STATION E
COMPUTED FROM SHORT-TERM SURVEYS

	Survey	Velocity (cm/day) AVG.	B.	LF
H	1-2	2.1	2.2	1.9
	2-3	1.9	2.0	1.8
	3-4	1.8	1.6	2.0
K	1-2	2.6	2.8	2.4
	2-3	2.5	2.7	2.4
	3-4	2.1	1.9	2.3
L	1-2	2.6	2.8	2.4
	2-3	2.2	2.4	2.1
	3-4	1.8	1.6	1.9
M	1-2	3.0	3.3	2.7
	2-3	2.3	2.4	2.1
	3-4	2.4	2.0	2.8
N	1-2	7.8	8.5	7.0
	2-3	6.1	6.6	5.8
	3-4	6.1	5.1	7.5

CHAPTER SIX

THE MASS BALANCE OF THE ICE PIEDMONT

INTRODUCTION

The mass balance of a glacier is the algebraic sum of incoming and outgoing mass. The mass balance of small cirque and valley glaciers can be treated rigorously, because they have clearly defined boundaries and because they are relatively easily mapped using modern photogrammetric techniques and aerial photography (LaChappelle 1962) or normal topographic mapping proceedures. Changes in ice volume can therefore be quite accurately determined. On a small feature it is also possible to obtain accumulation and ablation data from closely spaced locations either from stakes set in the surface or by probing to the level of the previous year's icy surface (Schytt 1962). The velocity field can be more precisely determined, not only from the point of view of concentration of velocity stations, but also because topographic conditions allow more accurate measurement of the velocity. Thus, for a small glacier it is possible to use precisely defined parameters (eg Meier 1962) in the mass balance treatment and take a more sophisticated approach to the problem.

As with meteorological data, mass balance measurements increase in value and significance the longer the period over which they have been gathered. Logistic accessibility is therefore an important consideration to a program of mass balance determination. Though there are several notable exceptions from mass balance studies in the Arctic (eg Paterson 1969; Hattersley-Smith and Sersen 1970) it is not surprising therefore, that the best known glaciers, from the mass balance aspect, are small, easily accessible valley and cirque glaciers, usually temperate and generally in the northern hemisphere (eg. LaChappelle 1965; Meier 1965; Hoinkes & Rudolph 1962; Wallen 1948; Schytt 1962). The significance of size and access was summed up by Schytt (1962) in answere to Dr. Hoinkes, who was impressed by the density of accumulation stations on Kebnekajse, and doubted that "one could go on for 50 years like that!"

Dr. Schytt: "You can if you choose a small glacier and have the men to work on it".

(Jour. of Glac. Vol 4 No 3 p 286)

The mass balance of ice sheets, ice shelves and other larger features is more difficult to determine primarily because of their size and the problems presented by that alone. Without an extremely mobile working force of almost army size, the number of accumulation and ablation measurement stations must be limited and it may not be possible to deliberately locate stakes where anamolous values might be expected, for example in an area of pronounced surface topography. Changes in surface elevation are more difficult to determine because of the manifest problems of detailed surveying over larger areas.

Logistically, these areas are often remote and aerial photography and photogrammetry are not feasible. It is frequently the case that data cannot be collected systematically year after year, so long-term averages are not available. Usually therefore, a small section must be treated in detail and assessed as representative of the whole. On the Antarctic Ice Sheet this is the only practical method and, as pointed out by Cameron (1964), the main difficulty is the unknown quantity of ice entering the study area from the inland regions. This is the obvious problem when a study area does not have sensible boundaries. On the Ross Ice Shelf this problem was overcome to some degree by first assessing the contribution to the shelf by the glaciers flowing from the Antarctic Plateau (Swithinbank 1963) and then at a later date, assessing the coastal discharge at the northern coastline (Hoffman & others 1964; Dorrer 1970).

On the Marr Ice Piedmont, mobility of the work party was assured by the use of Polaris and Foxtrack motor toboggans which provide rapid transportation over a smooth snow surface. Thus it was possible to establish the extensive and relatively intensive network of accumulation stakes, which in length amounted to over 90 km. A total of 1,200 bamboo poles were used to maintain the measurement program at over 600 separate locations, and though the density of stations falls far short of the $120/km^2$ obtained by Schytt (1962), the network does provide good coverage of the study area.

The pattern of ice velocity vectors (figure 30) confirms that the study area does in fact reasonably approximate a distinct morphological unit from the point of view of drainage. However, insufficient data are available for a mass balance evaluation of the entire 380 km^2 area defined in Chapter two, p 20-21. Access to the Cape Lancaster promontory by surface vehicle was prevented by crevasses and regular small boat operations were hindered by sea ice. Thus no data have been obtained from this area and an unknown quantity of ice may be entering the region from the foothills of Mt. Moberly and Mt. William. Ice movement station LINDA was lost in crevasses during the early stages of the field work and (she) was not replaced. This station had been established in what appeared to be a distinct and probably fast-moving ice stream (see figure 17) and would therefore have provided important information for the calculation of ice discharge.

The section of the accumulation zone believed to discharge between stations G3 and G5 (on the basis of the velocity vector distribution and surface topographic configuration) is that to the north of the dotted line on the accumulation map shown in Figure 28 (p 91 above). For the calculation of the mass balance, this section of the accumulation zone has therefore been omitted.

BUDGET YEAR

A time framework is required for the calculation of glacier mass balance. This is the "budget year", defined as the interval between the time when new accumulation exceeds ablation and the close of the following summer's ablation season, when accumulation again exceeds ablation. This is the "classical" budget year and on the Marr Ice Piedmont, it begins in late March to early April.

On Anvers Island this concept is not easy to apply, because the Marr Ice Piedmont lies almost exclusively above the firn line and the accumulation of mass is an almost continuous process. What little ablation (by melt) there is, is confined to very small areas on the peripheral ramps and is miniscule compared to accumulation. For the calculation of the mass balance described in this chapter, a "calendar" budget year has therefore been used. This is appropriate as it maximises the use of all available accumulation data, which, as is evident from chapter 4, were obtained on a calendar year basis, a shedule which did not coincide with the "classical" budget year. In a study of the mass balance of the Norsel and Bonapart ramps, described in Chapter 7, it is however, possible to discuss the "classical" budget year.

MASS INCOME

With the exception of nine stakes which were situated on the Norsel and Bonapart ramps, every stake in the network recorded a net gain of mass during the period of investigation.

Annual accumulation values for the entire stake network are available for the two-year period November 1965 to January 1968. The average of these values was used to prepare the accumulation map (figure 28). As explained in Chapter 4, the isolines on this map have been drawn at intervals of 10 g/cm^2/yr. and in order to arrive at the following figures, the area bounded by each isoline was obtained with a polar planimeter and the volume of water represented was calculated.

The total area of the "southern lobe" amounts to about 380 km^2 and excluding the Cape Lancaster promontory, the area to which the accumulation map applies is about 338 km^2. The total positive balance averaged over the two-year period for this area is 516 x 10^6 metric tons. However, because of the uncertainty of the ice discharge between stations G3 and G5, this total area cannot be used, so when the area to the north of the dotted line on the accumulation map is excluded, a total positive balance for the remaining area above the equilibrium line was calculated as 407 x 10^6 metric tons. However, the equilibrium line is situated at approximately 60 m. elevation and closely follows the cliff edge. Therefore, in order to assess the mass balance for the entire area from the cliffs to the foot of the mountains, it would be necessary to calculate the throughflow at the cliff face itself, but movement and thickness at the cliff face are not known. The closest "profile" to the cliff face, for which throughflow can be sensibly calculated, is defined by the line G3 through the "G" stations to E, then to H1, H2 and H3 passing through K3 and C1 to station N7. The total positive balance above this line, over an area of 320 km^2, was calculated as 380 x 10^6 metric tons.

VOLUME OF THROUGHFLOW

The surface area of ice passing annually through this profile has been calculated, using the measured surface velocities and the computed values of the azimuth of ice flow at each ice movenent station. To obtain the volume of ice passing through the profile, this area has been multiplied by the measured ice thickness and a reduction factor of 0.95 to account for the variation of velocity with depth. This factor is based on values obtained by Bull and Carnein (1970) from calculations made from borehole data from the Meserve Glacier, Antarctica and elsewhere.

Assuming an average ice density of 0.90 g/cm^3 throughout, the total mass flowing annually through the profile has been calculated as 360 x 10^6 metric tons.

The resulting mass balance equation is:

Total positive balance above the throughflow profile	380×10^6 metric tons	100.0%
Total throughflow at the profile	360×10^6 metric toms	94.7%
Excess of positive balance over throughflow	20×10^6 metric tons	5.3%

DISCUSSION OF THE MASS BALANCE EQUATION

The excess of positive balance over throughflow is the equivalent of approximately 8.8 g/cm^2 distributed evenly over the surface area of the accumulation zone considered here and could represent a real positive balance for the entire study area. However, with the data available, it is not possible conclusively to show this.

If this is a positive balance, then the surface elevation should be increasing. This is the "velocity of thickening" described by Meier (1960). In the present survey the vertical angles were not simultaneously reciprocal and are subject to an unknown error which would render any calculation of the velocity of thickening rather suspect. Therefore, it is not proven that a positive balance is reflected by elevation change. The total lack of any kind of accurate topographic map of the ice piedmont is also a hindrance to further consideration along these lines.

CONSIDERATION OF ERRORS IN THE MASS BALANCE EQUATION

Several parameters used for the calculation of the mass balance are subject to possible errors. Snow accumulation was measured against the stakes using a two-meter long "tally rod" and an error of ± 1.0 cm. at each reading can be estimated. Some stakes were measured as many as nine times a year and all stakes were measured at least three times a year.

It is difficult to assign a possible error to the densities derived from the snow pit and ice core studies. Almost invariably the snow pit samples were of well consolidated material and the density tubes were most certainly full. A possible error of ± 0.01 g/cm^3 is probably the limit of error which can be applied to these measurements. Error in the core density is likely to be greater though not excessively so because, unless it was obvious that the core was thin, which often resulted from brittle, friable material, the core was assumed to be 76 mm. in diameter. When this was not the case the specimen was measured and in no case was the diameter found to be less than 74 mm. The length of each specimen is a likely source of error but it again is not likely to exceed about 2-3 mm. Thus an estimated error for the core sample densities of ± 0.02 g/cm^3 can be considered as the upper limit. However, a small change in the density values used could significantly alter the mass balance equation.

Annual accumulation rates vary considerably from year to year. The records for positive balance presented in Chapter 4 show that the annual totals for the two years considered in the mass balance equation are considerably different. Furthermore, the velocities used for the throughflow calculation did not cover a period (1967) when accumulation was being measured for the mass balance calculation and it is therefore, strictly an average velocity which has been applied.

Meteorological records also are limited, both for Palmer Station and for the surrounding area as a whole. Thus it is not possible to assess the stability of meteorological conditions and therefore, of snow accumulation conditions, over Anvers Island. The differences between the three annual positive balance records from the Main profile (figure 20) are not readily explainable from the contemporary Palmer Station meteorological records (Appendix I). The difference in these final values is certainly not the result of greater or lesser ablation from year to year as is suggested by Sadler (1968) on the Argentine Islands, but results directly from differences in total snowfall from one year to another. The most likely explanation for the annual snowfall variation is probably related to differences in local sea ice conditions during the winter.

The errors in the velocity determination and ice thickness measurement have been considered elsewhere (Appendix II and Chapter Three respectively) but further comment is warranted. A large proportion of the mass discharging from the study area occurs through the major ice stream between the Central line and the Neumayer line (see figure 30). The discharge velocity used is in effect, the average velocity of stations C1 and N7. This velocity may not in actual fact, be representative of the velocities of stations K4 and K5 for which second-survey data are not available. These stations lie in the mid-section of this ice stream and possibly have greater velocities than stations C1 and N7. The discharge through the ice stream may therefore have been underestimated.

It has also been assumued that the section of the accumulation area which discharges between stations G3 and G5 has been accurately identified. This area, which amounts to approximately 62 km^2 showed a total positive balance of 108.5 x 10^6 metric tons which, to be discharged through the ice stream in which station LINDA was placed, would require of that station a velocity in the order of 170 m/yr. This value is reasonable when compared with the other ice stream velocities and particularly the nearby stations G2 and G3. A very significant error in the final mass balance evaluation has probably not therefore occurred as a result of this particular assumption.

The reduction factor of 0.95 used to account for the variation of ice velocity with depth is estimated and is based on the findings of Bull and Carnein (1970) for Meserve Glacier, which is cold. From their borehole data they found that a reduction factor for that glacier was between 0.83 and 0.92. Temperature measurements on the Marr Ice Piedmont and discussed in Chapter 10 below and consideration of the average temperature condition within this ice from surface to bottom, indicate a much warmer condition. Sliding velocities are high (Chapter 11 below) in the ice streams and in other areas, where the internal deformation velocities and the surface velocity are about the same, the ice could be moving as "plug flow". Under plug flow conditions, the reduction factor would be 1.00 and under conditions of high sliding velocities, the factor might well be in excess of 0.95. With regard to these parameters of average englacial temperature and flow characteristics, the factor used in the mass balance calculation may be low. It is not likely however, to be 1.00 for the entire study area.

An ice density of 0.90 g/cm^3 has been assumed for the throughflow calculation. Maximum density of ice (0.917 g/cm^3) has been used by some workers, though the average density may be less than this. The value used in the present calculation could be high by 2.0%.

This study has not been able to assess the contribution to the positive balance by ice and snow movement from the southwestern flanks of the Achaean Range because no accumulation, thickness or movement data are available. It is probable that the snow cover of these flanks is thin and that because the significant positive balance to the piedmont is by deposition directly on its surface, the contribution from the mountain walls is relatively small and is not likely to significantly effect the mass balance equation presented.

The major error in the entire mass balance study lies with the isolined accumulation map, from which total positive balance values were derived. Such a map is obviously subjective to a greater or lesser degree. The position of each isoline is accurate only at the points of measured accumulation. It is between these points that the lines become subjective. The inaccuracy cannot be quantitatively analysed but is likely to be greater in the northern part of the study area where known positive balance values are more scattered. Fortunately, much of this area has been omitted as it would discharge through station LINDA into Wylie Bay.

CONCLUSIONS

The possible errors in the mass balance evaluation are sufficient to account for the apparent positive balance for the study area. The physical dimensions of the ice piedmont do not appear to be changing significantly; the coastal boundary seems to have been stable for at least the past ten years. The mass of the study area is in the order of 90×10^9 metric tons, and if the 20×10^6 metric tons is real positive balance, it would represent a gain to the study area of less than one tenth of one percent annually, which is an insignificant departure from equilibrium.

This conclusion is drawn from a study of only part of the Marr Ice Piedmont and it is not known whether this part is truly representative of the piedmont as a whole. There is no evidence that precipitation in the northern parts is higher than in the southern section considered here. If it were it would require higher ice velocities to give equilibrium conditions. There are no velocity data available from the northern area so it must be concluded that if the study area is indeed representative of the whole, then the Marr Ice Piedmont itself is in or is very close to equilibrium. There is little reason to imagine that the piedmont could have a negative regime.

Such a conclusion is not surprising. It was suggested above in Chapter 3, p 45, that the horizontal dimensions of the piedmont are governed by sea level; a rise in sea level would cause a recession of the ice front and a fall in sea level would cause an advance. Therefore, because equilibrium is maintained by sea level-controlled calving, a positive or negative regime would not be detected. Any change in the activity of the regime would be reflected by changes in velocity which would change the amount of calving and probably reestablish equilibrium in a relatively short time. Only on the land based areas of the ice terminus (the ramps) would the true picture of the regime be revealed.

From studies on the Norsel Point and Bonapart Point ramps, this matter can be further pursued and the mass balance equation presented above (p130) can be elaborated upon and refined. The discussion follows in Chapter 7.

CHAPTER SEVEN

THE REGIME OF
NORSEL POINT AND BONAPART POINT RAMPS

On January 20 1965, 26 poles were emplaced on the Norsel Point ramp (figure 38) in a trail finding operation but the measuring program was not begun until February 3 when a safe trail through the crevasse field had been established. At that time the remaining 1964 winter snow was probed to the previous year's ice level or ice layer and depth was established. Depth ranged from 75 cm. to 129 cm. The lowest values were not at the foot of the ramp but some way up-slope, suggesting a tendency for snow to drift at the foot - a situation later confirmed by measurements in 1965 and 1966. On January 28 a 125 cm.-deep pit was excavated near pole number 2 and average snow density proved to be 0.532 g/cm^3. The content of the profile was computed to be 66.5 g/cm^2. Gradual ablation and subsequent surface lowering resulted in bare ice by March 3 at poles 1, 3 to 5, 7-11 and at number 14. At numbers 2, 6, 12 and 13 and all above number 14, heavily iced firn remained.

Steady ablation continued until April 8 and the resultant lowering of the ice surface ranged from zero at stake number 14 to 23 cm. at number 11 with an average of 12.2 cm. If the ice surface exposed on March 3 was old glacier ice and not superimposed ice formed during the winter of 1964, this surface lowering represents net loss of mass from the ramp. At an estimated ice density of 0.85 g/cm^3, the average net

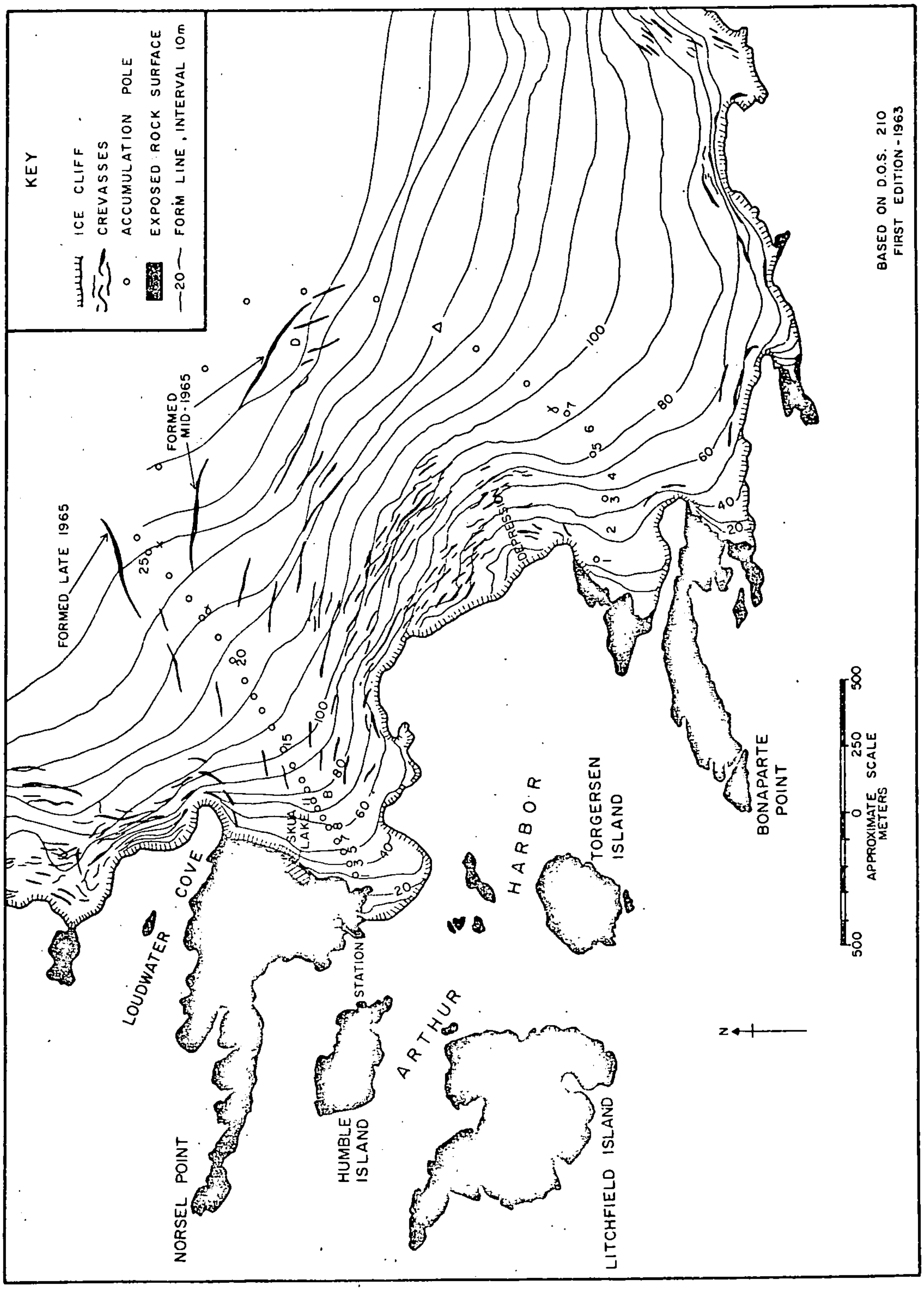


Fig. 38. Study area around Arthur Harbor.

loss at the nine stakes recording negative balance was 10.4 g/cm^2. Positive balance was recorded at stake number 2 where 8 cm. of iced firn remained (4.2 g/cm^2) and at number 6 with 7 cm. (3.7 g/cm^2). At numbers 12 and 13 mere icy patches remained; 2 cm. (1.1 g/cm^2) at number 12 and 3 cm. (1.6 g/cm^2) at number 13. At number 14 no significant surface lowering occurred after March 3 resulting in equilibrium at that point. No information is available on total accumulation during the 1964 winter and as the previous year's iced surface was not reliably located, it is not possible to establish the balance above stake number 14. However, it seems reasonable to assume that it was positive.

On April 8, 15 to 20 cm. of fresh snow remained on the surface and therefore, defined the beginning of the 1965 "classical" budget year. Frequent measurements were made throughout 1965 and the results from several representative stations are shown in Figure 39. These are cumulative curves of change in surface level (or snow accumulation trends in cm. of snow) because though density determinations were frequently made; they rarely coincided with the time of actual stake measurement, therefore the exact water content of the profile at many specific times is not accurately known. The results of the density studies, most of which were made on the lower part of the ramp, are given in Table XXIII and extrapolating the density values produces a generalised graphical summary of accumulation rates, in g/cm^2, at stake number 2, shown in Figure 40.

The curves in Figure 39 differ slightly from one another in detail

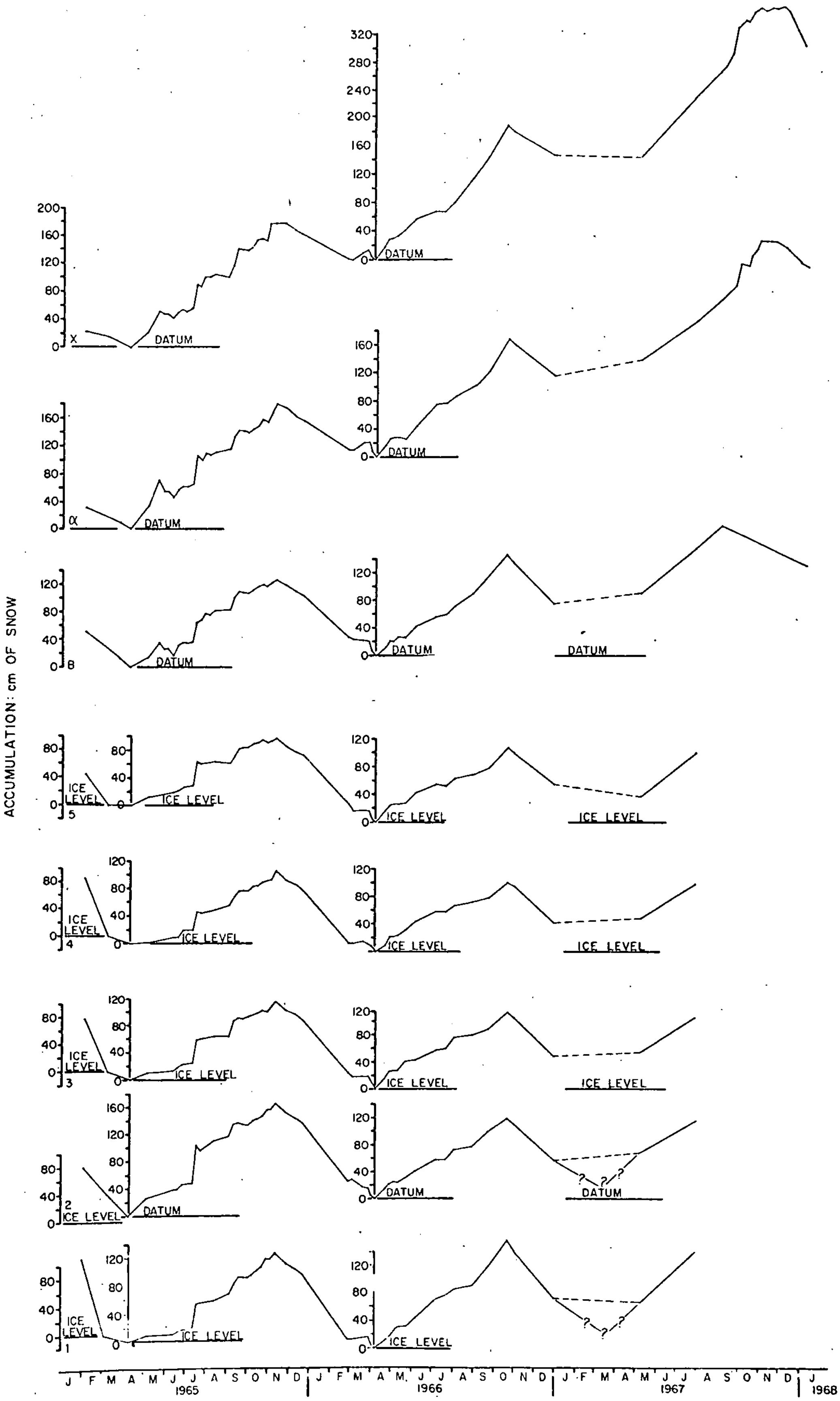


Fig. 39. Cumulative curves of snow accumulation from selected stations on Norsel Point ramp.

TABLE XXIII

SNOW DENSITY: NORSEL POINT RAMP AND PERIPHERAL AREAS.

Date 1965	Depth of Profile	Density gm/cm³	Profile Content	
Jan 28	125 cm	0.532	66.5	
May 23	42 cm	0.354	14.9	
May 31	35 cm	0.364	12.7	
Jun 5	49 cm	0.396	19.4	
Jun 9	41 cm	0.395	16.2	
Jun 15	39 cm	0.360	14.0	
Jun 21	38 cm	0.376	14.3	
Jun 26	72 cm	0.333	24.0	
Sep 16	90 cm	0.354	31.9	
Oct 27	113 cm	0.408	46.1	
Nov 19	114 cm	0.433	49.4	
1966				
Apr 8	40 cm	0.518	20.7	
Jun 2	30 cm	0.353	10.6	
Jul 22	56 cm	0.301	16.9	
Aug 13	131 cm	0.383	50.2	At Stake Nr. 18.
Sep 12	156 cm	0.366	57.1	At Cormorant Point.
Sep 23	116 cm	0.373	43.3	
Nov 19	122 cm	0.401	48.9	At Station Big X.

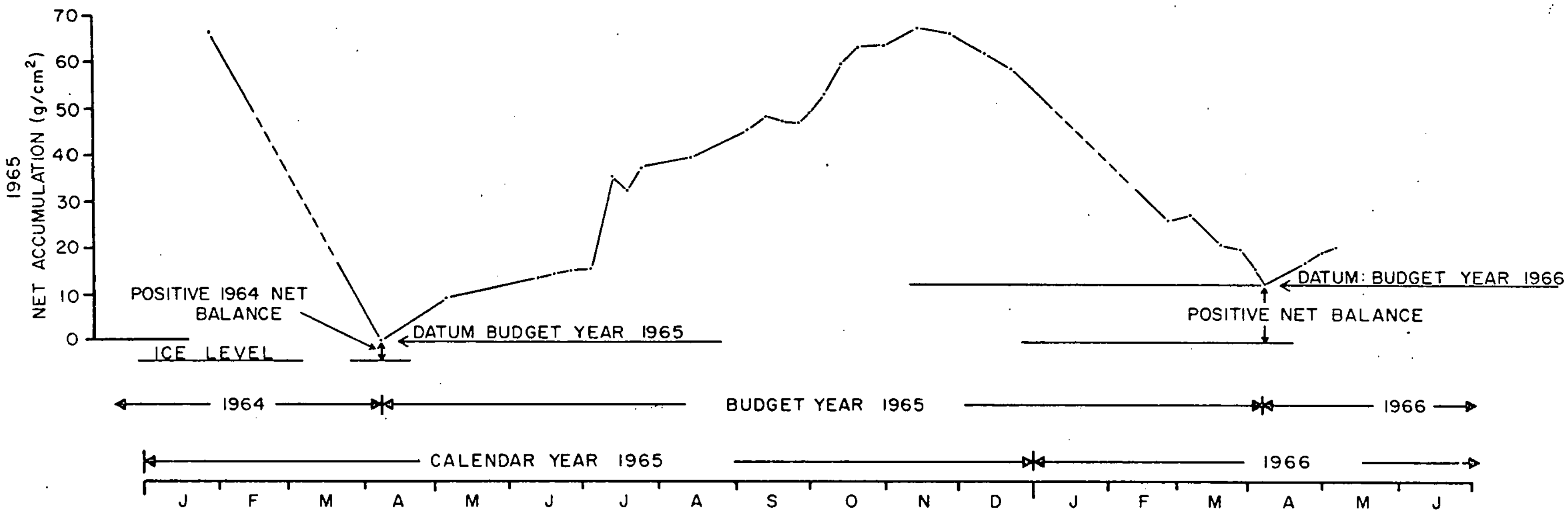


Fig. 40. Summary of mass balance at ramp stake number 2 based on density data taken from close to that stake. Budget year 1965.

but in general they all record nearly the same pattern of accumulation and ablation. The effect of drifting is clearly evident at the lower stakes but it is notable that the lower 1966 accumulation recorded over all the study area is not strongly evident on the ramp. With the exception of two heavy snowfalls in mid-July and early September, the rate of snow accumulation in 1965 was generally steady and reached a peak on November 13. At that time gross accumulation based on a density of 0.408 g/cm^3 recorded on October 27, had reached approximately 40 g/cm^2 at stake number 5, 51 g/cm^2 at stake 11, 73 g/cm^2 at number 22 and 72 g/cm^2 at number 25 at the top of the ramp.

Slow ablation after November 13 caused surface lowering of between 20 and 30 cm. by December 22, and by February 27 1966 almost all of the 1965 winter accumulation had been removed at the lower stakes. From then until March 28 the surface was generally stable under light snow and sleet conditions but from March 28 to April 7, ablation was heavy under high air temperatures and heavy rainfall and the surface was lowered below the previous year's level at six stakes; numbers 1 and 3 to 7.

The net loss at these six stakes ranged from 4 cm. of ice, density 0.85 g/cm^3 (3.4 g/cm^2) at stake 7 to 24 cm. (20.4 g/cm^2) at number 5. The average negative balance was 12.0 cm of ice (10.2 g/cm^2. Stake number 2 recorded a positive balance (probably due simply to the drifting effect) of 25 cm. of very wet snow with a measured density of 0.518 g/cm^3 giving 12.9 g/cm^2. Fresh snowfall on April 7 remained on the surface and by April 24 had reached an average depth of 15 cm. ranging from 6 cm. at stake 7 to 22 cm. at number 18. Thus, the 1966 budget year (by remarkable coincidence) began one day earlier than in the previous year.

All the stakes above number 7 recorded a positive balance during the 1965 budget year. The results, based on the April 8 1966 density of 0.518 g/cm^3 are shown in Figure 41. For comparison, the net budget for 1964 at stakes 1 to 14 is included in Figure 41.

Meanwhile, on the Bonapart Point ramp, six stakes had been emplaced below velocity station Gamma (see figure 38) on December 23 1965, when prevailing snow depths ranged from 71 cm. at stake number 5 to 52 cm. at number 2. By February 27 1966, bare ice was exposed at the first two stakes and 1965 winter snow remained at the other stakes, ranging from 5 cm. at number 3 to 30 cm. at number 6.

As on the Norsel Point ramp, the surface level was generally stable until March 28, after which heavy ablation lowered the surface to below the previous year's level, but at three stakes only; numbers 1, 2 and 3. The resulting negative balance at these three stakes was 36 cm (30.6 g/cm^2) at stake 1, 25 cm. (21.2 g/cm^2) at number 2 and 23 cm. (19.5 g/cm^2) at stake 3. Average net loss was 23.8 g/cm^2. This is twice the value from the Norsel Point ramp and might be a result of the more northerly facing aspect of the Bonapart ramp and its slightly greater exposure to direct summer radiation.

At stake number 5, the old surface was exposed but not ablated indicating equilibrium at that point, while stakes 4, 6 and 7 (Gamma) recorded positive net balance of 6 cm. (3.1 g/cm^2), 15 cm. (7.8 g/cm^2) and 8 cm. (4.1 g/cm^2) respectively. The results from the Bonapart ramp are shown in Figure 42.

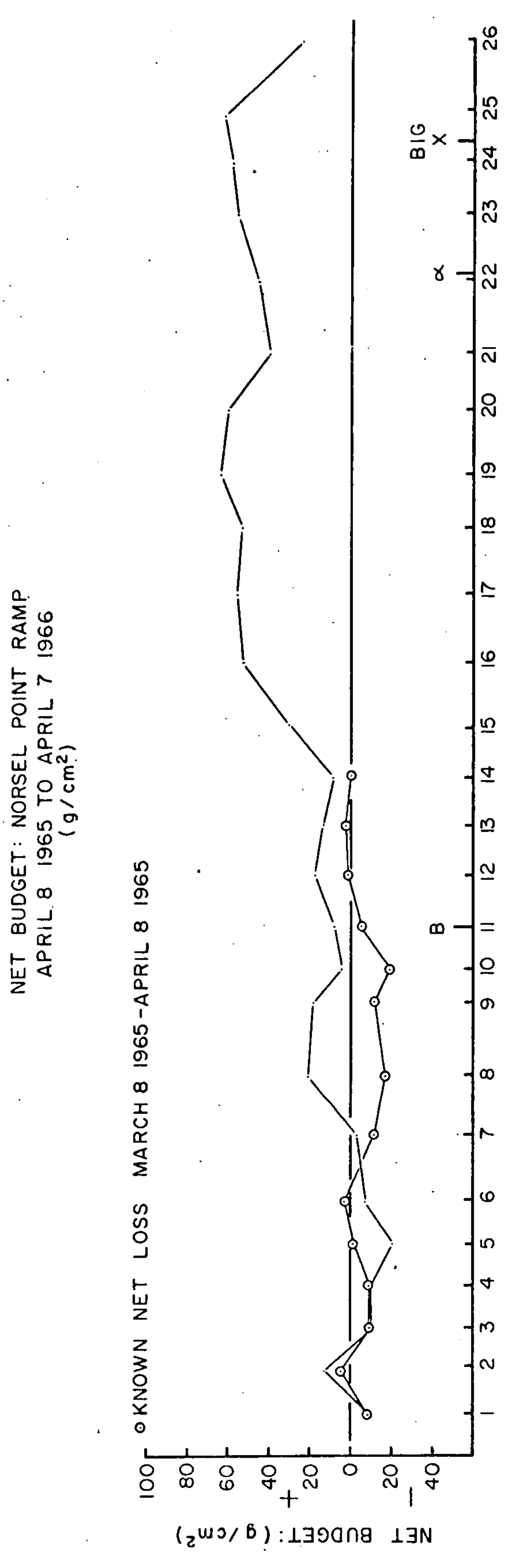

NET BUDGET: NORSEL POINT RAMP
APRIL 8 1965 TO APRIL 7 1966
(g/cm^2)
○ KNOWN NET LOSS MARCH 8 1965-APRIL 8 1965
NET BUDGET: (g/cm^2)
100 80 60 40 20 0 20 40
+
−
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26
B
∝
BIG X

Fig. 41.

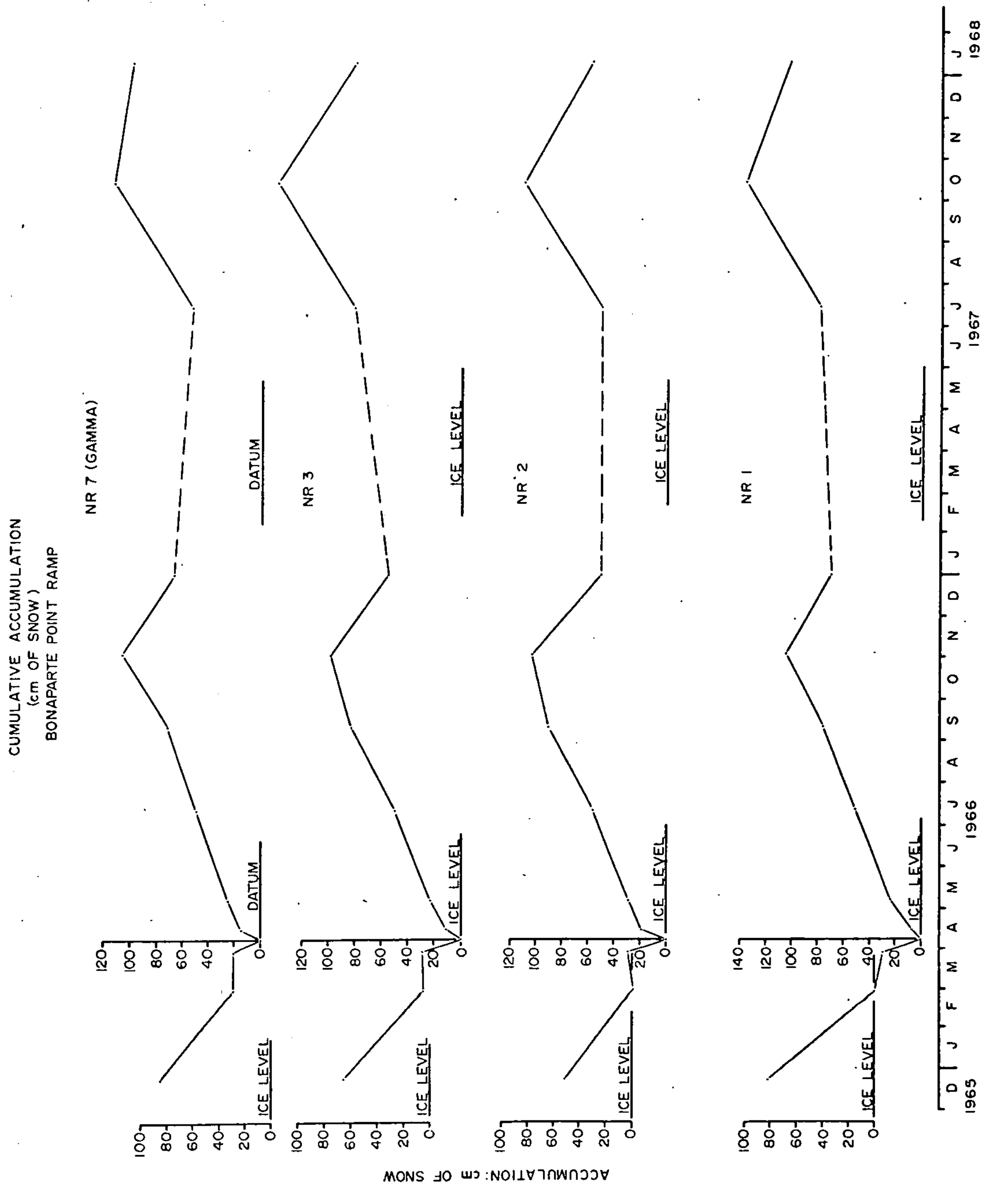

CUMULATIVE ACCUMULATION
(cm OF SNOW)
BONAPARTE POINT RAMP
NR 7 (GAMMA)
NR 3
NR 2
NR 1
DATUM
ICE LEVEL
ACCUMULATION: cm OF SNOW
1965
1966
1967
1968

Fig. 42.

At the end of the 1965 budget year, the equilibrium line was situated at approximately 60 m. elevation on the Norsel Point ramp (station B is at approximately 80 m. elevation) and at about 80 m. elevation on the Bonapart Point ramp (Gamma lies at 112 m. elevation). However, as is evident from Figure 41, this line was considerably higher at the end of the 1964 budget year on the Norsel Point ramp indicating considerable mobility of the equilibrium line from year to year. From the situation of the curves in Figure 41, it seems reasonable to speculate that the equilibrium line fluctuates generally between 60 m. (stake 7) and 100-110 m. (stake 14) elevation on the Norsel Point ramp. On the Bonapart ramp it may reach to higher elevations in some years; a British fuel cache at station Gamma was left on the surface,(reportedly) in 1963, by F.I.D.S. personnel on or about December 26. There are no significant signs of accumulation or ablation around this cache indicating that it must lie close to the equilibrium line in frequent years. However, from information supplied by Honkala's party, it appears that in some years the equilibrium line does not exist on either ramp.

During the 1966 winter, the pattern of snow accumulation was much the same as in the previous year, except that peak accumulation was recorded on October 22, three weeks earlier than in the previous year. On September 23, average snow density on the lower part of the ramp was computed to be 0.373 g/cm^3 and using this as a basis, the peak accumulation was approximately 40 g/cm^2 at stake 5, 54 g/cm^2 at stake 11, 63 g/cm^2 at number 22 and 72 g/cm^2 at number 25; much the same as in the previous year.

Gradual ablation and surface lowering were recorded until December 31, when this author's party made its final measurements, by which time approximately half the 1966 winter accumulation had been removed (54 cm. of snow then remained at stake 5 and the snow density on November 19 was 0.401 g/cm^3). Unfortunately, Honkala's party did not maintain a close record of the ramp stakes and budget figures for 1966 are not available. However, I.M. Whillans in a personal communication, reported that at no place on either ramp was bare ice exposed in the early part of 1967. Thus, all stakes recorded positive balance during the 1966 budget year, though the magnitude is not known. It is probable however, that at the lower stakes at least, the balance was small after continued ablation beyond December 31 1966 and that the curves dipped somewhat as shown in Figure 39.

The record from only two stakes, numbers 22 and 25, are of significant value in assessing conditions of regime during the 1967 budget year. At most stakes, measurements were made on September 7 1967 and January 9 1968, but the more important lower stakes were last measured on July 30 1967. Studies by I.M. Whillans, who recorded snow accumulation rates in detail at stakes 22 and 25 between September 7 1967 and January 9 1968, indicate that peak snow accumulation was reached on November 3 1967 but that the surface level remained stable until December 9, after which, heavy ablation rapidly lowered the surface until January 9 when his last measurements were made. However, much of the 1967 winter snow remained on the ramps when Honkala's party left Palmer Station in early January 1968. This can be seen in Figures 39 and 42. Thus, unless meteorological conditions in early 1968 were extremely favorable for ablation (high air temperatures and heavy rainfall, for example) it appears highly probable that a positive net balance resulted in the 1967 budget year.

DISCUSSION

On the basis of these studies it is clearly evident that the sign of the net budget of the ramps varies from year to year, in some years being slightly positive, in other years slightly negative. Both gross accumulation and ablation are strongly influenced by elevation. Year to year variations in the magnitude of the net budget are not strongly emphasised by this study and net ablation in the 1964 and 1965 budget years was almost identical. However, net ablation occurred over a much wider area in 1964 than in the following year. It is possible that this resulted because of lower gross accumulation during the 1964 winter, though the probed depths of January 1965 only partly support this hypothesis. The positive budget of 1966 was probably small and unless ablation factors were particularly strong in early 1968, the net budget for 1967 most probably was positive and greater in magnitude than the previous year. Gross winter accumulation in the 1965 and 1966 budget years was generally similar but the negative budget of 1965 was reversed the following year. On this basis, ablation rather than accumulation factors appear to have been the determinant of the sign of the regime. If the 1967 budget year was strongly positive, it resulted from a combination of high accumulation and, possibly, weak ablation factors.

SUMMARY AND CONCLUSIONS FROM ALL MASS BALANCE DATA

The ramp study was of too short a duration to allow firm conclusions to be drawn concerning mass balance, but taking the limited data at face value one might justifiably assume, that even though the sign of the regime changed from year to year, there was no serious departure from equilibrium during the period of investigation. Furthermore, it can be concluded that the true state of balance of the ice piedmont as a whole, will more likely be revealed by an understanding of the condition of the ramps rather than by such a study as was described above in Chapter 6. However, neither that study nor the ramp study are particularly helpful in determining or deducing the state of balance on a long-term basis; that is, over a long period of time, has the piedmont experienced a generally positive or negative regime? Circumstantial information relating to this problem is discussed briefly in Chapter 8 below.

Meanwhile, it is obvious that two mass balance studies were made, the "ice stream" balance and the ramp balance and because they utilised different time frameworks, they cannot be combined*. However, the data obtained from the two studies are helpful toward an understanding of the relative significance of the various processes of mass gain and loss in the overall regime of the ice piedmont.

The net balance figures recorded by the stakes in the study area account for all processes of accumulation and ablation. In the equation presented above (p 130), the total positive balance above the

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* See Journal of Glaciology. Vol. 8. No. 52. 1969. p 3-7.

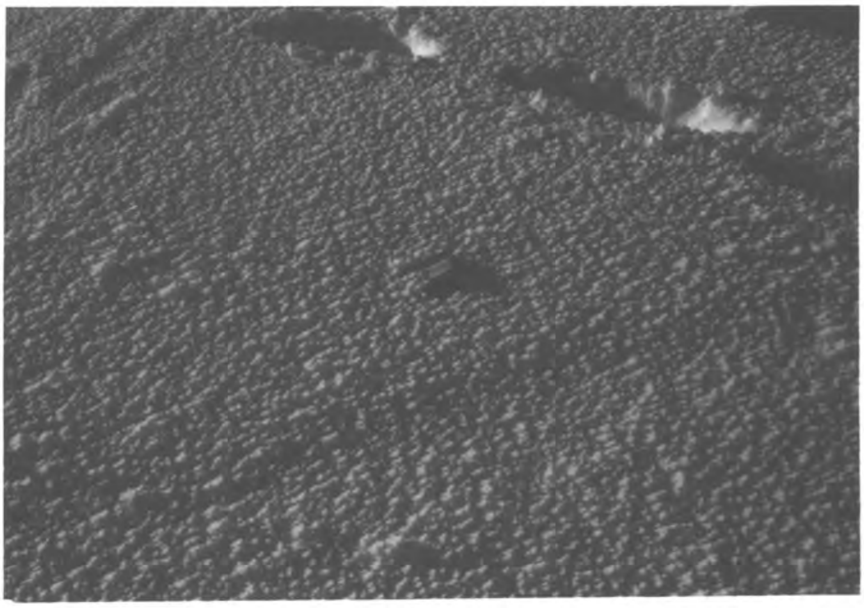
throughflow profile was equated with the throughflow at the profile and approximate equilibrium was concluded. However, that equation is derived entirely from the accumulation zone and it remains to be considered what processes contribute to the positive balance and the importance of processes other than calving in the loss of mass from the piedmont.

From the study in Chapter 6 above it was assumed that on the average, 360×10^6 metric tons is the annual net ablation from the study area. In reality, this is an oversimplification because that figure refers to throughflow and mass is being added to the area below the profile because virtually all of the study area lies above the equilibrium line. However, this does not detract from the relative importance of the various processes of loss. In the case of accumulation, the processes operating above the profile are restricted to snowfall and rime formation. Accumulation below the profile occurs in other forms but has not entered the preceding equation (p.130). Thus the discussion which follows concerns the entire study area and is, in some respects merely qualitative.

Accumulation

To summarise on accumulation (Chapter 4), snowfall is the major source of mass. Rainfall is a source of mass below the 200 m. elevation contour, but above this level, sleet rather than rain has been recorded.

In the higher parts of the study area, accumulation occurs in the form of rime as thin surface deposits (figure 43a). The remains of



a)

b)

Fig. 43. Rime formation. a) Surface deposit; depth approximately 2 cm. b) Formation on elevated objects. View looking ESE indicating southerly wind aided rime buildup. Deposit on stakes about 7 cm thick.

these layers, often up to 1.5 cm. thick, were seen in snowpit walls but were too thin to sample individually. It is estimated that their contribution to the total accumulation as exposed in the pit walls, is not likely to exceed one tenth of one percent.

Drifting snow causes a redistribution of accumulation over the surface and its contribution to the positive balance may vary from one location to another. It has not been possible to identify this contribution quantitatively in the study area though it must be assumed that snow did drift into the area from the northern part of the piedmont. However, an equal amount of snow may have been blown from the study area into the sea. Again, as in the case of surface rime, drifted snow deposits are not likely to constitute a major source of mass income to the study area.

Core studies on the Norsel Point ramp, discussed below in Chapter 8, indicate that superimposed ice forms between about 50 m. and 150 m. elevation. C.C. Plummer, who did the work during 1965, showed that "with the exception of one centimeter of fine grained ice that evidently formed during the winter" (Plummer's unpublished report to this author), there was no superimposed ice at 50 m. elevation but that it was 512 cm. thick at 150 m. elevation. Above this level the upper 134 cm. of ice contained iced firn. Later core studies at the same locations in 1966, and weekly stake measurements, failed to show any further superimposed ice formation. Either the accumulation by this process was too small to detect or it did not occur at all during that year. Similar work carried out by I.M. Whillans in 1967 was also inconclusive (pers.com).

On the basis of these considerations, a more refined statement of positive balance for the study area can be written thus:

Positive Balance

Snowfall and Rainfall	Approximately	100.00%
Surface Rime	Trace	0.10%
Drifting Snow	Not definable - insignificant	-
Superimposed Ice	Did not occur during investigation	-
Total Positive Balance		100.00%

Ablation

Ablation is not as readily dealt with as accumulation but the following ablation processes are known, or can justifiably be assumed to be operative in the study area: 1) Calving, 2) Surface Melt, 3) Basal Melt, and 4) Evaporation and Sublimation.

Net ablation by surface melt occurred in only two of the four summer seasons for which data are available, namely, 1964-1965 and 1965-1966. The average surface lowering in those two melt seasons was 12 cm/year on the Norsel Point ramp and 20 cm/year on the Bonapart Point ramp. These records show that the equilibrium line lies at about 60 m. elevation. Over these two ramps, the area lying below 60 m. elevation, obtained by planimeter, amounts to 312,000 m^2. Assuming that the stake measurements of surface lowering are representative of all this area, then the mass lost would have been 0.045×10^6 metric tons. Other parts of the study area below 60 m. elevation and occurring as ramps and not calving cliffs, again obtained by planimeter, accounts for 0.600×10^6 metric tons. Total loss by surface melt, when it occurred, is therefore, estimated as 0.645×10^6 metric tons.*

* See footnote at end of chapter.

Snow stratigraphy studies (Chapter 9 below) indicate that surface melt occurs over much of the piedmont above the equilibrium line. However, the resulting meltwater percolates downward and refreezes and though there may be a slight lowering of the surface in summer (see Figure 24, p.79) this process does not constitute net ablation from the piedmont.

Melting at the base of the piedmont may also occur though outwash streams at the base of the cliffs have not been recorded. More information is needed regarding the amount of ice that is afloat, if any, before a sensible estimate can be made.

Evaporation and sublimation probably occur over the entire piedmont throughout the year, though it is not possible to identify these parameters in the present study. The accumulation values derived from the stake measurements must in effect be regarded as net positive balance values. They probably are not representative of the true gross positive balance which has been reduced to the measured values by such ablation processes as evaporation and sublimation. Detailed studies of the energy exchange at the surface are needed before these processes can be identified quantitatively.

Calving is the dominant ablation process and may ultimately account for almost all of the 360×10^6 metric tons passing through the throughflow profile.

It is not possible to relate directly all the ablation data and the accumulation data and create an equation. Also, because of the difference in time framework and the fact that net ablation by surface melt occurred generally outside the time when accumulation was recorded, it is not possible to enter all the ablation data into the mass balance equation presented above on page 130 . In that equation calving is the ablation process. Quantitatively therefore, it can simply be said that ablation processes detected at some time during the period of investigation, accounted for negative balance as follows:

Calving	Approximately 360.000 x 10^6 m.t.
Surface melt - Arthur Harbor	0.045 x 10^6 m.t.
Surface melt - all other ramps	0.600 x 10^6 m.t.
Bottom melt	Trace?
Evaporation/sublimation	Not definable

Paradoxically however, and allowing for the slight overlap in the time frameworks, all data appear to point towards a slightly positive mass balance during 1966 and 1967 when there was no net ablation from the ramps and the main study suggested a possible 5.3% excess positive balance. This is significant in view of the discussion to follow in Chapter 8.

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* Footnote:

During the 1969-70 austral summer, Honkala** measured the flow of 135 melt streams at several locations along the Anvers Island coast. The total length of coastline studied was estimated at 30 km. Over the 54-day period December 18 1969-February 12 1970, 3.01 x 10^6 metric tons was discharged as water from the piedmont. This is six times the amount estimated by Rundle (1970) and this thesis but it still represents less than 1% of the 360 x 10^6 metric tons throughflow.

Honkala did not control his experiment with snow stake measurements and it is probable that much of his measured discharge was loss of the 1969 winter's gross accumulation and not net loss from the piedmont.

It seems probable therefore, that 1% is the upper limit for loss by surface melt and runoff.

** Honkala, R.A. (1971). Melt as a Factor in Ice Loss at Anvers Island, Antarctica. Unpublished report, National Science Foundation, Washington DC.

CHAPTER EIGHT

STRUCTURE OF NORSEL POINT RAMP

As emphasised above the land-based ramps are probably the more representative indicators of the piedmont's regime, which from the ramp study is slightly positive in some years and slightly negative in others. This raises an important question concerning the recent history of the piedmont. On a long-term basis, is there a greater frequency of positive regime or of negative regime? C.C. Plummer's study of the structure of Norsel Point ramp is helpful in approaching this question.

Ice samples were studied both macroscopically and in this section between crossed polaroids and the distinction between superimposed ice and "old glacier ice" was made on the basis of several criteria. The following information comes from Plummer's report to this author.

Superimposed Ice

1). Grain Size:- Although there were minor variations vertically in grain size within the core sections, grain size was essentially constant in a horizontal direction. The mean diameter of grain cross section was nearly always 1 mm to 5 mm. Rarely, the grain size was as small as 0.5 mm or as large as 15 mm. The ice was generally very bubbly and exhibited a pronounced banding of the bubbles. The bubble banding however, did not necessarily correspond to grain size banding.

2). Texture:- The texture was very regular, a sugary mosaic which was obvious even in the newly extracted cores.

3). Bubble-Banding:- The most prominent feature of the superimposed ice was the concentration of bubbles into bands, which varied both in the concentration of the bubbles and in actual bubble size. The majority of the bubbles were near-spherical and most had a diameter in the range 1 mm to 5 mm. Isolated bubbles were as large as 10 mm in diameter while bubbles of less than 1 mm were common but accounted for only a small volume of the air in the samples.

Glacier Ice

1). Grain Size:- Generally the grains were very large with mean diameter ranging from 5 mm to 20 mm. Grains with 50 mm diameter were not uncommon. Some fine grains were dispersed throughout the glacier ice.

2). Texture:- The grains were irregularly shaped with curved and embayed inter-grain boundaries. There was a tendency for the grains to be elongate vertically.

3). Structures:- The glacier ice in places showed evidence of fracture in the form of small cracks. The ice filling these cracks had the same texture and grain size as the surrounding glacier ice but was almost bubble-free.

4). Bubbles:- There were abundant bubbles but these were not banded or stratified. Most were elongate with the long axis inclined at about 20° from the horizontal.

5). Dirt:- Several globules of cryoconite were observed in the upper part of the glacier ice. Under certain lighting conditions this part of the ice had a darker appearance than the superimposed ice, which is believed to result from microscopic dust particles dispersed the sample. This dirt and dust suggests that the glacier ice surface was at some time exposed at the surface prior to the superimposed ice being formed upon it.

Thickness of Superimposed Ice

The cores showed a rapid increase in the thickness of superimposed ice through a relatively short distance up the ramp. A core taken from between stakes 10 and 11 (see Figure 44) contained no superimposed ice (with the exception, as mentioned above, p.153 of 1 cm of fine-grained ice which had evidently formed during the winter). At pole number 12 the superimposed ice was 52 cm thick. At pole 13 it was 312 cm thick and by pole number 14 it was 519 cm thick, but here the upper 134 cm contained fine layers of consolidated iced firn.

Thus, a large wedge of superimposed ice restes on the glacier ice in this area and from these observations can be inferred a general structure as shown in Figure 45.

Discussion

Though somewhat circumstantial in nature, these results are provocative and the cryoconite concentrations and the dusty glacier-ice surface are intriguing. They allow a tentative hypothesis on the recent history of the piedmont. It is possible that the dirt and dust may have washed downward from a prevailing snow surface to concentrate on the old ice surface but it seems equally probable that the old ice surface, below the superimposed ice, was at some time exposed as an ablation surface. This leads to the futher hypothesis that at some time in the past the ramp ice surface was in recession and that later a positive regime, reflected by the superimposed ice, was initiated. On a long-term basis, with variations between slightly positive and slightly

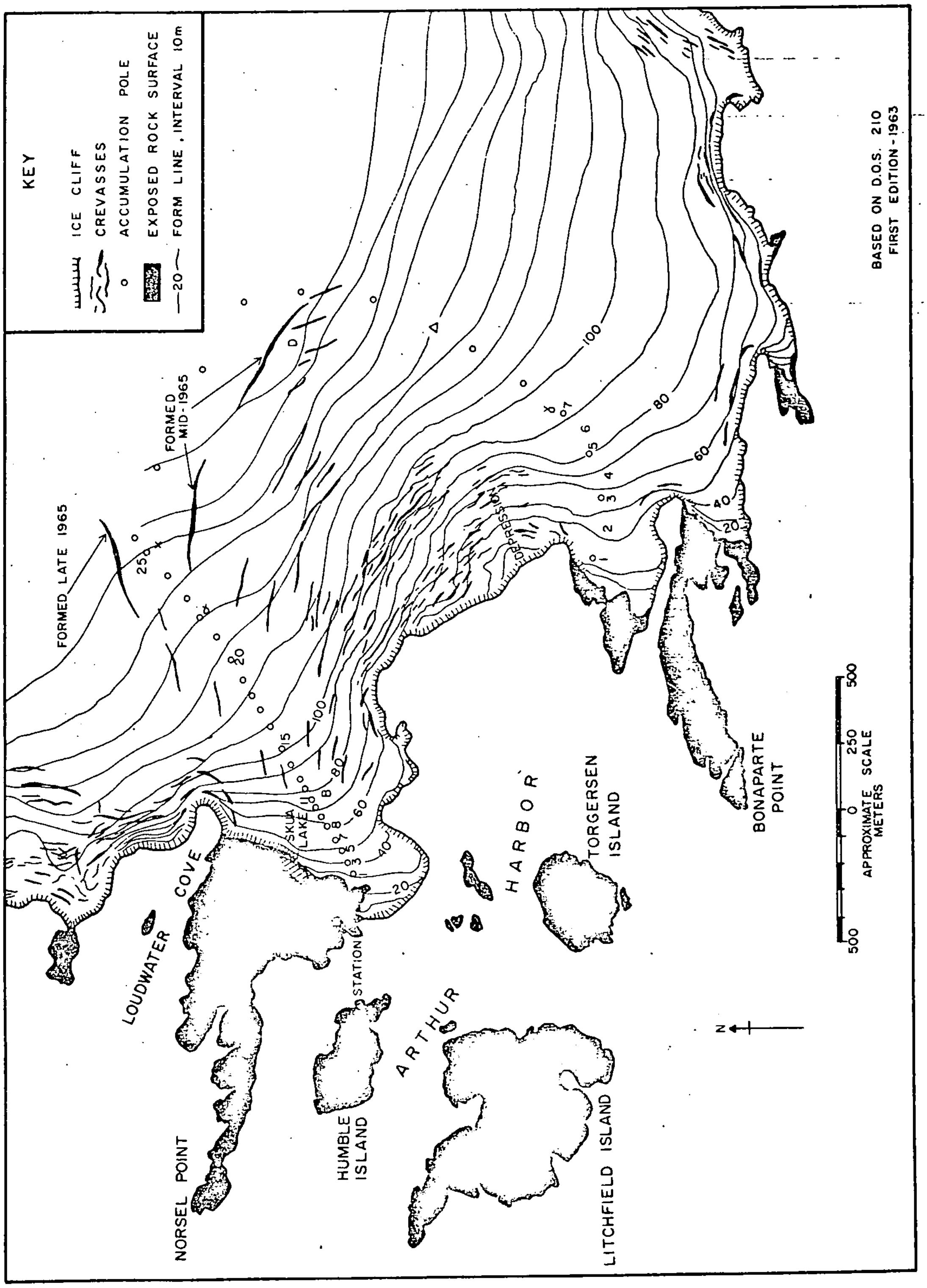


Fig. 44. Study area around Arthur Harbor.

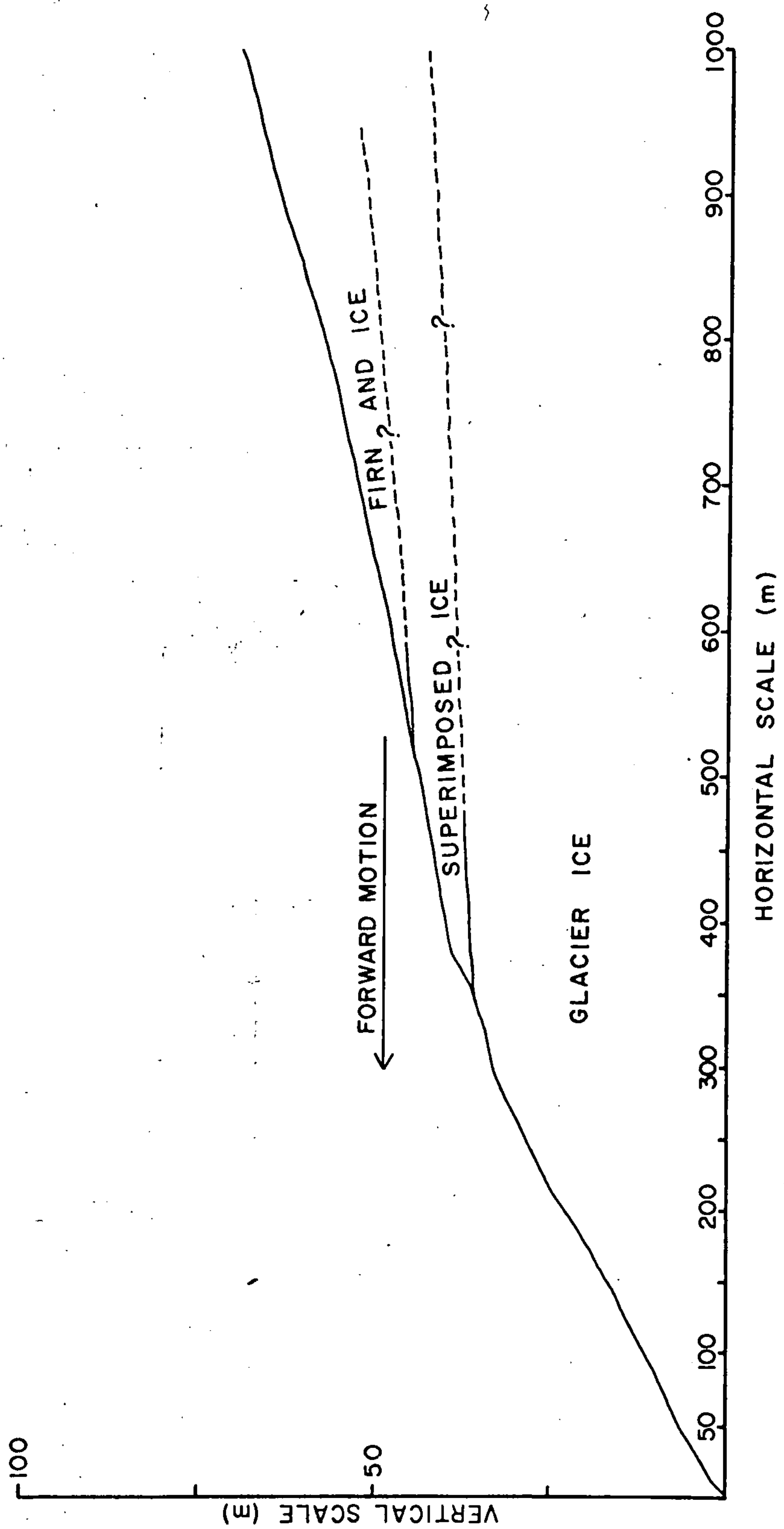


Fig. 45. Schematic diagram of Norsel Point Ramp. Longitudinal profile. V.E X5

negative annual budgets, an overall positive regime has been maintained resulting in the gradual buildup of the superimposed ice. Indeed, such a suggestion is supported by the observations between 1965 and 1968 when, during two summer seasons net loss resulted by surface melt, while in the remaining two summers a positive budget existed. Observations over a long period of time might reveal a slightly higher frequency of positive budget years.

The suggestion of a long-term positive regime for the (land-based parts) of the Marr Ice Piedmont accords with that of Bryan (1965) for the Fuchs Ice Piedmont on Adelaide Island where "it appears probable that the regime of the ice piedmont is either slightly positive or in equilibrium" (p. 62). Sadler (1968) came to similar conclusions from studies on the ice caps of Argentine Islands. The Anvers Island study also confirms the inaccuracy of Fleming's (1940) observation that the "fringing glaciers" are temporary features, out of phase with the present climate, and will disappear in a few years. Bryan (1965) suggested that "temporary" deglaciation in the Marguerite Bay area was contemporaneous with long periods of negative regime of the Fuchs Ice Piedmont and that a shift in the depression tracks over the Antarctic Peninsula had resulted in increased precipitation over Adelaide Island and the restoration of positive regime conditions. At Anvers Island, there is only circumstantial evidence to support the idea of a previously negative regime and this is a question which warrants further investigation. It is possible therefore, that Anvers Island lies within the supposed area of recent climatic and precipitation change.

CHAPTER NINE

ENGLACIAL STUDIES I: SNOW FACIES ZONATION

INTRODUCTION

A knowledge of the physical conditions within a glacier is fundamental to a complete understanding of the glacier's characteristics and behavior and is a prime prerequisite to detailed glaciological investigation and analysis. Therefore, investigations of the firn and ice on Anvers Island to shallow depth formed an important part of the overall study program.

Stratification of the firn cover, as revealed by snow pit excavations and ice cores, results from variations in the conditions of deposition and subsequent change. Physical changes within the firn cover with time can be directly observed and the causes of these changes often deduced. However, though differences between firn layers are frequently easy to see or feel, sometimes they are only detected by measured variations in density, hardness and grain size. Stratigraphic work in firn, therefore, consists of identifying and describing these variations in a given layered sequence and if possible, extending recognisable features laterally by correlation with other measured sections. The primary problem however, is to determine the conditions which prevail at present in the glacier.

The identification of annual units of accumulation is one of the most significant results derived from the study of firn stratigraphy and may allow an extension of accumulation records, back in time, often with reasonable certainty.

It is also possible to recognise and describe physical differences in the firn cover in terms of diagenetically produced facies, which in turn allows a greater resolution of glacier characteristics than does Ahlmann's more straightforward geophysical classification. In particular, a "facies classification" permits a subdivision of large glaciers which span the entire range of environments from temperate to polar.

The objectives of the englacial studies of the Marr Ice Piedmont were, therefore, to obtain information on the division of the ice cover into diagenetically produced facies and the magnitude and variation of past accumulation. Concurrently, information on the internal temperature distribution, albeit to shallow depth, was also obtained and is discussed below in Chapter 10.

THE CONCEPT OF DIAGENETICALLY PRODUCED FACIES

The concept of facies as applied to firn stratigraphy became firmly established after the work of Benson (1959), who formulated several basic definitions which have since been almost universally employed. Reference to these definitions is made in this thesis.

The ablation facies extends from the edge of the glacier to the firn line. The firn line is the highest elevation to which the snow cover recedes during the melt season. In this facies the accumulation of the current budget year is removed by ablation to reveal glacier ice or the accumulation of the previous budget year.

The soaked facies. When a mass of snow is at 0°C and wet throughout, it is said to be water-saturated or "soaked". The soaked facies is defined as that in which wetting reaches to the level of the previous year's surface. It extends from the firn line to the uppermost limit of complete wetting, the saturation line. The saturation line therefore, is the maximum elevation at which the 0° C isotherm penetrates to the melt surface of the previous year. Ice temperatures at a depth of 10 to 12 meters below the surface often reflect the mean annual air temperature above the saturation line. Below the saturation line, this is not the case.

The percolation facies occurs when the degree of wetting is not sufficiently strong to penetrate to the previous year's surface. Localised percolation in this facies permits the downward penetration of melt water through part of the current year's accumulation. The process can occur in snow with negative temperature and has been observed by Benson (1959) to be taking place in snow with ambient temperature of -7° C. to -18° C. The downward penetration is along restricted channels, slightly slushy when active, which may expand laterally on certain layers. Only the percolation channels are at the melting point and they refreeze to form ice glands, ice lenses and layers. Ice layers extend over large areas with only limited interruptions, while ice lenses can be seen to obviously pinch out laterally. Both layers and lenses are parallel to the strata. Ice glands are discordant pipe-like masses extending vertically downward and which occasionally spread laterally to form lenses and layers. When the firn temperature is not appreciably below 0° C. at the onset of melt, the penetration of melt water is more even and when refrozen produces only iced firn. The uppermost limit of the percolation facies is t he dry snow line above which lies the dry snow facies in which negligible summer melt and percolation occur (Benson 1959). From studies at altitudes in excess of 3,000 m. on the East Antarctic Plateau, this author prefers that by definition the dry snow facies should involve absolutely no transition of the snow cover to the mobile liquid phase and that radiation crusts are the only acceptable "melt" features in the dry snow facies. On this basis, the dry snow facies has not been recognised on the Marr Ice Piedmont proper.

A frequently useful division in the lower parts of a glacier is the <u>superimposed ice facies</u> where net accumulation occurs only by the re-freezing of melt water on the old ice surface and which is separated from the <u>net ablation facies</u> by the <u>equilibrium line</u>. Thus the equilibrium line separates the area of net accumulation from the ablation zone and is therefore the most important datum line on a glacier since it divides the glacier into areas of net mass gain and net mass loss.

SNOW FACIES ZONATION

Most of the deep pit work was carried out during August 1965 when pits, one to three meters deep were excavated by the author at each of the 14 stations along the Main and Mountain lines. Messrs. Ahrsnbrak and Plummer followed one or two days behind taking a core through the bottom of the pit and then transporting the core sections to Palmer Station for later examination by the author. Regrettably, the cores taken from the stations between H and R were destroyed during a sudden rise in air temperature at Palmer Station in early September 1965 (cold storage was not then available at the station) and only at stations H and K were cores later recovered and adequately studied. Snow pits and shallow-depth cores were studied in the field at other times during 1965 and 1966 but no data of this kind were obtained by the 1967 working party.

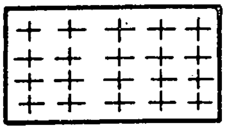
Standard SIPRE 500 cc. snow-sampling tubes were used for density measurements in the snow pits and the 7.6 cm. diameter ice cores were recovered with a standard SIPRE coring auger. An Ohaus Corporation triple beam balance, recording wieght to 0.1 gm., was used for all weight determinations.

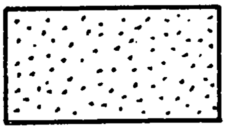
Studies above the Saturation Line

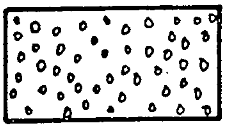
Station Little x:- The highest elevation at which a complete study was accomplished was at station Little x in the lee of the mountains (see figure 9, p. 23). In this area total snowfall during the year, as recorded by the stake networks, amounts to over 5 meters and is equivalent to some 230 cm. water. In August 1965 at this station, a pit was ecavated to a depth of 3.20 meters and which penetrated the entire snowfall since the height of the 1964-1965 summer. A continuous core was later recovered from a total depth of 13.00 meters below the surface. Core recovery at this point was almost perfect.

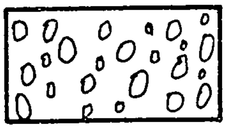
The observations at this station are shown in detail in Figure 46. The stratigraphy throughout the entire section is remarkably uncomplicated and the general absence of icy features indicates that this station lies close to, but not in, the dry snow facies. The homogeneity of the pit wall stratigraphy down to about two meters depth suggests fairly uniform conditions of deposition since about early April and relatively little subsequent winter-time diagenesis. However, it must

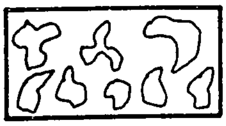
KEY TO PIT STRATIGRAPHY DIAGRAMS

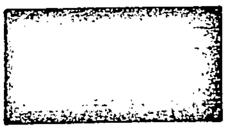

NEW UNDISTURBED SNOW

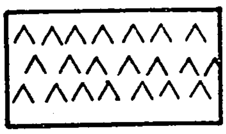

VERY FINE OR FINE-GRAINED FIRN (<1mm)

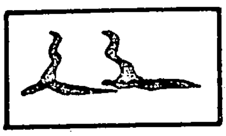

MEDIUM-GRAINED FIRN (1-2mm)


COARSE-GRAINED FIRN (2-3mm)


VERY COARSE-GRAINED FIRN (3-5mm)


SOLID ICE


RIME LAYERS


ICE GLANDS & LENSES

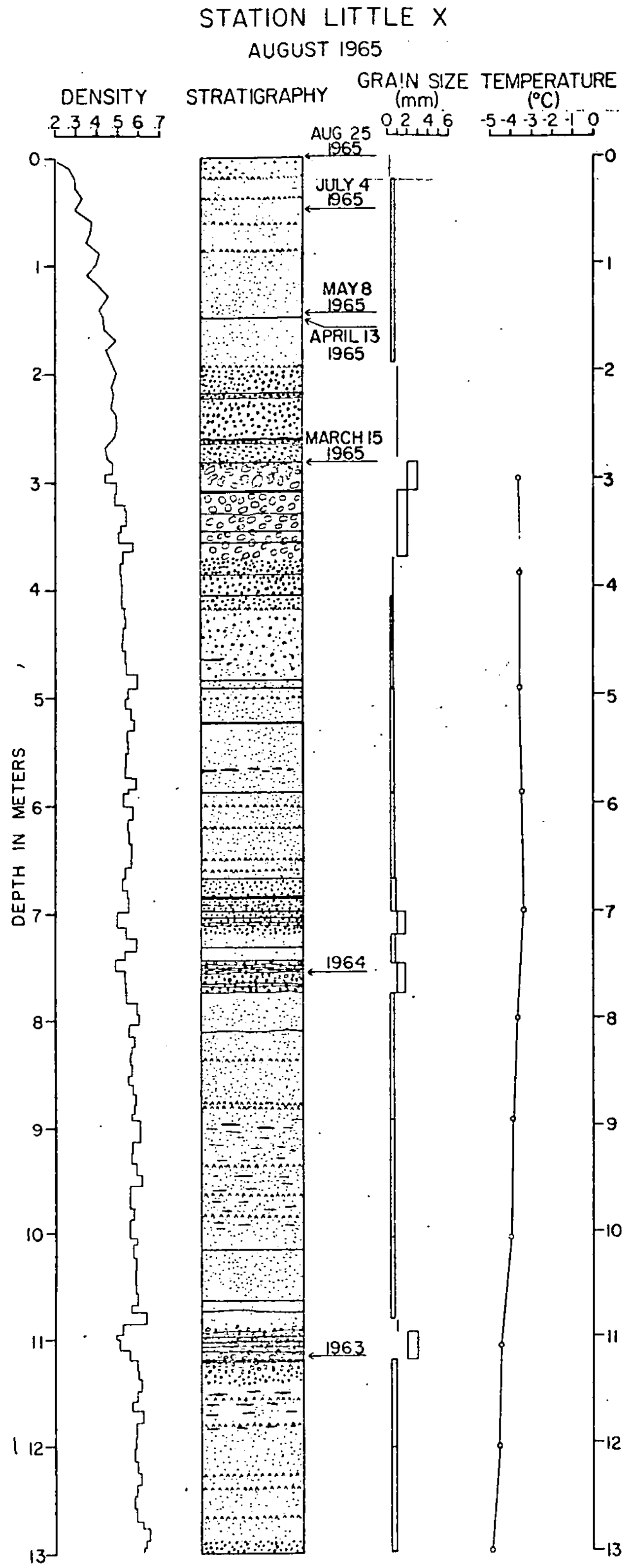


Fig. 46. Firn stratigraphy at station Little x.

be borne in mind that the excavation was made close to the time of maximum stability and that increasing stability would have taken place as this snow cover was laid down since the height of the previous summer.

The onset of real stability appears to have been about early April and the only significant features seen in the pit wall were the buried bands of surface rime and a thin ice crust which was at the surface on April 13. From the field records, this crust can be related to heavy fog prevailing over much of the piedmont on that date. The lower parts of the pit stratigraphy record some diagenesis of the snow cover and the conditions of increasing instability can be traced towards the height of the 1964-65 summer at a depth of approximately 3.5 m. This summer surface of 1964-65 is interpreted to be at about 3.5 m. depth on the evidence of the relatively large grain size of 2.0-2.5 mm.

On the same basis, the 1963-64 summer surface is interpreted at about 7.5 meters depth with the 1962-63 surface at about 11.25 meters depth.

Within the limits of these surfaces, there appears to be a slight variation from one year to the next in the degree of diagenesis that has taken place. In the firn of the 1964 winter there is evidence of diagenesis in the early part of the winter and in the latter part but it is not notably severe. This evidence is not apparent in the 1963 winter firn and suggests colder conditions throughout that year than in 1964. The stratigraphy also suggests that the period of maximum instability and diagenetic change was prolonged and relatively severe during the 1964-65 summer, whereas in the 1962-63 summer it was short and relatively ineffective.

The presence of the large grains in the summer surface firn and the existance of the small icy features in the stratigraphy prevent the inclusion of this station in the dry snow zone. The is no evidence of actual downward percolation of surface meltwater in the form of ice glands but this station does represent the upper reaches of the percolation facies.

<u>Station W:-</u> Station W lies at approximately 800 meters a.s.l. or about 20 meters below Little x but in spite of so small a difference in elevation there is, as can be seen from Figure 47, a noticeable difference in the stratigraphy. Icy features are generally more prominent and the severity of the 1964-65 summer is more pronounced. In certain parts of the core section there is evidence of downward percolation of meltwater in the form of small ice glands, which in places extend laterally as ice lenses and layers and indicate that this station lies in the true percolation facies.

The difference in iciness within so small an elevation change is probably the result of aspect relative to direct solar radiation in summer. In this region the mountains of the Achaean Range form an arc generally surrounding station Little x and this configuration tends to protect Little x from direct radiation - it lies in the shadow of the mountains for almost all the year. Station W however, is far enough removed from the base of the mountains to be out of the shadow for much of the year and consequently experiences considerably more of the relatively intense direct summer solar radiation which

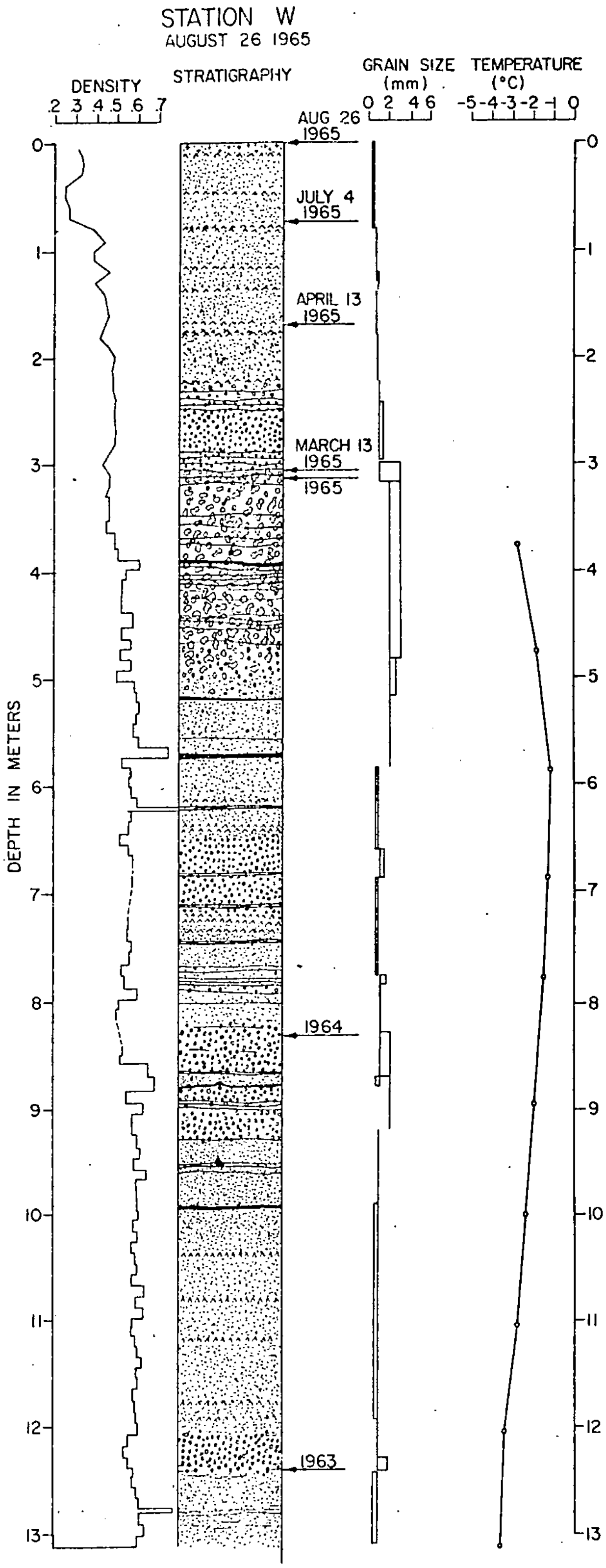


Fig. 47. Firn stratigraphy at station W.

results in slightly more summer surface melting. In some years this is strong enough to cause downward movement of the meltwater.

Apart from the generally grater iciness of the firn at W, the stratigraphy at the two stations is basically the same. The more severe diagenesis of the firn in the 1964-65 summer is very pronounced and the general lack of diagenetic change during the winter of 1963 is clearly evident, as is the diagenesis of the firn in the early and latter parts of the 1964 winter.

Though there is a slight variation in degree, these features are strongly evident in the stratigraphy from other stations(eg. station U, figure 48) moving outward from the base of the mountains but the stratigraphy becomes increasingly complex and more difficult to interpret and reflects the gradual change from the percolation facies to the true soaked facies zone.

Station S:- From the stratigraphy shown in Figure 49, it appears that station S (elevation 677 meters a.s.l. or about 150 meters below Little x) lies in a transition zone between the two facies but the recent history of the firn cover at this station compares well with that at the higher altitude stations.

At station S the onset of winter stability, from the evidence of the very fine-grained, relatively unaltered winter snow, appears to have taken place at more or less the same time as at Little x and W, but the intensity of the metamorphism in late March to early April 1965 apparently was greater than at higher elevations. The height of the summer of 1964-65 seems to have been later (around mid-March) than

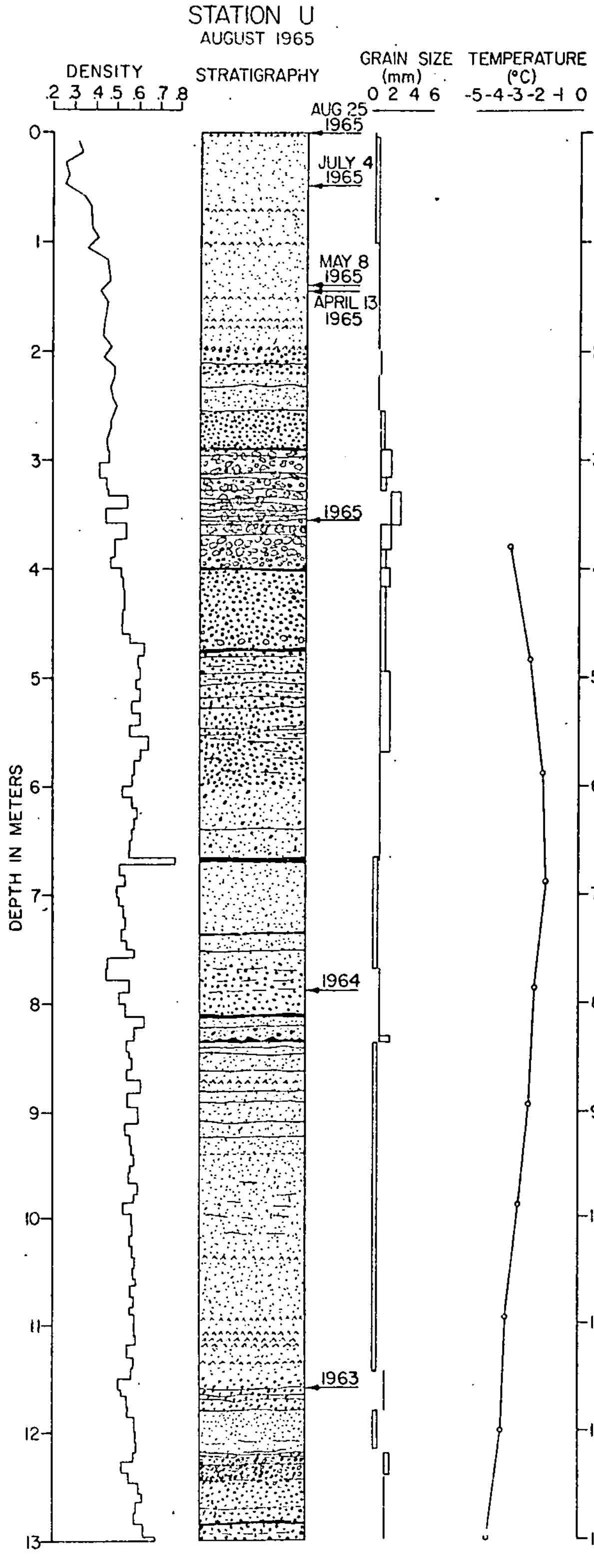


Fig. 48. Firn stratigraphy at station U.

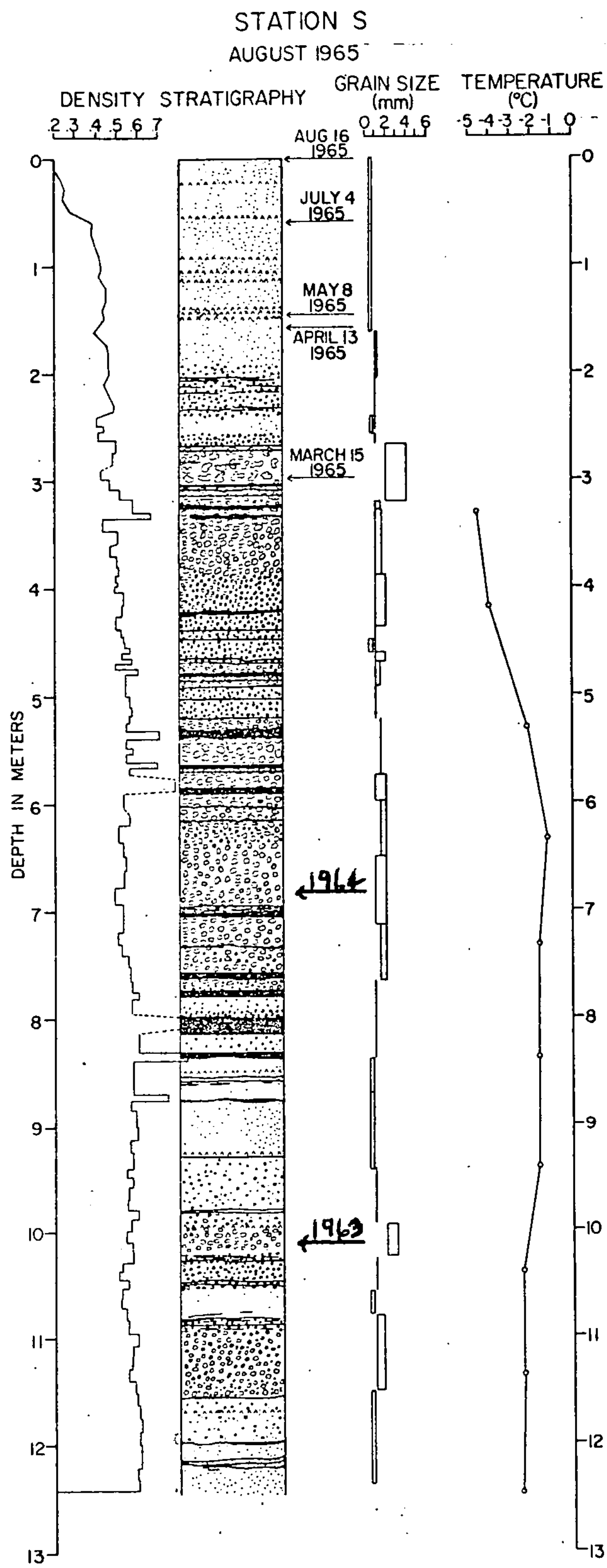


Fig. 49. Firn stratigraphy at station S.

at Little x (probably about mid to late February). This may be a reflection of the difference in elevation but is probably also attributable to the relative effects of the altitude of the sun and the difference in receipt of solar radiation by the two stations.

The most remarkable feature of the stratigraphy at station S is that the entire 1964 firn cover has undergone marked diagenetic change. This is in contrast to W and Little x which showed significant change only in the early and latter parts of that year. Also significant is that the upper layers of the 1963 firn cover (representing the latter part of that winter) have been severely changed and contain fairly massive layers of ice, again in contrast to the higher altitude stations. In this particular core no ice glands were seen to have been feeding into these ice layers but their origin probably is the result of pronounced percolation of the 1963-64 summer meltwater. In fact, the degree of metamorphism during that summer prevents a conclusive identification of the 1963-64 summer surface and that shown in Figure 49 is based not only on grain size but on data subsequently obtained from the accumulation stake network.

Despite the uncertainty as to the actual position of the 1963-64 horizon, there can be little doubt, in the absence of fine grained, unaltered firn, that station S was situated in the soaked facies in 1964. The middle portion of the 1963 section of the core does contain about one meter thickness of unaltered or almost unaltered firn, indicating that in that year this station lay in the percolation, not the soaked facies zone. From this evidence, which generally spans a three-year period, and assuming that any diagenetic change in late 1965 did not penetrate to the depth of the early April surface, this station was in the soaked facies during one year but in the percolation facies for the other two. Thus the saturation line shows mobility from one year to another. Its general position however, is in the vicinity of the 600 meters a.s.l. contour.

Studies Below the Saturation Line

Station K:- There can be no doubt that station K is situated well into the soaked facies (figure 50). Throughout almost the entire core there is evidence of severe melting and metamorphism of the firn and the stratigraphy is complex. There is however, a half-meter thickness of relatively fine grained and unaltered firn at 2.5 to 3.0 meters depth which is puzzling. Grain size in this piece was generally 1 mm. and the density varied between 0.514 g/cm^3 and 0.548 g/cm^3 but it was not particularly icy. Above, the grain size was 4 mm. and below, it was 2 mm. Its origin is not clear, but it may result from being protected from percolating meltwater by the heavily iced firn above.

The pit stratigraphy indicates that real stability was not attained until about late May to early June and that there was considerable metamorphism of the firn up to that date. This probably reflects the comparatively low elevation of this station and the greater incidence of rain and sleet in this area.

Though the stratigraphy of the whole core is highly complex, it is possible to make a rough correlation with the stratigraphy at the higher elevation stations and the summer horizons have been tentatively

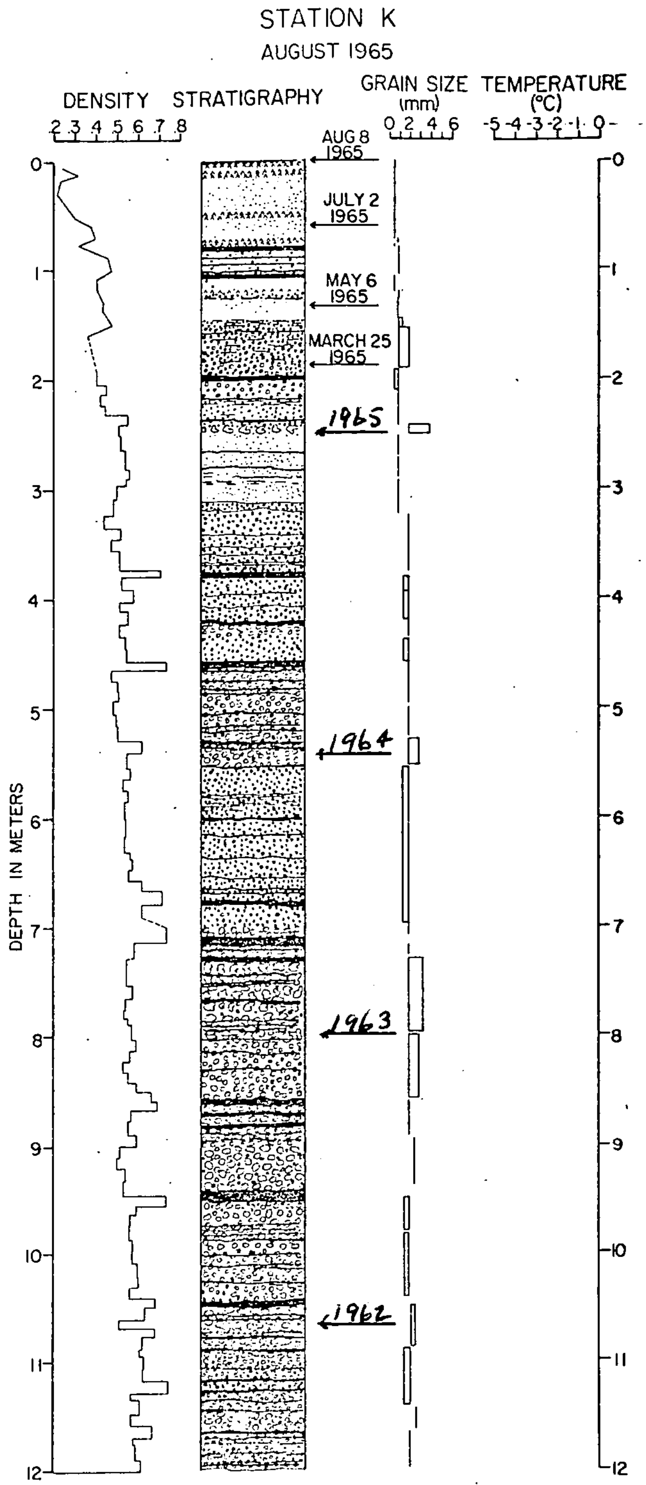


Fig. 50. Firn stratigraphy at station K.

identified on the basis of grain size and data from the accumulation stakes. There are some notable differences however, in the K stratigraphy.

The grain size data suggest that the 1964-65 summer was not particularly intense and very large grains were evident in only a small portion of the core. This is contrary to the evidence from higher elevations, which suggests that that summer was more intense than any other recorded in the core. The K station core also suggests that the summer of 1962-63 was very intense whereas at the higher stations it was the least significant in terms of diagenetic metamorphism. The possible reasons for these differences however, are not clear.

Station Big X:- This was the lowest altitude station at which a complete stratigraphic analysis was accomplished to significant depth. The stratigraphy is shown in Figure 51 and clearly shows an intensification of the soaked facies characteristics evident higher up at station K. Throughout the section there has been severe alteration of the firn and a buildup of massive ice features. In places the development of nodular ice masses, with grains up to 5 mm. diameter, was observed. The results of percolating surface meltwater are more strongly evident in this core than in the one from station K and there is more evidence also of surface leaching which is reflected in the density curve. The leached surfaces, which typically are associated with large and very large grain sizes, are porous and friable with densities of about 0.4 to 0.5 g/cm^3. The meltwater derived from these surfaces forms the ice layers below and raises the density to about 0.7 to 0.85 g/cm^3.

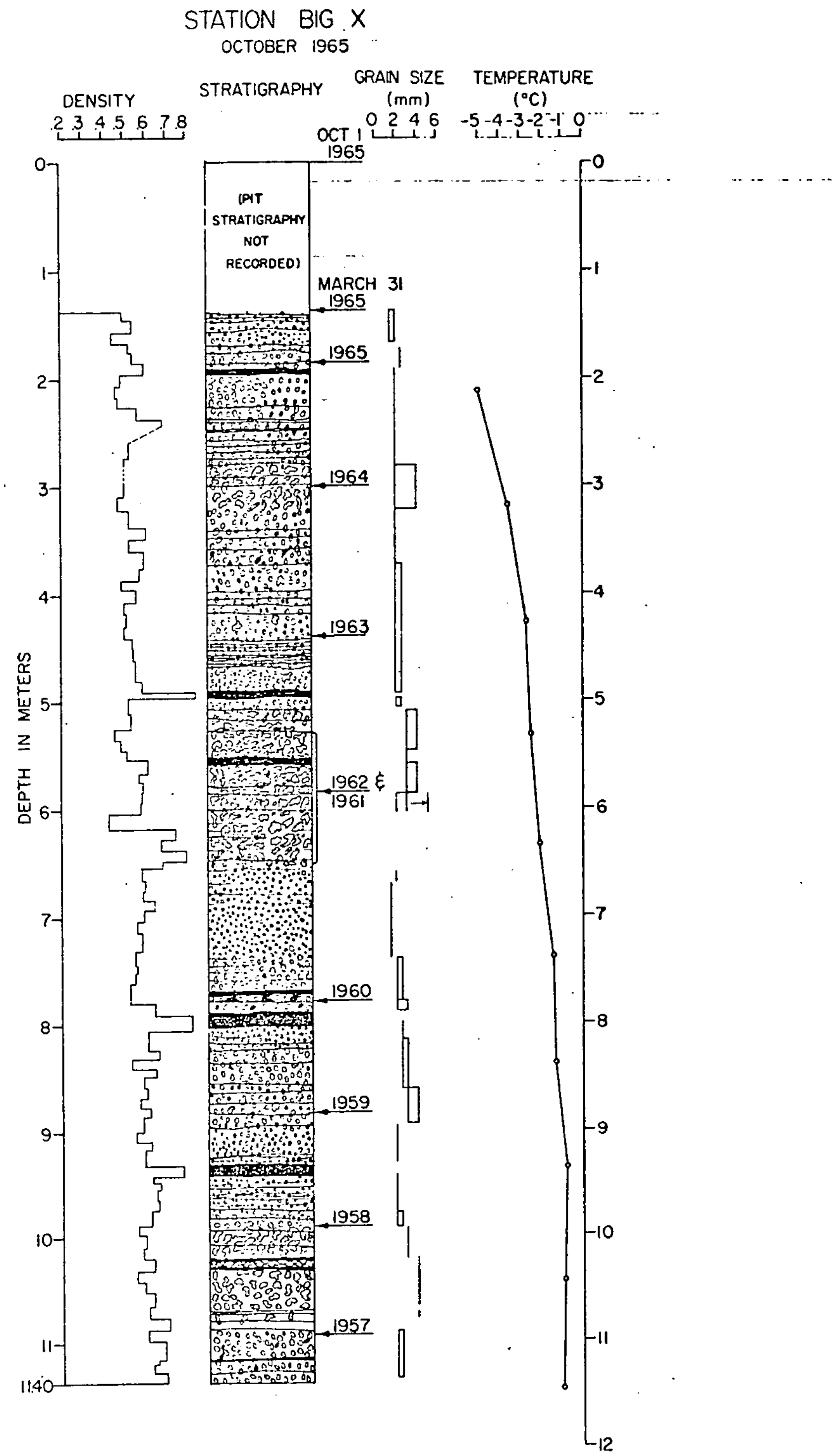


Fig. 51. Firn stratigraphy at station Big X.

Past Accumulation

The principal criterion used in interpreting the annual horizons is grain size, based on the premise that the relatively intense solar radiation and higher air temperatures in summer allows rapid grain growth at the time of minimum stability. Data from the accumulation stakes has also been used to determine the general depth at which a horizon should be, and at station Big X, the variations in the density were taken into account. The horizons shown in Figures 46 through 51 therefore, are interpreted as late February to early March.

On the basis of this interpretation, accumulation in the previous two years at stations Little x, W, U and S, and the previous three years at K have been calculated. At station Big X annual rates of accumulation are comparatively much lower and the ten meter core penetrated an eight-year period. The results are set out below. Column 2 shows the depth range of the year's accumulation, column 3 is the average ice density (g/cm^3) within the depth range and column 4 is the calculated total accumulation in g/cm^2.

	2	3	4
Little x			
1964	310 cm - 760 cm	0.540	243.0
1963	760 cm -1120 cm	0.564	203.0
Station W			
1964	320 cm - 830 cm	0.541	275.9
1963	830 cm -1240 cm	0.583	239.0
Station U			
1964	360 cm - 790 cm	0.556	239.1
1963	790 cm -1160 cm	0.574	212.4
Station S			
1964	270 cm - 680 cm	0.544	223.0
1963	680 cm -1010 cm	0.626	206.6

Station K

1964	240 cm - 540 cm	0.528	158.7
1963	540 cm - 800 cm	0.576	149.8
1962	800 cm -1070 cm	0.590	159.3

Station Big X

1964	190 cm - 300 cm	0.531	58.4	
1963	300 cm - 440 cm	0.537	75.2	
1962	440 cm - 585 cm	0.556	80.6	
1961	585 cm - ? cm)	0.604	117.8	(58.9
1960	? cm - 780 cm)			(58.9
1959	780 cm - 880 cm	0.614	61.4	
1958	880 cm - 990 cm	0.626	68.9	
1957	990 cm -1090 cm	0.624	62.4	

Comments

In any stratigraphic interpretation there are likely to be errors and uncertainties in deducing annual horizons. At the four high altitude stations it appears from this interpretation, that the 1963 accumulation was significantly less than in the following year. This could be a real difference or it could reflect a faulty interpretation. From an analysis of contemporary accumulation records discussed in Chapter 4 above, annual variation in accumulation rates, of significant proportion, are concluded and on the basis of that record, this stratigraphic interpretation may well be sound. At station K however, there is no annual variation evident unless the 1963 horizon has been interpreted at too great a depth. This could well be the case and which would result in lower accumulation in 1963 and much higher rates in 1962. The record from Big X suggests that 1962 was a high accumulation year, but this record also indicates that 1963 was a higher accumulation year than 1964.

The stratigraphy at Big X however, is so complex that possible horizons span 30 cm. to 40 cm., a variation which significantly affects the calculated values of annual accumulation. This complexity is particularly evident between 5 and 6 meters depth where it is impossible to conclusively identify the true position of the 1960-61 and 1961-62 summer surfaces and the general horizon shown in Figure 51 is arbitrary. Despite uncertainties about specific horizons, there can be little doubt that this core spans the period from March 1957 to March 1965 and the average annual accumulation for the 8-year period is 65.6 g/cm^2 which is very close to the observed value in 1965 (Chapter 4 above).

SUMMARY AND CONCLUSIONS

It is difficult to correlate these observations with those from other areas as very little information is available, on facies distribution in particular, from other parts of the Antarctic Peninsula. Nowhere does the subject appear to have been approached specifically by previous workers and much of the limited information there is emerges incidentally from determinations of past accumulation and mass balance.

From Koerner's (1961; 1964) reports, the glaciers and piedmonts of the Trinity Peninsula area have extensive ablation facies and the superimposed ice facies is often of significant size; in places of lee accumulation it becomes extensive. On Depot Glacier the equilibrium line lies at approximately 75 meters a.s.l. in the lee accumulation area due to the formation of superimposed ice from the freezing of melt water at the end of the ablation season. Kenny Glacier, flowing into Depot Glacier, has no lee accumulation area and the equilibrium line reaches to as high as 270 m. elevation. The climatic firn line in this area is at 150 meters but local lowering results from shade and avalanching snow.

From the Anvers Island study similar conclusions can be drawn. By definition the percolation facies exists only if surface melt water does not percolate to the depth of the previous summer surface and on the Marr Ice Piedmont it exists only by virtue of the massive annual accumulation rates rather than of temperature conditions alone. Under present conditions the saturation line lies at about 600 meters a.s.l. and all the snowfield above this elevation lies in the percolation facies. With lower annual accumulation rates over the piedmont, it is likely that the percolation facies would not exist at all, the saturation line would disappear and the firn line and equilibrium line would be at considerably higher elevation. In fact, both the equilibrium and firn lines on Anvers Island lie close together and occur, with slight fluctuations from year to year, between 60 and 100 meters a.s.l. When positive balance is recorded, as the ramp studies showed, neither line exists. On the Marr Ice Piedmont the soaked facies is by far the largest but there is a significant percolation facies above 600 m. elevation. The net ablation facies, when it exists, is small and is probably restricted to the small coastal ice ramps and though information on the superimposed ice facies is limited, it is known to exist on the upper parts of the ramps.

Thus it is apparent that though there may be a general trend along the length of the Antarctic Peninsula, the prevailing distribution of facies and associated parameters is not simply determined on the basis of latitudinal effects alone. Local causes and conditions are important factors. On Deception Island, where atmospheric temperature conditions are not significantly different from those at Anvers Island (Burdecki 1957), the firn line and equilibrium line lie close together at about 250 meters a.s.l. (Orheim 1970a) but the winter balance is only 70-100 g/cm^2 (Orheim 1970b; Kläy and Orheim 1969) and there is a low albedo as a result of volcanic activity. Both the dry and the percolation facies are absent and the soaked facies reaches to the highest parts of the glaciers at 450 m. elevation.

On Livingston Island, about 65 km. northward in the same island group, the same facies zonation exists but the equilibrium line reaches down to 100-150 meter a.s.l., with a superimposed ice facies extending through about 20 m. of elevation (O. Orheim, personal communication). This is similar to the elevation of the firn line on Depot Glacier (Koerner 1964) to the east suggesting that in this area the elevation of the regional climatic firn and equilibrium line is about 150 meters. To the south, the ice piedmont between Hope Bay and Trepassey Bay (Koerner 1964) had a winter balance of only 0.3 m. (snow cover) resulting in an average elevation of 300 m. for the firn line, with the ablation and superimposed ice facies being dominant.

From pit digging and visual observations Koerner (1964) placed the firn line somewhere below 150 meters a.s.l. on the west coast of Trinity Peninsula, between 230 m. 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 meters a.s.l. on the east coast islands (p.29). On Russel East Glacier the firn line lies at about 230 m. elevation and the saturation line ranges between 610 and 760 meters a.s.l. On Louis-Phillipe Plateau ice existed in the pit stratigraphy at more than 1000 m. elevation and it can be concluded that it is unlikely that the dry snow facies exists anywhere in the Trinity Peninsula area. The variation in the elevation of the firn line in this area was attributed by Koerner (1964) to several factors including deflation by cold southwesterly winds which raises the height of the firn line in the Hope Bay area.

It is notable however, that the firn line is higher on the east coast than the west coast in this area. The reverse appears to the true to the south. No information is available from the eastern side of Anvers Island but on the eastern side of Adelaide Island, the firn line was placed at about 77 m. a.s.l. by Bryan (1965) and 190 m. a.s.l. on the western coast near Adelaide station. This is considerably higher than on Anvers Island and probably results from a lower accumulation rate at Adelaide Island; compare about 25 g/cm^2 on Adelaide Island with about 60 g/cm^2 at similar elevation on Anvers Island.

On the eastern side of the island the relatively low elevation of the firn line seems to reflect the lower temperatures of the Weddell Sea coast and possibly the dry snow facies exists at high elevation on that side of the island. It appears to be absent on the Fuchs Ice Piedmont proper as the dry snow line is tentatively set at 850 m. a.s.l. (Bryan 1965) and much of the piedmont lies between the saturation line (below 370 m. elevation) and the dry snow line and is therefore in the percolation facies. Thus the soaked facies is comparatively smaller than on the Marr Ice Piedmont and probably results from the more southerly and slightly colder position of the Fuchs Ice Piedmont; the mean annual air temperature at Adelaide station is -6.2° C. (Bryan 1968) compared with -3.3° C. at Palmer Station (see Appendix I, this thesis).

From the observations of Thomas (1963) and Sadler (1968) it appears that the ice caps of Argentine Islands lie exclusively below the saturation line and that the facies present are; soaked, superimposed ice and ablation, the low elevation of these ice caps being the primary cause.

A general trend can be tentatively discerned from these observations. From the north to the south of the Antarctic Peninsula there is a gradual lowering of the various boundary lines which can be related in part to air temperature factors but also to increasing rates of annual accumulation, which probably reach their maximum in the west central region near Palmer Station, towards the south. On the basis of currently available information alone, it appears that the dry snow facies does not exist on any of the glaciers and piedmonts of the western peninsula and may be restricted to the mountains of the eastern side. The ablation, superimposed ice and soaked facies exist throughout the western peninsula but the percolation facies, on the west coast, does not exist north of about 64° S. latitude.

The data are extremely limited and considerably more observations are needed before the overall distribution of facies can be explained.

CHAPTER TEN

ENGLACIAL STUDIES II: ICE TEMPERATURE

INTRODUCTION

The distribution of temperature within a glacial body is governed by the climate at the surface, the geothermal gradient of the bedrock over which the glacier flows, the rate of movement, the shear stress at the base and the rate of accumulation at the surface. Thus, the study of temperatures within a glacier can provide a great deal of information on the thermal properties of the firn and ice, the dynamics of the glacier and the climate at the surface.

On the Marr Ice Piedmont ice temperature observations were made at each of the velocity stations along the Main and Mountain lines in 1965 in conjunction with the snow pit and ice core studies. Electronic or thermo-electric instrumentation was not available and all ice temperatures were obtained from mercury-thallium liquid-in-glass thermometers graduated to 0.1° C. All the instruments used had previously been checked and calibrated by by the U.S. Bureau of Standards. No significant errors were reported by the Bureau and the temperatures reported below are direct field observations.

Temperatures recorded in the bore holes were obtained from insulated thermometers; the insulation consisted of approximately 1 cm. thick coating of candle tallow bound with medical adhesive tape. The thermometer was lowered to the newly drilled level, the hole was capped and allowed to stabilise for 45 to 60 minutes. Experiments at Palmer Station indicated that the lag time of the thermometers was 8 to 12 minutes. Temperatures obtained from the snow pit walls were recorded with uninsulated thermometers placed directly into the pit wall to the immersion level as the excavation progressed. They were allowed several minutes to stabilise and read in situ. The reading interval in the pits ranged from 5 cm. near the surface to 20 cm. near the floor. In the bore holes the interval was approximately 1 meter. Temperatures were obtained from depths up to 13 m. below the prevailing surface.

ICE TEMPERATURES

The observations are given in Table XXIV and are included in the stratigraphy diagrams (figures 46-51) but for emphasis are shown graphically to larger scale in Figure 52. The anamalous curve from station H results from the drill having been frozen in and subsequent cooling of the hole in the lower ambient air temperature. It should be discounted, though the lowest reading from 11 meters depth may be credible.

These are typical winter temperature curves with the prevailing winter cold wave much in evidence. In detail they differ subtly from one another and indicate a gradual change in temperature regime with elevation. They all show that by August the winter cold wave had penetrated to about 6 or 7 meters depth. The warming effect of storms and the subsequent recovery of the temperature profiles are evident in some of the temperature curves in the upper firn. The drilling operation was interrupted on two occasions by storms; on August 12 and 13, when at Palmer Station, the temperature changed from -18.6 °C on the 10th to -6.1 °C on the 13th and back to -18.4 °C on August 16. In the second

TABLE XXIV

FIRN AND ICE TEMPERATURES: MAIN AND MOUNTAIN LINE STATIONS

Depth(cm)	*C	Depth(cm)	*C	Depth(cm)	*C
BIG X		E		K	
74	-5.0	195	-5.4	226	-3.4
180	-3.6	306	-3.2	343	-2.2
278	-2.7	399	-2.4	466	-1.2
393	-2.5	511	-1.4	582	-2.2
497	-2.1	610	-0.8	712	-1.0
600	-1.4	702	-0.6	790	-0.9
700	-1.3	796	-0.8	913	-0.8
798	-0.8	924	-0.5	1018	-0.7
905	-0.9	1032	-0.6	1150	-0.8
1008	-1.0				

Depth(cm)	*C	Depth(cm)	*C	Depth(cm)	*C
L		M		N	
207	-3.2	270	-4.8	300	-4.8
323	-2.7	392	-3.2	403	-3.8
436	-2.4	525	-1.8	510	-2.2
555	-1.4	652	-1.1	610	-1.6
676	-1.2	755	-1.0	709	-1.2
790	-1.0	857	-0.8	813	-1.0
891	-0.8	965	-0.5	903	-1.2
1021	-0.6	1070	-0.5	1005	-0.8
1106	-1.2	1171	-0.7	1112	-1.3
				1214	-1.2

Depth(cm)	*C	Depth(cm)	*C	Depth(cm)	*C
P		R		S	
282	-2.0	393	-2.7	331	-4.6
389	-2.2	499	-2.1	417	-4.0
496	-2.0	603	-2.1	531	-2.2
593	-1.7	706	-1.6	636	-1.2
694	-1.6	799	-1.6	731	-1.6
789	-1.5	912	-1.4	839	-1.6
888	-1.5	1012	-1.5	941	-1.6
995	-1.6	1111	-1.6	1042	-2.4
1101	-1.2	1208	-1.0	1137	-2.4
1205	-1.3			1246	-2.5

TABLE XXIV (cont)

Depth(cm)	°C	Depth(cm)	°C	Depth(cm)	°C
T		U		V	
358	-2.4	380	-3.3	382	-2.6
483	-1.4	483	-2.3	476	-2.2
578	-1.1	588	-1.7	587	-2.3
673	-1.2	689	-1.6	686	-2.3
780	-1.4	786	-2.1	793	-2.5
887	-1.6	893	-2.4	894	-3.0
1000	-1.6	989	-2.9	991	-3.4
1091	-2.4	1095	-3.5	1097	-3.8
1187	-3.6	1200	-3.7	1201	-4.1
1294	-4.1	1299	-4.4	1299	-4.6

Depth(cm)	°C	Depth(cm)	°C
W		LITTLE x	
376	-2.8	387	-3.6
475	-1.9	493	-3.6
587	-1.2	589	-3.5
687	-1.4	699	-3.4
790	-1.6	800	-3.7
897	-2.0	894	-3.9
1001	-2.4	1006	-4.0
1105	-2.8	1109	-4.5
1205	-3.4	1204	-4.6
1312	-3.6	1300	-4.9

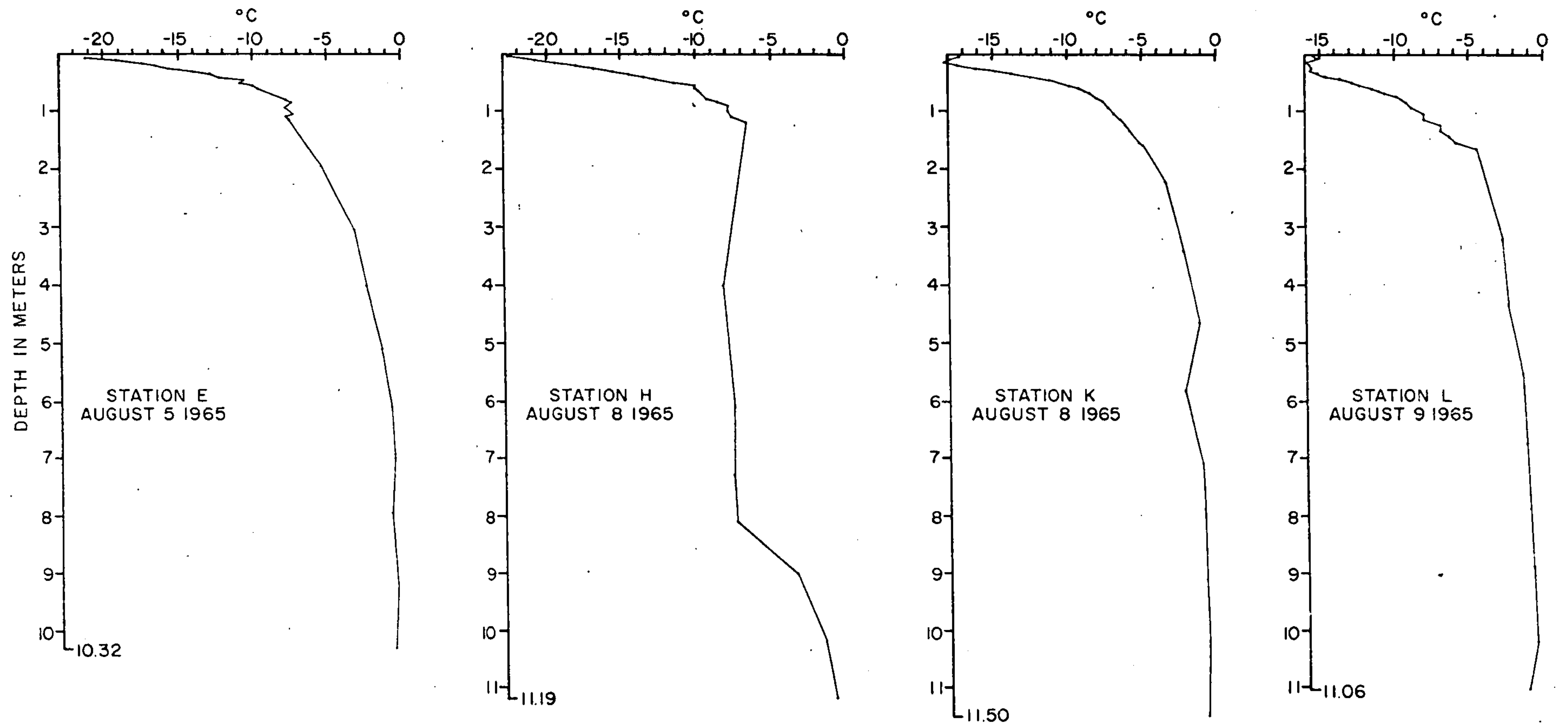


Fig. 52. Firn temperature profiles, Main and Mountain line stations.

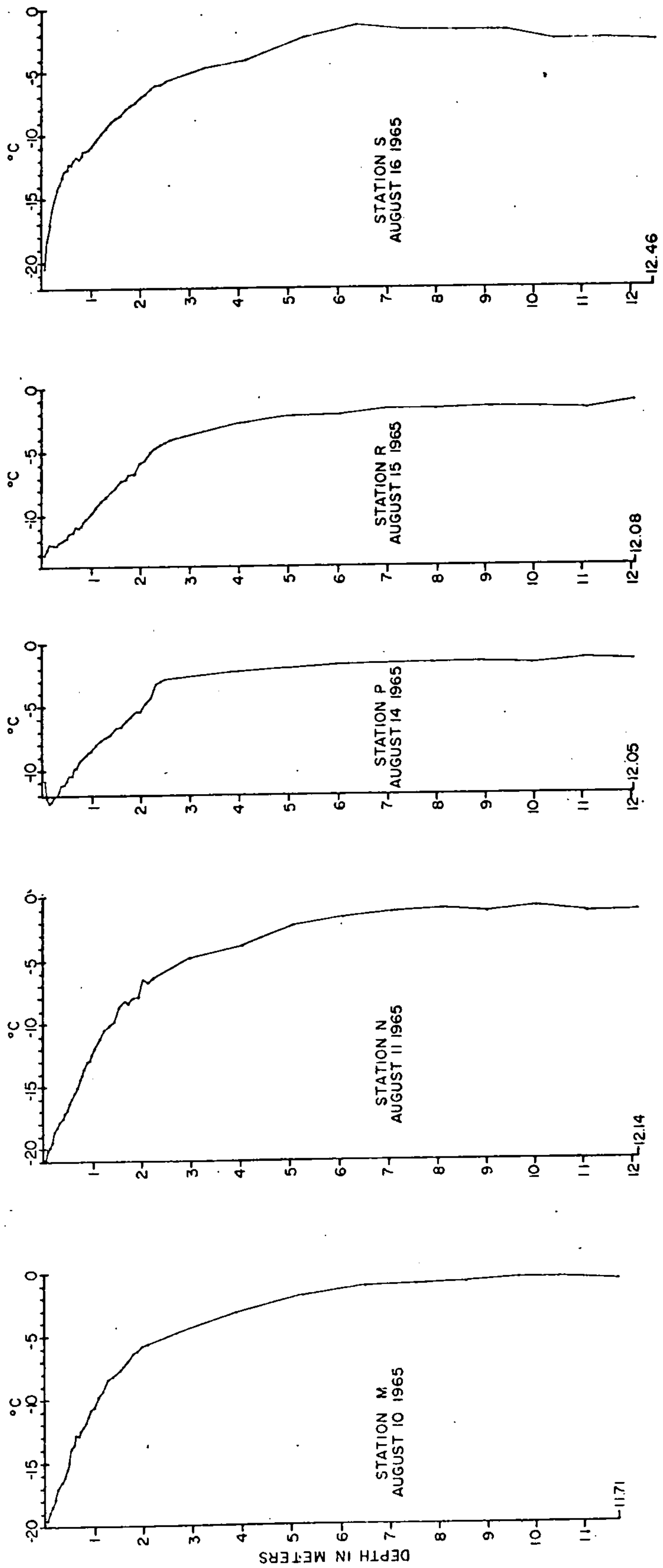


Fig. 52 (continued)

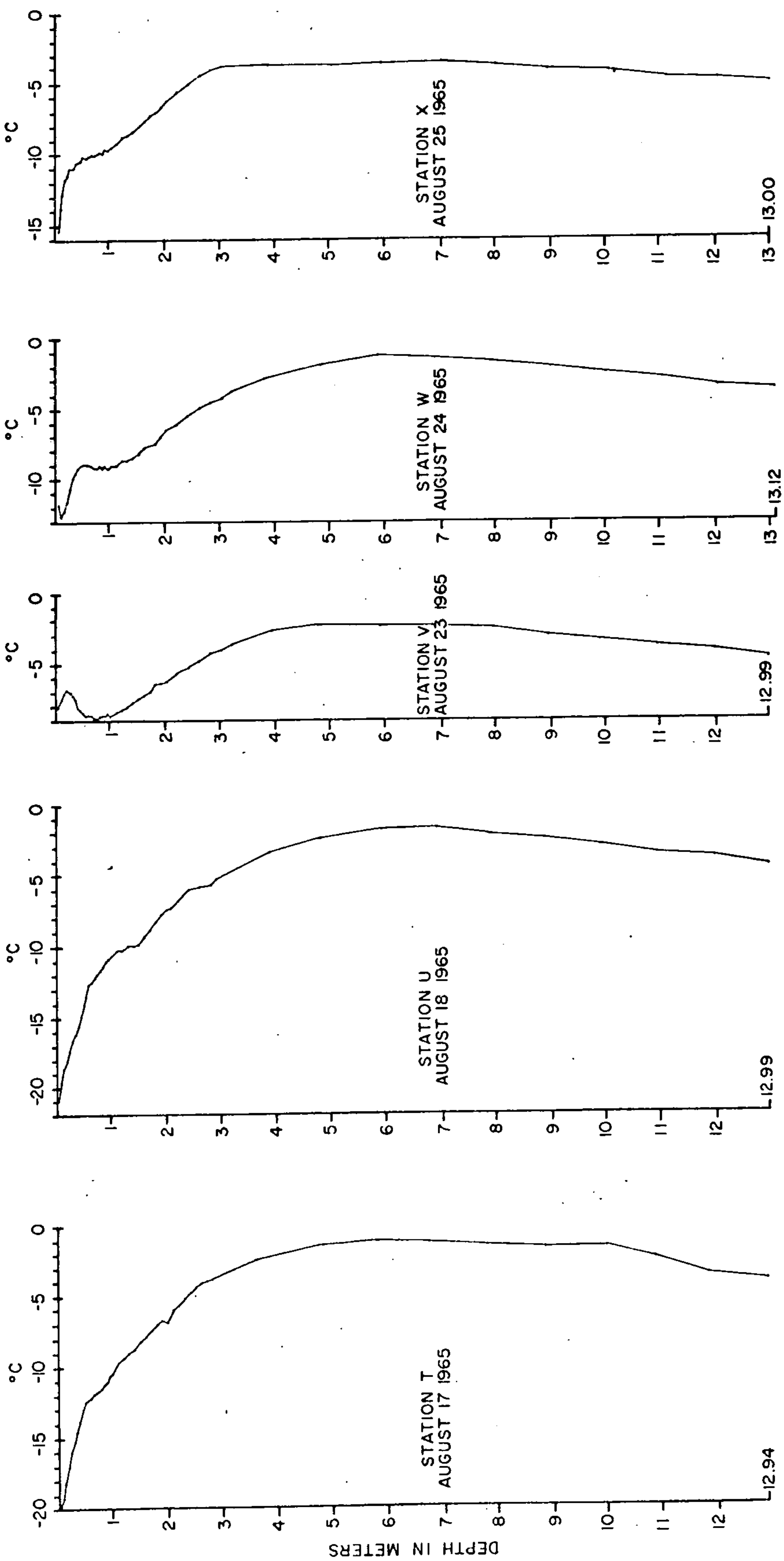


Fig. 52 (continued)

storm which halted work from August 19 to 22, the temperature jumped from -20.4 °C on the 17th to plus 1.2 °C on August 21 and back to -5.5 °C on the 25th.

The upper firn temperatures at stations N through S and stations T through Little x reflect these conditions when the negative temperatures were supressed by higher air temperatures and the deposition of warm fresh snow during the storm. The restoration of lower air temperatures under clear sky after the storm, resulted in the slow recovery of the original firn temperature curves(the anatomy of this particular storm is discussed in Appendix I). The pattern of recovery suggests a temperature penetration rate of about 0.5 m. per day in the upper layers, similar to the value determined by Bryan (1968) on the Fuchs Ice Piedmont of Adelaide Island.

Temperatures Below the Saturation Line

Below the saturation line surface melt in summer occurs and the downward percolation of melt water is sufficient to completely wet the firn and the descending water can transport heat to depth. On refreezing, the latent heat of fusion contributes to the raising of the ice temperature at depth. The temperature curves from the lower profile stations reflect this process. The prevailing winter cold wave is evident in the rapid increase of temperature with depth (ie. a positive temperature gradient) within the 1965 winter accumulation, from about -20 °C, the prevailing air temperature, near the surface to about -5 °C at a depth of 1.5 to 2.0 m. which was the excavated depth of the snow pits. Below this depth to about 6 or 7 meters, the gradient is less steep and suggests that the conductive properties of the recent, relatively unaltered winter snow is greater than the iced firn below. At depth below about 7 m. the temperature is generally constant at slightly below 0 °C.

In this area the previous winter's cold wave has been completely obliterated as a result of high summer air temperatures and particularly the heating effect of the percolating surface meltwater. Evidence of this process extends to the region of stations P and R or approximately 500 to 600 m. elevation. The 10 to 12 m. ice temperatures therefore, do not reflect the mean annual air temperature in this area.

Temperatures Above the Saturation Line

At elevations in excess of 500-600 m. ice temperatures at depths below 6 or 7 m. decrease slightly with depth (a negative temperature gradient) and this can be inferred as the remains of the previous year's (1964) winter cold wave. In this area, though some surface melt occurs, with percolation to shallow depth, it is not sufficient to completely soak the firn. The heating effect is too weak to remove the old winter cold wave and the absorption of solar radiation in summer and the effects of thermal radiation and conduction are the main processes by which the cold wave is reduced and stabilisation attained. In this area the deeper temperatures near the bottom of the holes should be close to the mean annual air temperature and are of particular interest.

Temperatures at 10 m. to 12 m. Depth

On a high polar glacier, where surface melt and percolation of meltwater are insignificant, or above the saturation line on other glacier types, the temperature of the firn at a depth of about 10 m. is approximately equal to the mean annual air temperature at the surface (Loewe 1956 p. 663). It could be reasonably expected therefore, that as a result of the atmospheric adiabatic lapse rate, the ice temperatures from the higher parts of the piedmont (supposedly above the saturation line as indicated by the ice core studies) should be considerably lower than those actually recorded.

The ice temperatures from approximately 10 m., 11 m. and 12 m. depth are shown in Figure 53. This plot of temperature versus elevation generally describe two intersecting straight lines. At lower elevations the line is approximately vertical while at higher elevations the line indicates a slow change of temperature with elevation. These are not statistical lines but free-hand approximate best fit lines to emphasise the trends.

The adiabatic lapse rate varies with temperature and humidity from a dry rate of approximately 1.0 °C per 100 meters to a moist lapse rate of about half this value at high temperature (Petterssen 1941 p. 53). With decreasing temperature, the moist lapse rate approaches the dry rate. The standard-atmosphere lapse rate, based on 15 °C air temperature and 1013.25 mb. atmospheric pressure, is 3.6 °F per 1000 feet (2 °C per 300 m.) or approximately 0.7 °C per 100 m. (Bowditch 1962 p. 794). Over Anvers Island, because of the relatively low air temperatures, the moist lapse rate should approach the dry lapse rate and should, and in fact do lie between 0.7 °C and about 1.0 °C per 100 meters (see Appendix I).

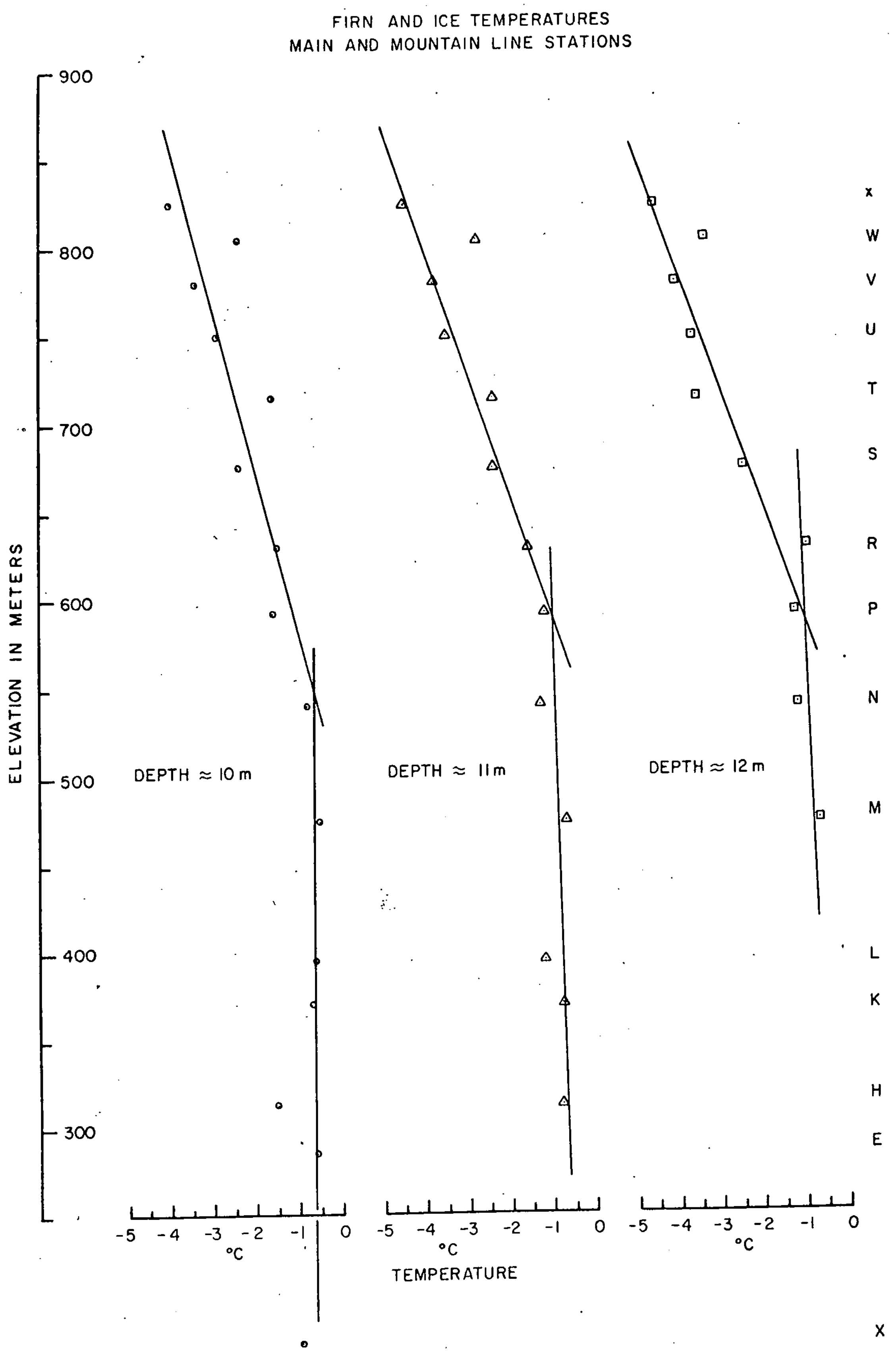

FIRN AND ICE TEMPERATURES
MAIN AND MOUNTAIN LINE STATIONS
ELEVATION IN METERS
900
800
700
600
500
400
300
DEPTH ≈ 10 m
DEPTH ≈ 11 m
DEPTH ≈ 12 m
x
W
V
U
T
S
R
P
N
M
L
K
H
E
X
-5 -4 -3 -2 -1 0
°C
-5 -4 -3 -2 -1 0
°C
-5 -4 -3 -2 -1 0
°C
TEMPERATURE

Fig. 53.

The mean annual air temperature at Palmer Station (at sea level for all practical purposes) was -3.6 °C in 1965 (11 months), -3.8 °C in 1966 and - 2.8 °C in 1967. The three-year average was -3.3 °C. Station Little x lies at 825 meters a.s.l. and, from considerataion of the adiabatic lapse rates, should experience a mean annual air temperature of between -12.7 °C (dry lapse rate) and -10.0 °C (standard atmosphere lapse rate), while the lower stations should experience temperatures slightly higher, in proportion to their altitudes. In no place do the temperatures plotted in Figure 53 reflect the expected mean annual air temperature. At station Little x the negative ice temperature is only about 40% of what would have been calculated.

It is improbable that heat is being imparted to the ice from the geothermal heat source or from ice movement at a sufficiently high rate to account for so large a difference in ice temperature. It must be concluded therefore, that if the adiabatic lapse rate recorded between sea level and 300 m. elevation (see Appendix I) is constant up to the base of the mountains, then the unexpectedly high ice temperatures are in fact the result of heat transfer by the small amounts of percolating summer meltwater and that nowhere on the Marr Ice Piedmont do the 10 m. to 12 m. deep ice temperatures reflect the mean annual air temperature. However, the generally constant deep temperatures below 500-600 m. elevation and their gradual decrease with elevation at higher altitudes suggest that the saturation line on the Marr Ice Piedmont lies in the vicinity of 500 to 600 meters a.s.l. and strengthens the conclusion made in chapter 9 above that the facies zonation of the piedmont is more a function of the accumulation rate than of air temperature conditions alone.

Temperatures in the Norsel Point Ramp

An elaborate study of ice temperatures and surface meteorology was originally conceived to follow the process of superimposed ice formation on the Norsel Point ramp. The project proved to be too large and complicated to be adequately pursued within the main work schedule and was never properly completed. However, considerable ice temperature data were obtained, some of which are discussed below.

Eight temperature measuring stations were established in April 1966. Each consisted of 17 copper-constantan thermocouples set at intervals ranging from 5 cm. near the surface, then 10 cm., 20 cm., 50 cm., 100 cm. and 200 cm. with increasing depth, with the lowest junction at 4.5 meters depth. Several junctions were suspended above the surface in a rack to record snow temperature as they gradually became buried. At three stations a copper-constantan thermohm was set at 10 m. depth. All temperatures were read with a Leeds and Northrup Direct Recording Temperature Indicator. The system was previously calibrated at Palmer Station using a distilled water/distilled water-ice mixture in a vacuum bottle, against the mercury-thallium liquid-in-glass thermometers. The errors detected in the thermocouples have been applied to the results below.

The observations from the three stations with 10 m. thermohms are shown in Figures 55, 56 and 57. The location of these stations is shown in Figure 54; site 1 was situated at ramp pole number 12,

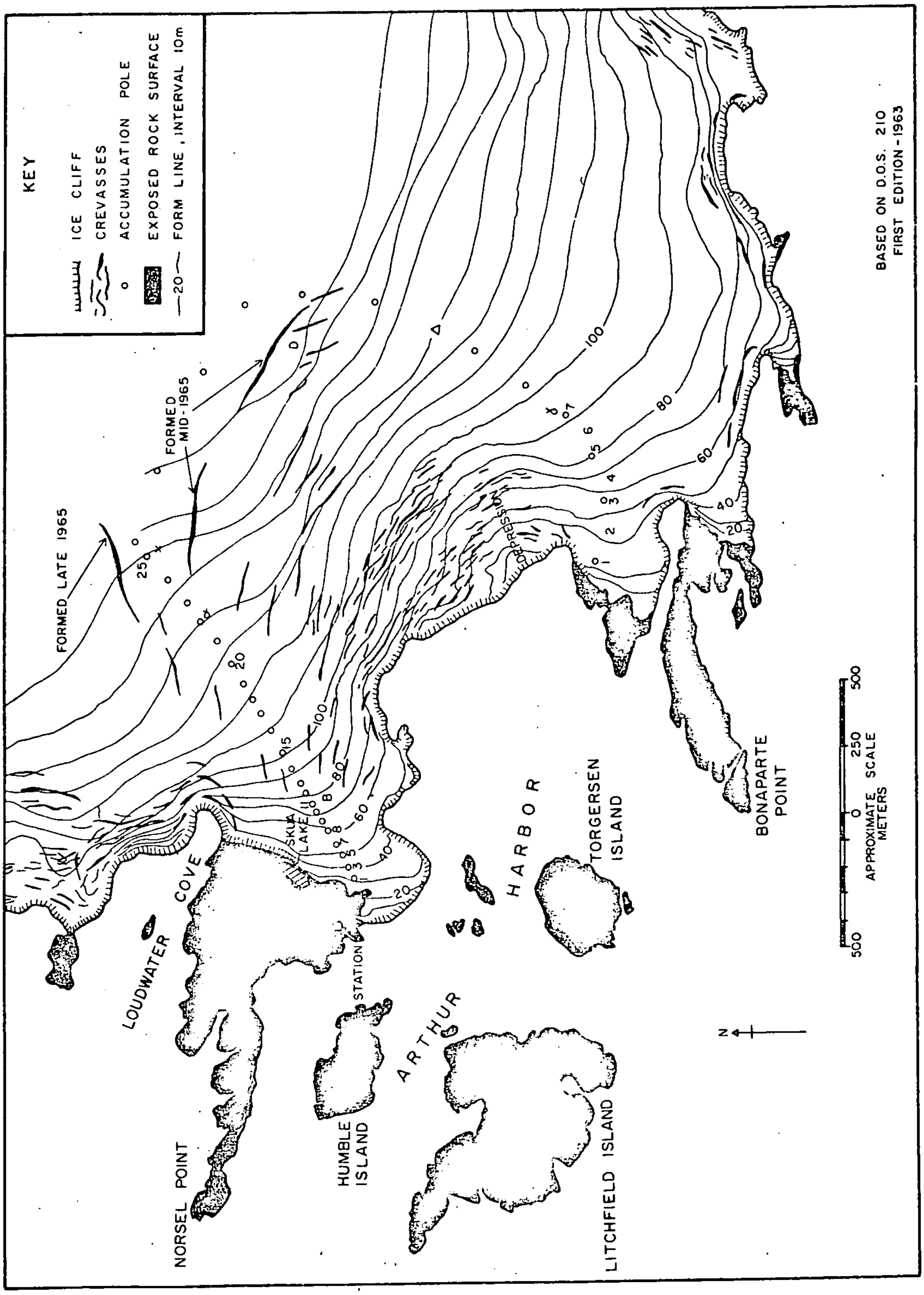


Fig. 54. Study area around Arthur Harbor.

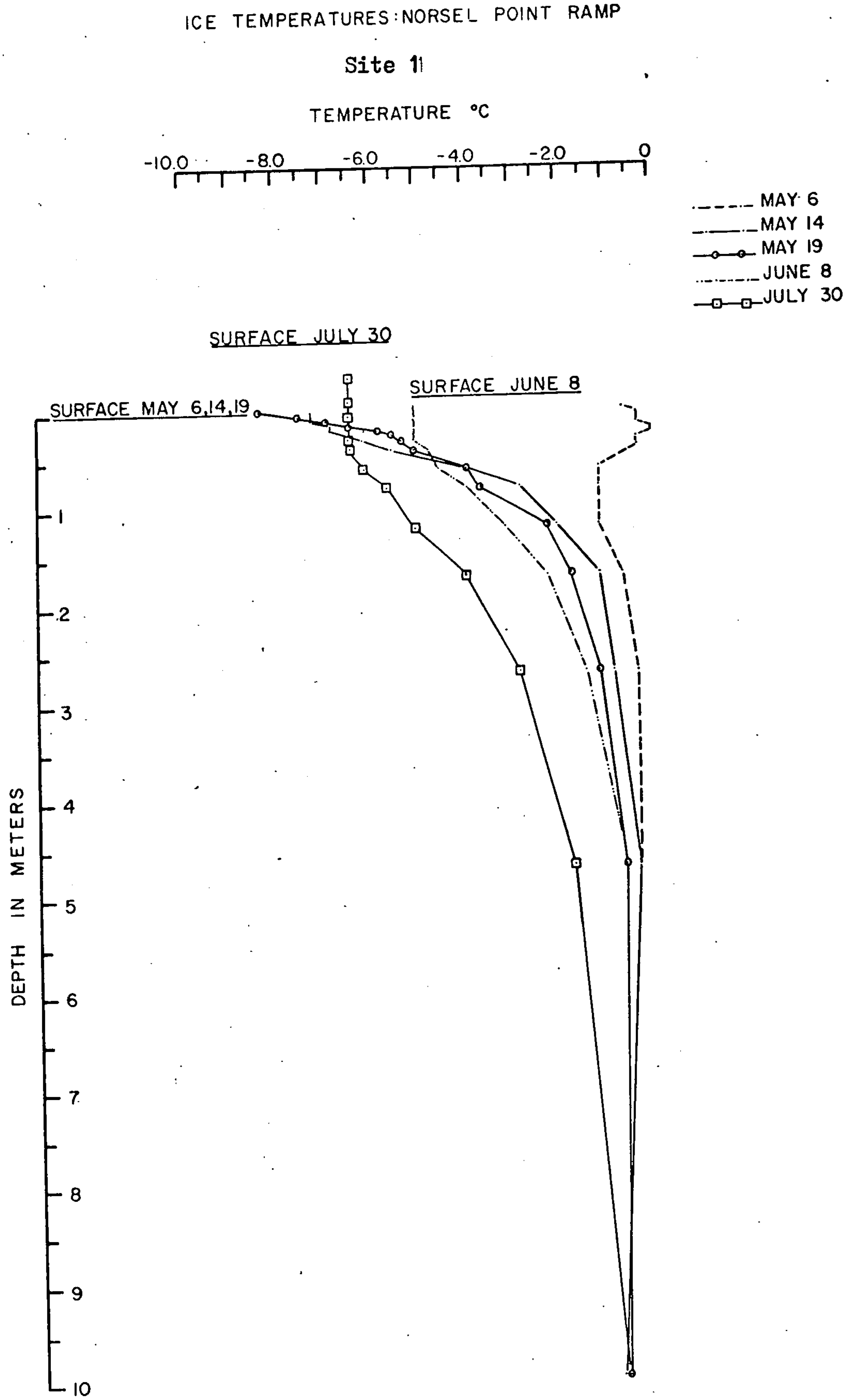

ICE TEMPERATURES: NORSEL POINT RAMP
Site 11
TEMPERATURE °C
-10.0
-8.0
-6.0
-4.0
-2.0
0
MAY 6
MAY 14
MAY 19
JUNE 8
JULY 30
SURFACE JULY 30
SURFACE JUNE 8
SURFACE MAY 6,14,19
DEPTH IN METERS
1
2
3
4
5
6
7
8
9
10

Fig. 55.

ICE TEMPERATURES NORSEL POINT RAMP

Site 1

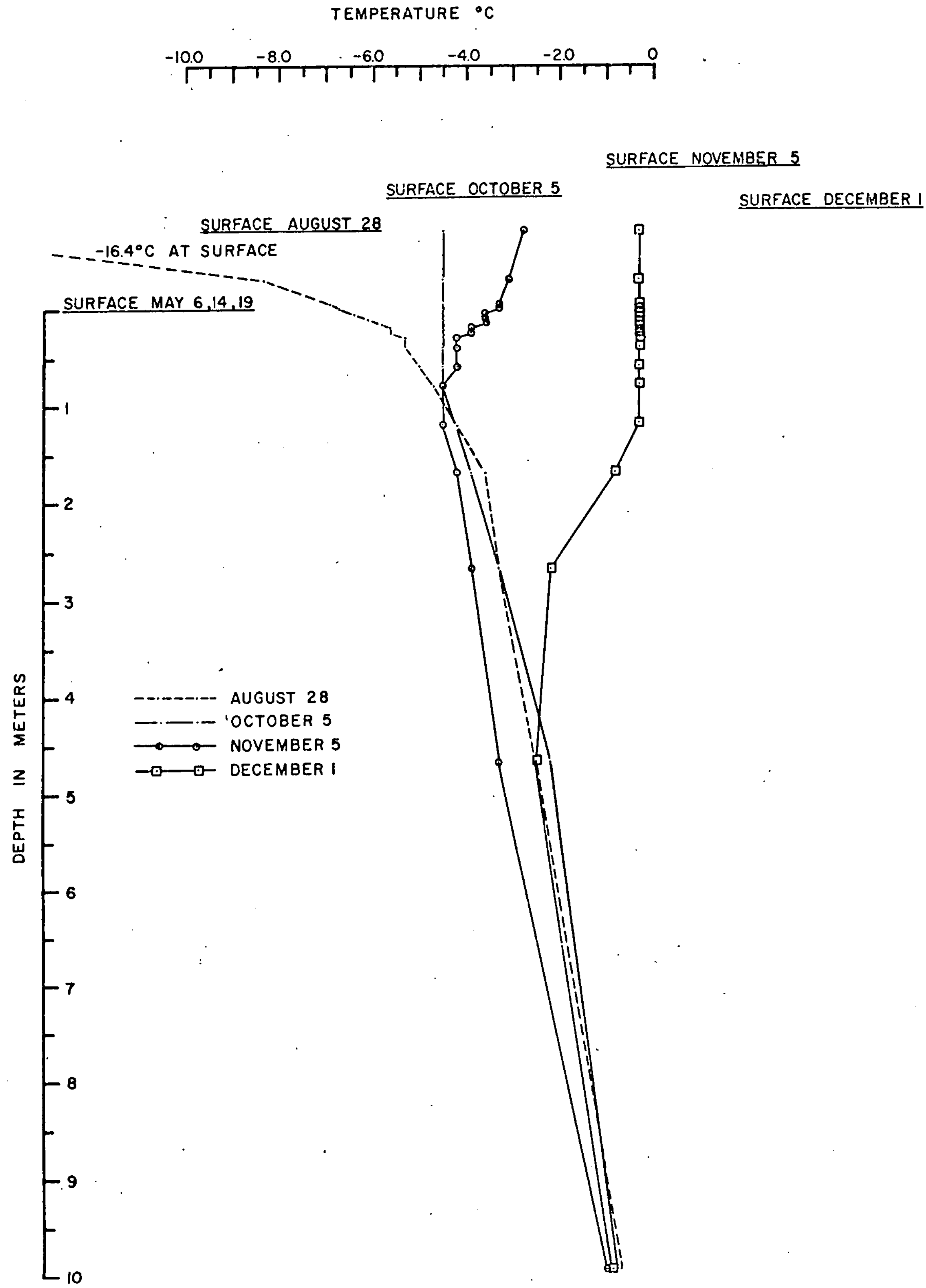


Fig. 55 (continued)

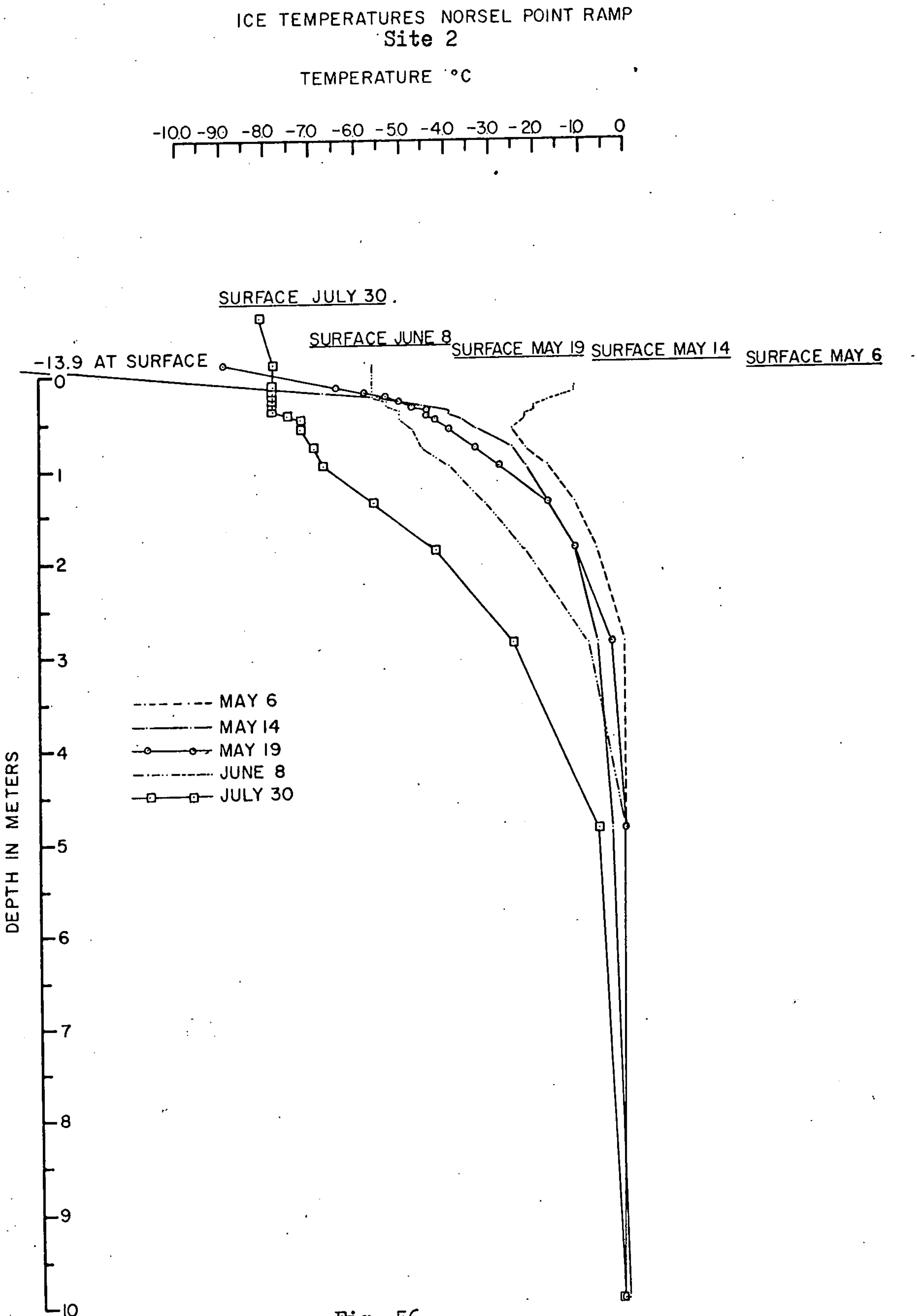

ICE TEMPERATURES NORSEL POINT RAMP
Site 2
TEMPERATURE °C
-10.0 -9.0 -8.0 -7.0 -6.0 -5.0 -4.0 -3.0 -2.0 -1.0 0
SURFACE JULY 30
SURFACE JUNE 8
SURFACE MAY 19
SURFACE MAY 14
SURFACE MAY 6
-13.9 AT SURFACE
0
1
2
3
4
5
6
7
8
9
10
DEPTH IN METERS
MAY 6
MAY 14
MAY 19
JUNE 8
JULY 30

Fig. 56.

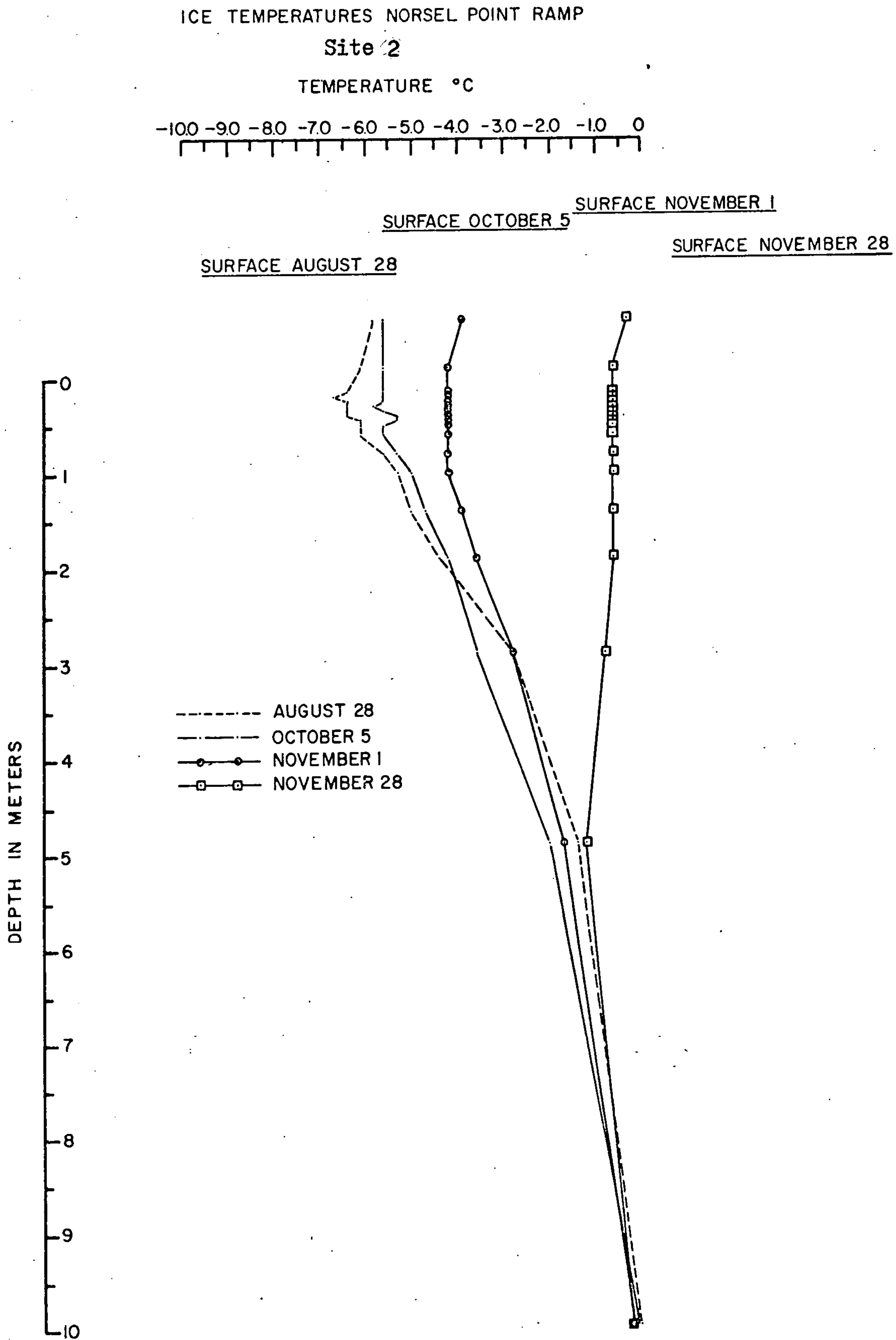
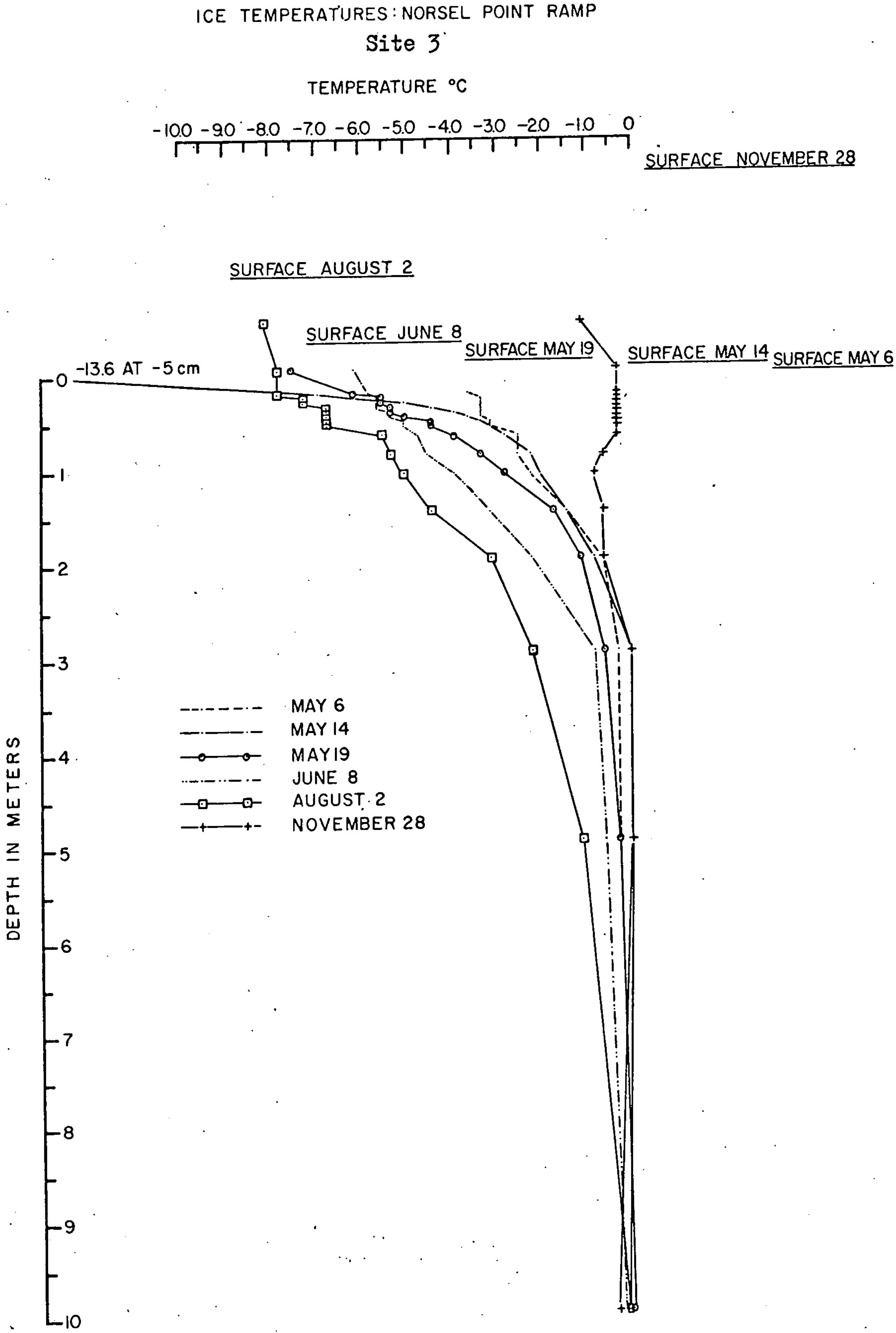


Fig. 56 (continued)



ICE TEMPERATURES: NORSEL POINT RAMP
Site 3
TEMPERATURE °C
-10.0 -9.0 -8.0 -7.0 -6.0 -5.0 -4.0 -3.0 -2.0 -1.0 0
SURFACE NOVEMBER 28
SURFACE AUGUST 2
SURFACE JUNE 8
SURFACE MAY 19
SURFACE MAY 14
SURFACE MAY 6
-13.6 AT -5 cm
0
1
2
3
4
5
6
7
8
9
10
DEPTH IN METERS
MAY 6
MAY 14
MAY 19
JUNE 8
AUGUST 2
NOVEMBER 28

Fig. 57.

site 2 at station Alpha and site 3 at station Big X. Thus site 1 was set in ice, while the other two were set in very icy firn.

The three set of curves are essentially similar and show the gradual penetration of the winter cold wave, with near-surface temperature fluctuations in response to air temperature variations and deposition of fresh, warmer snow. The winter cold wave reached its maximum effect by early October (the results at site 1 from November 5th are suspect and may result from malfunction of the recording bridge but more likely from observer error). After early October a gradual warming of the upper layers in response to general atmospheric warming, began to supress the winter cold wave.

It is noticeable that the penetration of the winter cold wave was more rapid at site 1 than at the other two sites and was greater at site 2 than at site 3. After early October however, the supression of the cold wave was more rapid at sites 2 and 3 than at site 1. This can best be explained in terms of the internal structure of the ramp.

Site 1 was situated very close to the position where Plummer found 52 cm. of superimposed ice resting on the old glacier ice surface. The other two sites were set in iced firn and where an alteration of ice layers and iced firn at depth might be expected. Thus the more rapid penetration of the cold wave at site 1 results from the relatively greater conductive ability of ice than the firn. The entrapped air in the firn has an insulating effect against the cold wave. The reverse condition results with the suppression of the cold wave. In the firn the downward percolation of surface meltwater is not impeded and the liberated heat of fusion is effective in raising the firn temperature. At site 1 however, the percolating water does not penetrate the ice to any great extent. It either forms superimposed ice (transient or permanent) or flows off the ice surface. Thus, by December 1, the warming of the ice at site 1 was generally completed only to a little over 1 meter depth, while at the other sites the warming was almost complete to much greater depth. The sharp increase in temperature at the base of the superimposed ice at site 1 is very pronounced in the final temperature curve. It is also well marked at about 2 meters depth at site 2. From Plummer's observations, this cannot be attributed to the same cause but probably does reflect the existance of a prominent ice layer at about 2 m. depth, which is impeding the downward flow of meltwater. At site 3, the percolation apparently is virtually uninterrupted and by November 28 almost all the profile was close to zero °C.

The 10 m. temperatures derived from the thermohms are of interest. Their fluctuations during the period of investigation are probably due to slight error in reading the Wheatstone Bridge but do not show well in the three sets of curves. They are given numerically in Table XXV. At site 1 the temperature varied between-0.7 °C and -1.0 °C with an average of -0.8 °C. This compares favorably with one observation at the base of the ice.

This was made on October 22 1965 in an ice cave behind Skua Lake. A horizontal borehole was drilled about 20 cm. above the ice bed for a distance of 8 meters into the basal ice. The temperature at 8 m.,

TABLE XXV

TEN METER TEMPERATURES: NORSEL POINT RAMP

Site 1 Date	°C	Site 2 Date	°C	Site 3 Date	°C
May 6	-0.8	May 6	-0.2	May 6	-0.1
May 14	-0.8	May 14	0.0	May 14	-0.2
May 19	-0.7	May 19	-0.1	May 19	-0.1
Jun 8	-0.8	Jun 8	-0.1	Jun 8	-0.25
Jul 30	-0.7	Aug 2	-0.1	Aug 2	-0.2
Aug 28	-0.7	Aug 28	-0.05	Aug 28	-0.1
Oct 5	-0.8	Oct 5	-0.1	Oct 5	-0.1
Nov 5	-1.0	Nov 1	-0.2	Nov 1	—
Dec 1	-0.9	Nov 28	-0.2	Nov 28	-0.4 ?
Avg	-0.8	Avg	-0.12	Avg	-0.15

Note: On November 1 site 3 was buried.

obtained with an insulated thermometer, was -0.8 °C

CONCLUSIONS

The ice temperature observations, because of the somewhat limited quality of measuring devices, were by no means as adequate as this author would have wished and would best be considered as "cursory". However, they do indicate that the main body of the ice piedmont is not truly sub-polar and is probably best defined, geophysically, as being between sub-polar and temperate.

The observations on Norsel Point ramp support the belief that the foot of the ramp at least, is frozen to the bed and is confirmed by the fact that subglacial melt streams have not been seen. At sites 2 and 3 on the ramp, the measured 10 m. temperatures were also below zero °C. but only slightly so and before any conclusions as to bottom condition there can be made, temperatures from greater depths are needed. The average 10 m. temperature at site 2 was -0.12 °C while at site 3 it was -0.15 °C.

Future workers on the Marr Ice Piedmont should devote considerable effort toward ice temperature studies so that its geophysical description can be more precisely determined and its dynamic behavior more clearly understood.

CHAPTER ELEVEN

ASPECTS OF THE GLACIER MECHANICS

In Chapter 5 above, the distribution of surface velocity was discussed in the simple terms of its relationship to the distribution of accumulation, subglacial topography and ice streaming effects. However, the total observed surface velocity of a glacier has been shown by Nye (1952; 1953) to be the sum of the velocity due to creep or deformation within the glacier and the velocity due to the glacier sliding over its bed. These predictions are based on the creep law for ice, valid for ice at a temperature close to the melting point, by Glen (1955). On the Marr Ice Piedmont the relative distribution of the velocity due to internal deformation and that due to basal sliding presents a more complex situation than was discussed earlier in Chapter 5 and has possible implications in terms of erosion by the piedmont.

BASAL SLIDING

Nye (1952) noted that mathematical analysis might well be able to explain the observed relative velocities in a glacier but that no satisfactory answer seemed to have been given to the question of what ultimately determines the basal sliding velocity. This question was a serious one in that it was known that sometimes the major contribution to the observed surface velocity came from the basal sliding and that only a small part resulted from differential motion within the ice. For example, in the Jungfrau borehole experiment (Gerrard and others 1952), it was noted that over 50% of the total measured surface velocity was accounted for by basal sliding. On Highway Glacier on Baffin Island, Ward (1955) found that over 80% was accounted for by basal sliding and on Veslskautbreen, McCall (1952) reported a value of over 90%.

Theories of glacier sliding have been proposed and developed principally by Weertman (1957; 1964) and Lliboutry (1965; 1968a; 1968b) for a glacier whose bottom surface is at the pressure melting point. In essence, the two theories are quite similar, though in detail there are several divergences but the one according to Weertman appears to be the more widely accepted (Paterson 1969).

Two mechanisms are invoked by Weertman (1957; 1964). The first is the pressure melting phenomenon. With this mechanism pressure is built up on the upstream side of a bedrock obstacle. Ice is melted on this high pressure side of the obstacle and the meltwater flows to the low pressure side where it refreezes. Thus, the glacier can slide. This mechanism has been observed and verified directly by Kamb and LaChappelle (1964) on Blue Glacier. However, with this mechanism the rate of sliding will be controlled by the rate of the heat flow through the larger obstacles and would be negligible. The second mechanism is the enhancement of the creep rate of the ice in the vicinity of the glacier bed through stress concentrations. In this case, the creep flow of the ice around the smallest obstacles will determine the rate of sliding and again the movement will be negligible. However, the combination of the two mechanisms operating simultaneously, will lead to appreciable sliding velocities.

However, theory is generally inapplicable to data from the field because the data are deficient in the respect of the "roughness Factor" of the glacier bed. This factor is dominant in the sliding equations. Unless the roughness of the bed is measured, and to date it rarely has been, the sliding velocity must be "assumed" or inferred as the difference between the observed surface velocity and the calculated velocity due to internal deformation of the glacier ice. As no information is available on bed roughness beneath the Marr Ice Piedmont, the sliding velocities reported below are assumed from the deformation calculations.

DEFORMATION AND SLIDING VELOCITY

Basal Shear Stress

The shear stress (τ_{zx}) acting on a plane parallel to the surface for unit width of ice, is approximately the tangential component of the pressure due to the weight of the overlying column of ice. Thus the simplest assumption that can be made about the state of deformation which produces a known velocity gradient is that the deformation occurs as simple shear in a direction parallel to the surface and that the only shear stress components are those which produce the simple shear flow (Meier 1960). Here it is necessary to assume that the simple shear planes are parallel at all depths and that the increase of the shear stress with depth, is linear (Nye 1952). The flow of ice therefore, is assumed to be laminar.

The shear stress on a plane at depth d is, to a good approximation (Nye 1952) given by:

$$\tau_{zx} = \rho g d \ \mathrm{Sin}\ \alpha$$

where ρ is the density of ice, g is the acceleration due to gravity and α is the slope of the surface. Nye (1952 p.86) has shown that the surface slope rather than that of the bed is the effective slope. The shear stress acting at the bed (τ_b) is therefore given by:

$$\tau_b = \rho \ gh \ \sin\alpha$$

where h is the measured ice thickness.

For practical purposes, ρ and g are taken as constants: $\rho = 0.9$ g/cm^3 and the acceleration due to gravity (g), 982.312 cm/sec^2, is the value recorded on a concrete pier at Palmer Station, by Dewart in 1967.

Measurement of Surface Slope

Values of surface slope have more generally been taken from the differences in elevation between successive stations, (for example Orvig(1963), Bull (1957))which is a quite satisfactory method if the slope is reasonably constant between stations. On the Marr Ice Piedmont surface topography is often too pronounced and detailed to determine in this way. It was therefore, measured directly.

At each velocity station the theodolite was set up and precisely leveled using a 3.6" Wild striding level. A circle, 200 m. in diameter, was described about the station with the aid of a 100 m. string. At 20* intervals on the circle's circumferance, the instrument's height was read from a level staff. The "preceeding station" in the ice movement survey sequence was taken as 0° on the circle. In this way it is possible to ultimately resolve the true azimuth of the maximum and minimum slope. The author is grateful to L.E. Brown who made these measurements, often alone, during 1966. An example of the results is shown in Figure 58. In few cases was the slope symmetrical as can be seen from the differences in the oposing maximum and minimum readings. Slight curvature of the surface is probably real in many cases but the difference in readings can probably be attributed also to slight local differences on the surface, for example, small sastrugi, which over such short sighting distances would have appreciable effect. Errors in reading the staff are probably negligible. For the calculation of τ_b the angle of maximum slope was determined as the mean of the maximum and minimum readings in the direction of maximum slope.

Values of the Basal Shear Stress

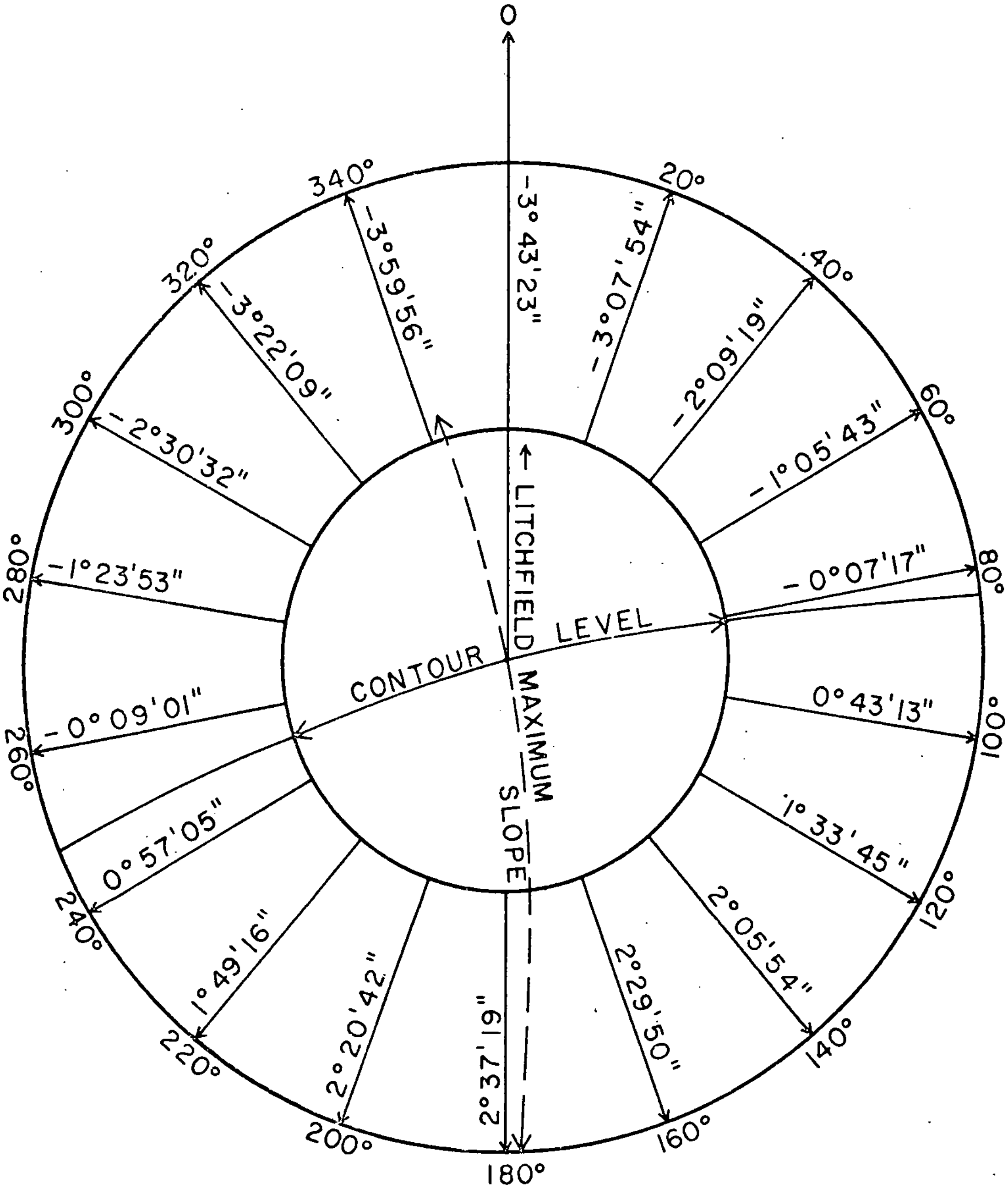
Values of τ_b range from 0.47 bar to 1.56 bar and are shown in Table XXVI (1 bar= 10^6 dynes/cm^2). This range of values is not unreasonable; Nye (1952) found that for 16 Alpine glaciers the range was 0.5 bar to 1.5 bar. C.B. Bull (personal communication) states values up to 1.6 bar on Sherman Glacier. Values from the Barnes Ice Cap on Baffin Island (Orvig 1953) and Greenland (Bull 1957) are much lower however, but in these cases the values can be explained by the much lower velocities recorded.

The Assumption of the Sliding Velocity

The flow law of ice has the general form $\dot{\gamma} = K\tau^n$, where $\dot{\gamma}$ is the rate of shear strain produced by a shear stress τ. K and n are taken as constants even though laboratory experimental data indicate that this is not actually the case. Nye (1965) has elaborated on the dev-

STATION E.

SLOPE CONFIGURATION



ASPECT 345°

SHAPE CONVEX AND SLIGHTLY STEEPER IN THE THIRD AND FOURTH QUADRANTS

Fig. 58.

TABLE XXVI

ICE THICKNESS, SURFACE SLOPE AND CALCULATED BASAL SHEAR STRESS

		Slope (α)			
Station	h (m)	°	'	"	(bar)
E	255	3	18	38	1.30
H	312	2	03	58	0.99
K	348	2	08	35	1.15
L	364	1	32	51	0.86
M	348	1	55	04	1.03
N	392	1	30	15	0.91
P	448	1	29	47	1.03
R	496	1	17	45	0.99
S	522	1	11	00	0.95
T	550	0	58	41	0.82
U	581	0	58	00	0.87
V	589	0	56	00	--
W	557	0	48	00	--
x	510	0	40	00	--
C1	361	2	30	47	1.40
C2	306	3	18	30	1.56
C3	372	2	44	09	1.56
C4	416	2	00	44	1.28
C5	447	1	57	26	1.35
C6	461	1	44	00	1.23
T1	506	1	18	00	1.01
T2	535	1	08	00	0.93
T3	558	0	55	08	0.79
T4	598	0	45	00	0.69
T5	612	0	40	00	0.63
N1	452	1	22	55	0.96
N2	425	1	33	42	1.02
N3	382	2	01	21	1.19
N4	366	2	19	10	1.31
N5	401	1	58	05	1.22
N6	322	2	33	12	1.27
N7	306	2	50	00	1.34

Continues

TABLE XXVI (Cont)

Station	h (m)	Slope (α) °	'	"	(bar)
Pi1	471	1	05	00	0.79
Pi2	453	1	10	00	0.81
Pi3	536	1	08	00	0.84
Pi4	423	1	16	51	0.84
K1	317	1	48	44	0.89
K2	332	2	35	49	1.33
K3	356	2	24	10	1.32
G	274	3	35	20	1.52
G1	287	2	50	51	1.26
G2	380	1	43	47	1.01
G3	339	2	00	00	1.05
H1	260	3	01	13	1.21
H2	245	3	55	04	1.48
H3	205	1	30	00	0.47
Mu1	404	1	54	00	1.18

elopment of Glen's (1955) creep experiments by other researchers and on the variation of the value of the exponent n. It seems well established however, that n increases with increasing stress with a range of about 1.6 to 2.2 for shear stresses between 0.25 and about 1.0 bar. At much higher stresses reaching up to 15.0 bars, the value of n reaches to about 4.0. The value of n therefore, can be taken as equal to 2, 3 or 4 depending on the stress range but if n is taken as 3, the power law of flow gives good agreement with observation. Thus for the purpose of calculation, $n = 3$ is taken as applicable over the whole stress range and is a good "average" value.

An average value of K must also be taken in the practical case of a moving glacier because of differences in internal structure, the degree of bubbliness and the amount of impurities contained within the ice and variations in temperature.

In deducing a flow law from the deformation of a bore hole in the Jungfraufirn, Gerrard and others (1952) assumed laminar flow and obtained good agreement with their observations and the power law $\dot{\gamma} = K \tau^n$, where K and n are constants. The stress was 0.75 bar and n proved to be 1.5 but was expected to rise to about 4 at stresses higher than 0.75 bar. Except for the upper 15 meters of the firn, the Jungfraufirn was at or close to 0 °C and the results of the experiment were regarded as applicable to ice at 0 °C. When the shear stress is measured in bars and the shear strain rate is expressed per second, the value of K proved to be, $K \approx 10^{-8}$.

No firm conclusions can be drawn about the general temperature distribution within the Marr Ice Piedmont. The temperature at 10 to 12 m.

depth ranged from −4.9 °C in the highest parts of the piedmont to slightly below 0 °C near the coast. The temperature at the base of the piedmont, in the absence of any flow, can be estimated from Robin's (1955) graphical summaries (reproduced in Figure 59) to be of the order of −2 °C. However, additional heat will be supplied as a result of the work done by movement. In calories, this would amount to about 20 cals/cm^2/yr. or approximately the geothermal heat flow, for a movement rate of 18 m./yr. In all cases on the Marr Ice Piedmont the movement rates are high enough to produce heat at least equal to the geothermal heat and in most cases considerably in excess of this. It follows that the heat produced by such rates of movement will raise the temperature of the ice considerably above the theoretical −2 °C, and there is little reason to believe that the average temperature of the piedmont is far from 0 °C. If the basal temperature of −0.8 °C, which was recorded in the ice of the cave behind Palmer Station, is valid, then an average temperature for the ice piedmont must be of the order of −1.0 °C to 0 °C.

Thus there appears to be a close physical and thermal similarity between the ice piedmont and the Jungfraufirn. Therefore, internal deformation velocity within the pidmont have been calculated from the Jungfraufirn flow law (Gerrard and others 1952). Taking $\dot{\gamma} = K\,\tau^n$, with $\dot{\gamma}$ expressed as the strain rate per second, $n = 3$ and $K \approx 10^{-8}$, integration over the thickness of the ice (Nye 1952) leads to the following expression for the velocity due to internal deformation (U_c).

a b

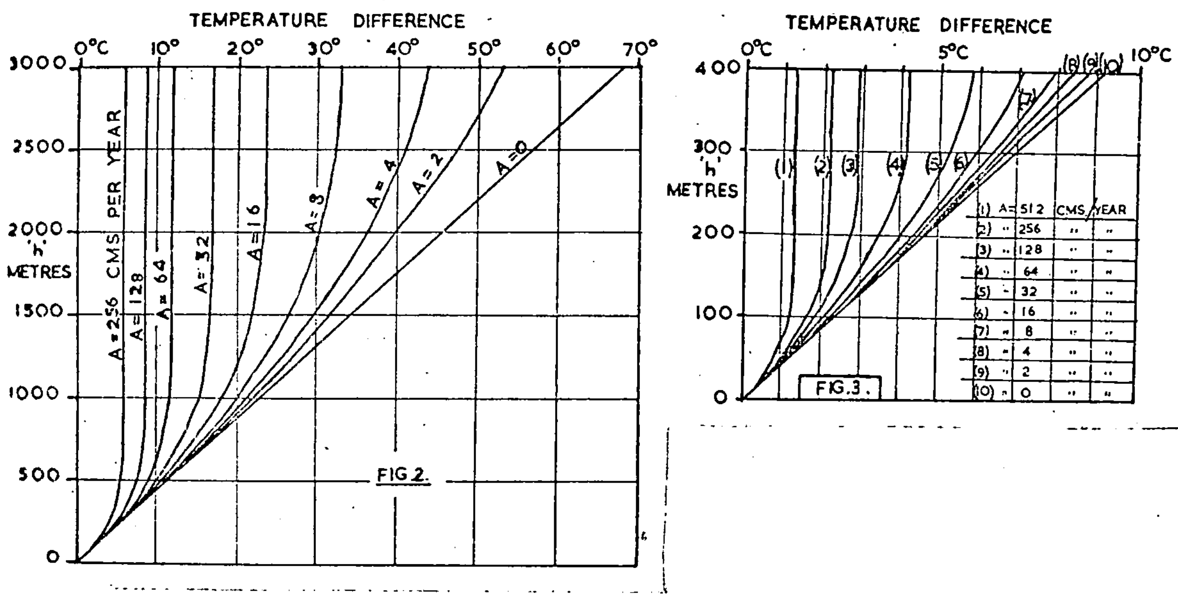


Fig. 59. (From Robin, 1955, p. 527). Temperature difference between the base and other levels near the center of an ice sheet: a Central thickness 3,000 meters. b Central thickness 400 meters.

$$U_o - U_b = U_c = K(\rho g)^n \frac{h^{n+1}}{n+1} \sin^n \alpha$$

$$= K \frac{\tau^n}{n+1} \cdot h$$

where U_o is the surface velocity, U_b is the basal sliding velocity and U_c is internal deformation velocity. Values of these parameters are given in Table XXVII.

Validity of the Assumed Sliding Velocity

Basal sliding is believed to account for up to 90% of the observed surface velocity in extreme cases for temperate valley glaciers. Generally, 50% of the observed total appears to be accounted for by sliding. On the Marr Ice Piedmont basal sliding accounts for 10% to 30% of the surface velocity in areas not effected by ice streaming, while in the ice streams themselves 50% to 70% of the surface velocity is accounted for by sliding. The higher range is remarkably similar to that of temperate and near-temperate valley glaciers. The sliding velocities assumed for the ice piedmont do therefore, appear to be acceptable.

BASAL CONDITION OF THE PIEDMONT

In Table XXVI are given the calculated values for the shear stress acting at the base of the ice and Table XXVII shows the assumed sliding velocity and its percentage contribution to the total, observed annual velocity rate at the surface. Some of the results in Table XXVII are

TABLE XXVII

VALUES OF THE "ASSUMED" SLIDING VELOCITY (U_b)
(Meters per Year)

Station	$U_o - U_b (U_c)$	U_o	U_b	U_b as % U_o
E	44.36	20.46	-	-
H	24.18	25.96	1.78	6.86
K	41.66	28.26	-	-
L	18.63	26.94	8.31	30.84
M	29.91	27.01	!	-
N	23.26	39.45	16.19	41.03
P	39.03	49.13	10.10	20.56
R	38.12	49.89	11.77	23.59
S	35.61	43.37	7.76	17.89
T	24.35	35.06	10.71	30.55
U	29.80	28.90	!	-
C1	77.99	97.33	19.34	19.87
C2	91.74	108.98	17.24	15.82
C3	110.50	108.51	!	-
C4	67.97	95.27	27.30	28.65
C5	86.67	81.15	!	-
C6	68.07	65.55	!	-
T1	41.70	42.65	0.95	2.20
T2	34.53	32.49	!	-
T3	21.79	23.55	1.76	7.47
T4	15.66	17.25	1.59	9.22
T5	12.05	13.89	1.84	13.24
Pi1	18.12	64.13	46.01	71.74
Pi2	19.36	79.13	59.77	75.53
Pi3	25.11	83.50	58.39	69.93
Pi4	19.49	83.79	64.30	76.73
N1	31.90	70.70	38.80	54.88
N2	35.99	88.92	52.93	59.52
N3	50.97	118.22	67.25	56.88
N4	64.77	156.64	91.87	58.65
N5	57.17	207.35	150.25	72.45
N6	51.77	210.47	158.70	75.40
N7	57.67	218.74	161.07	73.63

TABLE XXVII (cont)

Station	$U_o - U_b (U_c)$	U_o	U_b	U_b as % U_o
K1	17.40	14.25	!	-
K2	61.59	31.42	-	-
K3	64.48	74.68	10.20	13.65
G	75.30	31.34	-	-
G1	45.31	29.37	-	-
G2	31.25	109.95	78.70	71.57
G3	30.58	118.34	87.76	74.16
Mu1	52.90	116.58	63.68	54.62
H1	36.40	16.77	-	-
H2	62.60	20.87	-	-
H3	1.72	42.63	40.91	95.96

inconsistant and obviously something is in error because it is impossible for the internal deformation velocity to be greater than the actual velocity measured at the surface. The surface velocity survey is regarded as sufficiently accurate to be eliminated as the probable cause of the inconsistencies and the remaining possible errors are three-fold: a) Ice thickness; b) Surface slope; c) The constants in the equations used.

The assumption of the sliding velocity is a function of the calculated internal deformation velocity which in turn is a function of the calculated basal shear stress. Therefore, the basal shear stress could be in error because of errors in ice thickness and surface slope. The ice thickness values include an estimated possible error of ± 10% which is quite considerable and would effect the basal shear stress in the order of 12%. The surface slopes which have been used could be an even more significant source of error, being greatest in the topographically irregular peripheral area of the piedmont. The slope used in the shear stress calculations ideally should be that over a distance at least equal to the ice thickness (Nye 1952) and this condition has probably not been met at all the velocity stations. At high elevations, where the slope is more constant, the values used are probably more valid than at lower elevations where the measured slope is local. The error of ice thickness can be applied to reworked calculations but it is difficult to assign an error to the slope measurements.

The calculation of the internal deformation velocity used the calculated basal shear stress and the constants n and K. This equation is a good approximation at best and Nye (1965) has discussed at length the possible values of n and it is known that it varies with the stress. The constant K is very temperature sensitive and the values used in this thesis were experimentally derived from ice at close to zero degrees C (Gerrard and others 1952). It is assumed that the Marr Ice Piedmont is not significantly far from being temperate and that its classification lies between true temperate and true sub-polar. Therefore, its overall temperature must be close to zero °C also. Thus the errors in ice thickness and surface slope and an uncertainty as to the exact temperature condition of the ice, hardly justify manipulation of the values of the constants n and K in the equations. The ice thickness values are the only ones which justifiably can be manipulated, within the limits of the estimated error, in the calculation of internal deformation velocity. Table XXVIII itemises the results obtained for stations E to U for the assumed sliding velocity using the (average) ice thickness from Chapter 3 and these values plus 10% and minus 10%. The justification for these results is based on the assumption that some basal sliding is occuring at all the stations, which is reasonable considering that the ice is regarded as being close to temperate. However, in order to obtain a sensible distribution of internal deformation velocity and basal sliding, as shown in Figure 60, the values of U_b have to be selected subjectively, which is not reasonable. The values in Table XXVII are therefore used for discussion.

In this table the symbol (!) in column 4 is used to indicate that

TABLE XXVIII

VALUES OF SLIDING VELOCITY (U_b) FOR MAXIMUM, MINIMUM AND AVERAGE ICE THICKNESS (h) AND RESPECTIVE PERCENTAGE OF THE SURFACE VELOCITY (U_o)

Stations E-U

Station	U_b m/yr (Avg. h)	% U_o	U_b m/yr (h + 10%)	% U_o	U_b m/yr (h - 10%)	% U_o
E	-23.90	-	-44.02	-	- 8.90	-
H	+ 1.78	6.86	-19.36	-	+10.05	38.71
K	-13.40	-	-32.84	-	+ 0.98	3.47
L	+ 8.31	30.84	- 0.23	-	+14.66	54.42
M	- 2.90	-	-16.87	-	+ 7.45	27.58
N	+16.19	41.03	+ 5.46	13.84	+24.15	61.22
P	+10.01	20.56	- 8.10	-	+25.83	52.57
R	+11.77	23.59	- 5.68	-	+25.19	50.49
S	+ 7.76	17.89	- 8.70	-	+19.96	46.02
T	+10.71	30.55	- 0.60	-	+20.15	57.47
U	- 0.90	-	-13.80	-	+ 9.34	32.32

Note: Negative values of U_b result when calculated $U_o - U_b$ exceeds U_o in value.

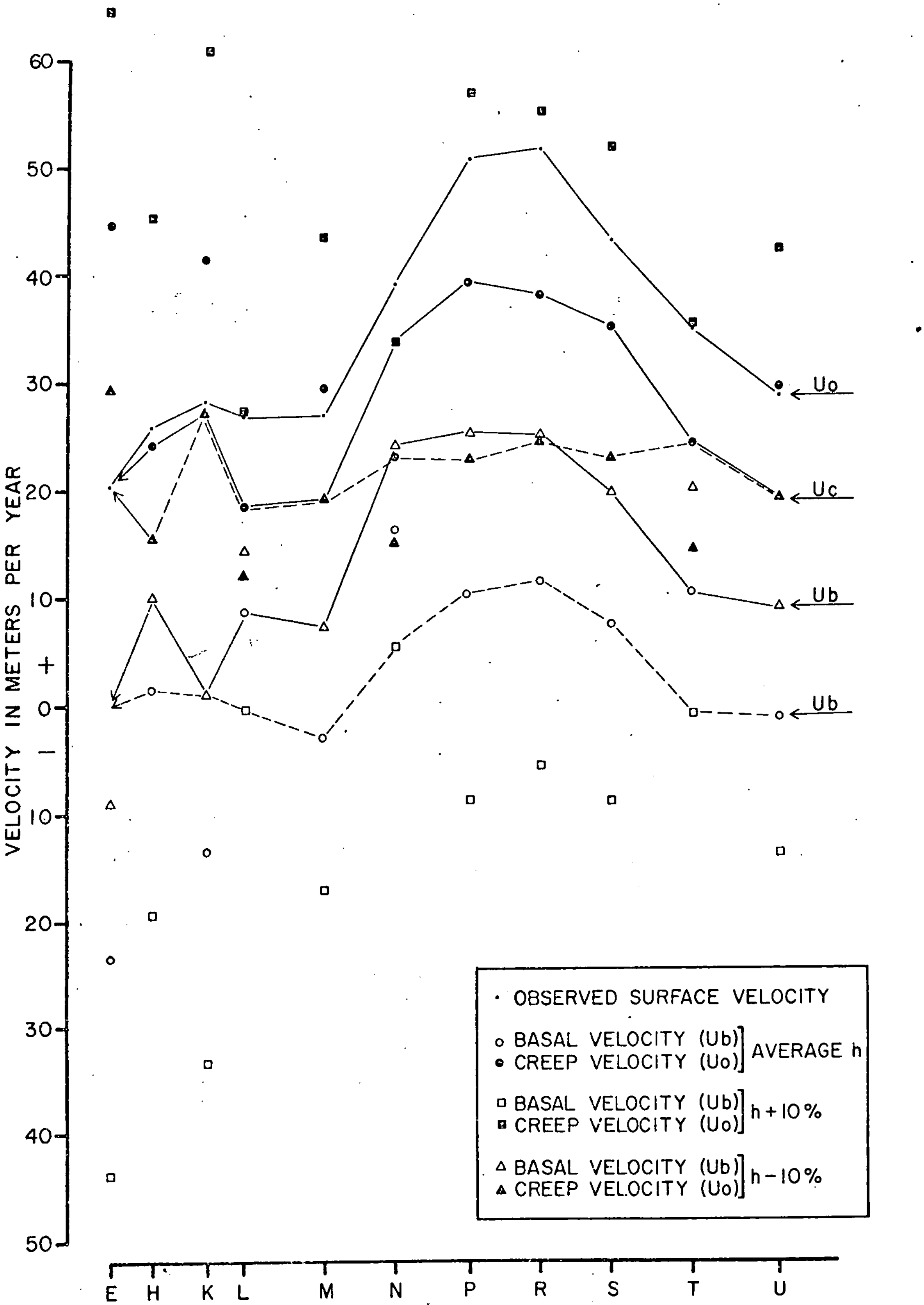


Fig. 60. Possible curves of basal sliding and internal deformation velocity calculated with various values of ice thickness (h).

though no basal sliding has been assumed, the situation is marginal and that a reasonable variation in the equation components could render a positive sliding velocity value. The corollary is that the calculated internal deformation velocity is almost equal to the observed surface velocity. In the same column the symbol (-) is used to emphasise that the internal deformation velocity calculates to be greatly in excess of the observed surface velocity and that no reasonable alteration of the equation components could render a sensible sliding velocity value. The symbol (-) in the final column signifies that from the calculations, basal sliding does not contribute to the total observed ice movement.

Despite the inconsistencies in these results, a sensible pattern can be discerned. In the ice stream flowing towards Biscoe Bay (the Pi line and Neumayer lines principally), the proportion of the total velocity attributed to basal sliding is clearly defined, even within the limits of possible error, and accounts for well over 50%. Stations Mu1, G2 and G3 have sliding velocities of similar percentage and are very similar to those reported for temperate glaciers elsewhere.

Consider the distribution of the marginal condition symbolised by (!). This condition generally embraces the higher parts of the piedmont outside of the ice streams and it is evident that it is generally associated with low assumed sliding velocities nearby. The unreasonable situation (-) is generally peripheral. The distribution of the three situations, areally over the study area is shown in Figure 61: I, sliding; II, marginal; III, unreasonable. This is an interesting and suggestive pattern.

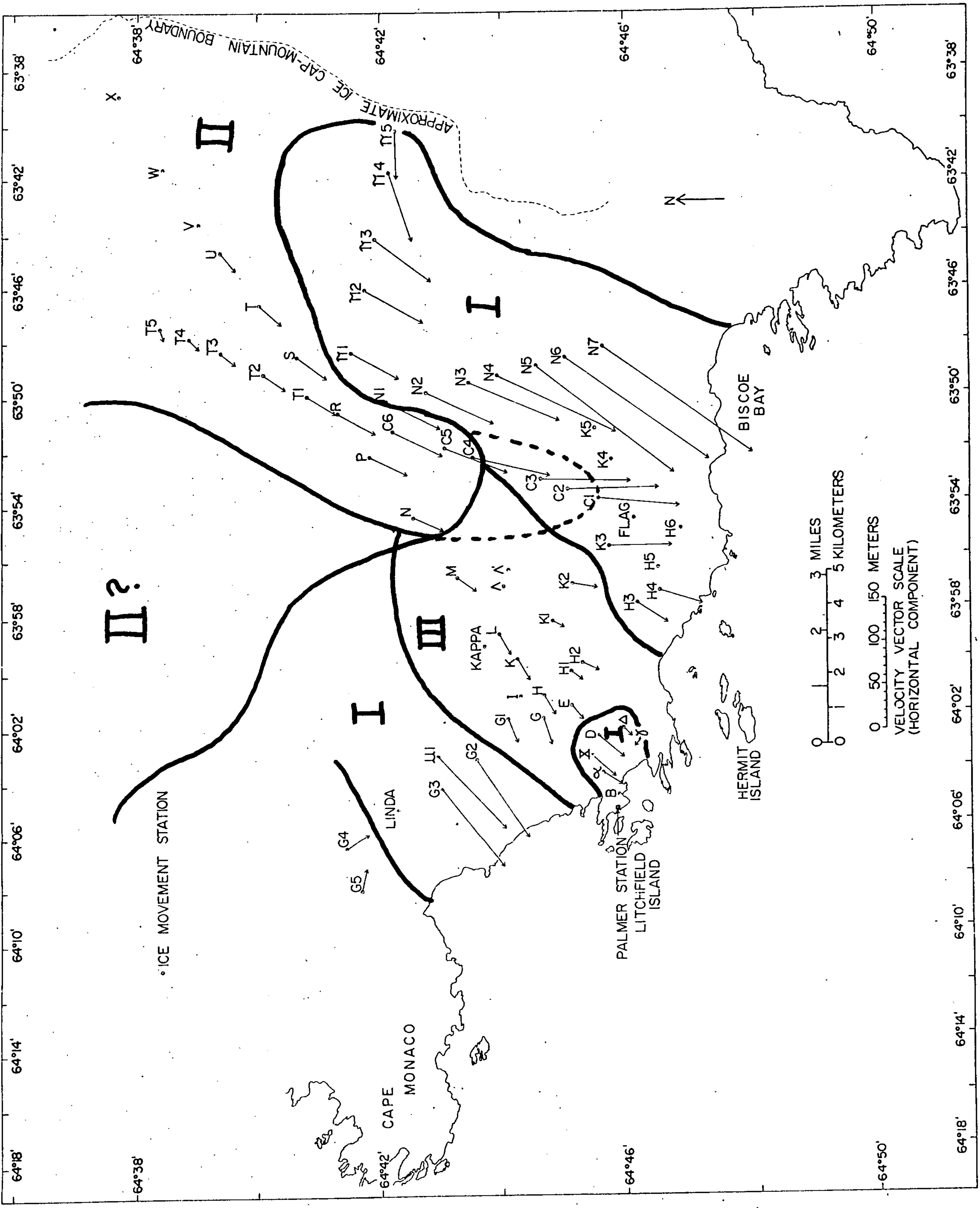


Fig. 61. Proposed zonation of basal conditions in the study area.

The sliding streams are associated with the subglacial valleys. The marginal condition is associated with the high subglacial platform and the flanks of the subglacial valleys and the unreasonable situation is associated with the low, peripheral subglacial saddle separating the Wylie Bay and Biscoe Bay subglacial valleys. The position of all the stations, relative to the subglacial topography, is shown in Figures 62 and 63 (also Figures 14 & 16 above). The position of station H3 is note-worthy.

The distribution of the basal shear stress and the observed surface velocities are also of significant interest. From Table XXVI and XXVII it is evident that there are considerable differences in the basal shear stress from one part of the study area to another and that these differences cannot be explained solely by differences in the rates of movement. Figure 64 is a plot of surface velocity versus basal shear stress and it is evident that widely differing velocities are associated with similar shear stresses. Orvig's (1953) argument for the Barnes Ice Cap, that the higher shear stresses can be explained by greater ice movement, does not hold. For example, from Figure 64, the surface velocities associated with a basal shear stress of about 1.2 bar range from 16.77 m. per year at H1 (1.21 bar) to 118.22 m./year at N3 (1.19 bar) to 218.74 m./year at N7 (1.34 bar). Similarly, the very high shear stresses in the vicinity of E, K, G, H1 and H2 are not explainable in terms of greater ice movement. Differences in the basal condition must therefore, be the explanation.

It is suggested that the three zones described in Figure 61 reflect these differences in basal condition. Zone I is formed by the Biscoe

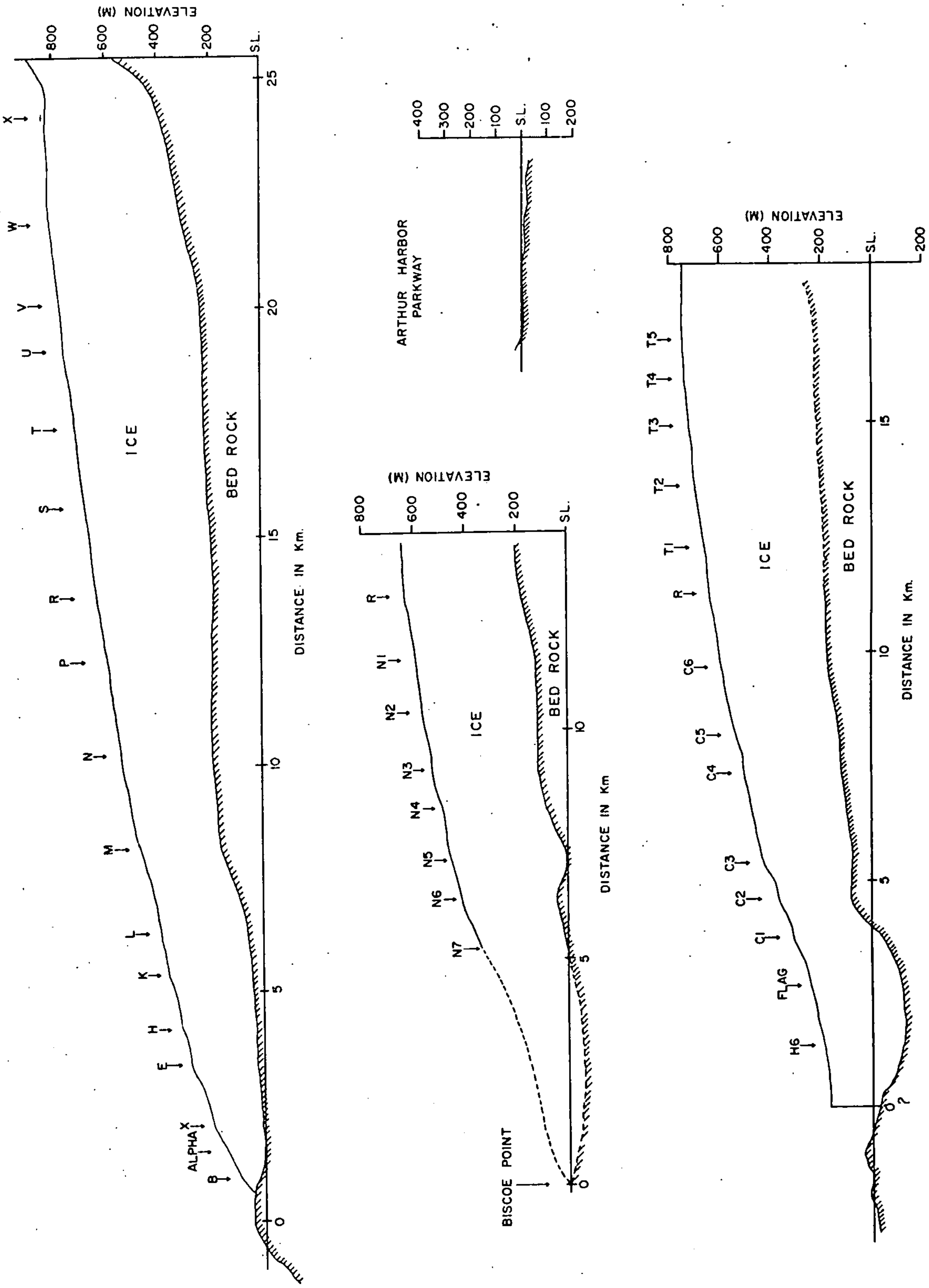


Fig. 62. Longitudinal profiles of bedrock and ice surface.

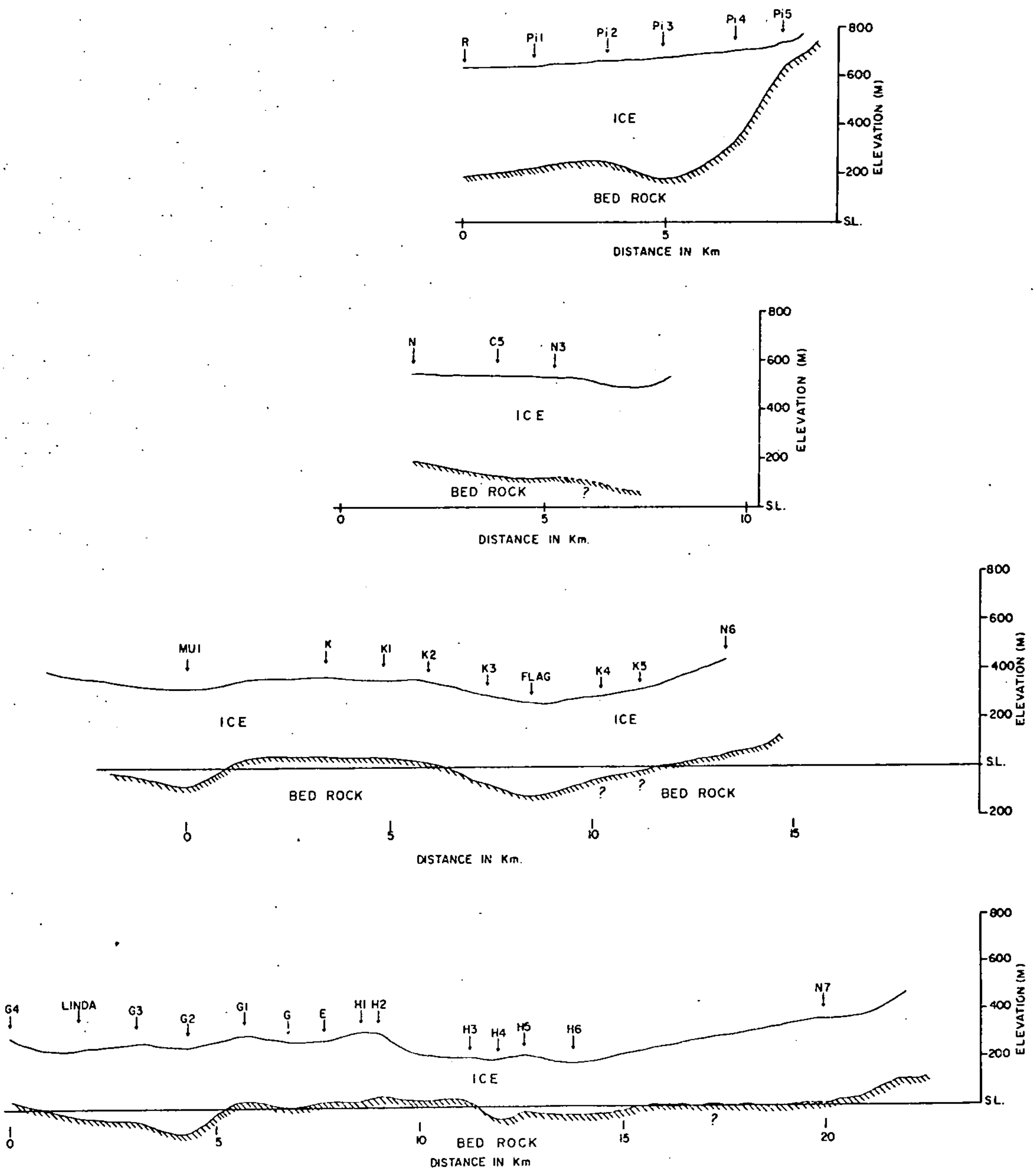


Fig. 63. Transverse profiles of bedrock and ice surface.

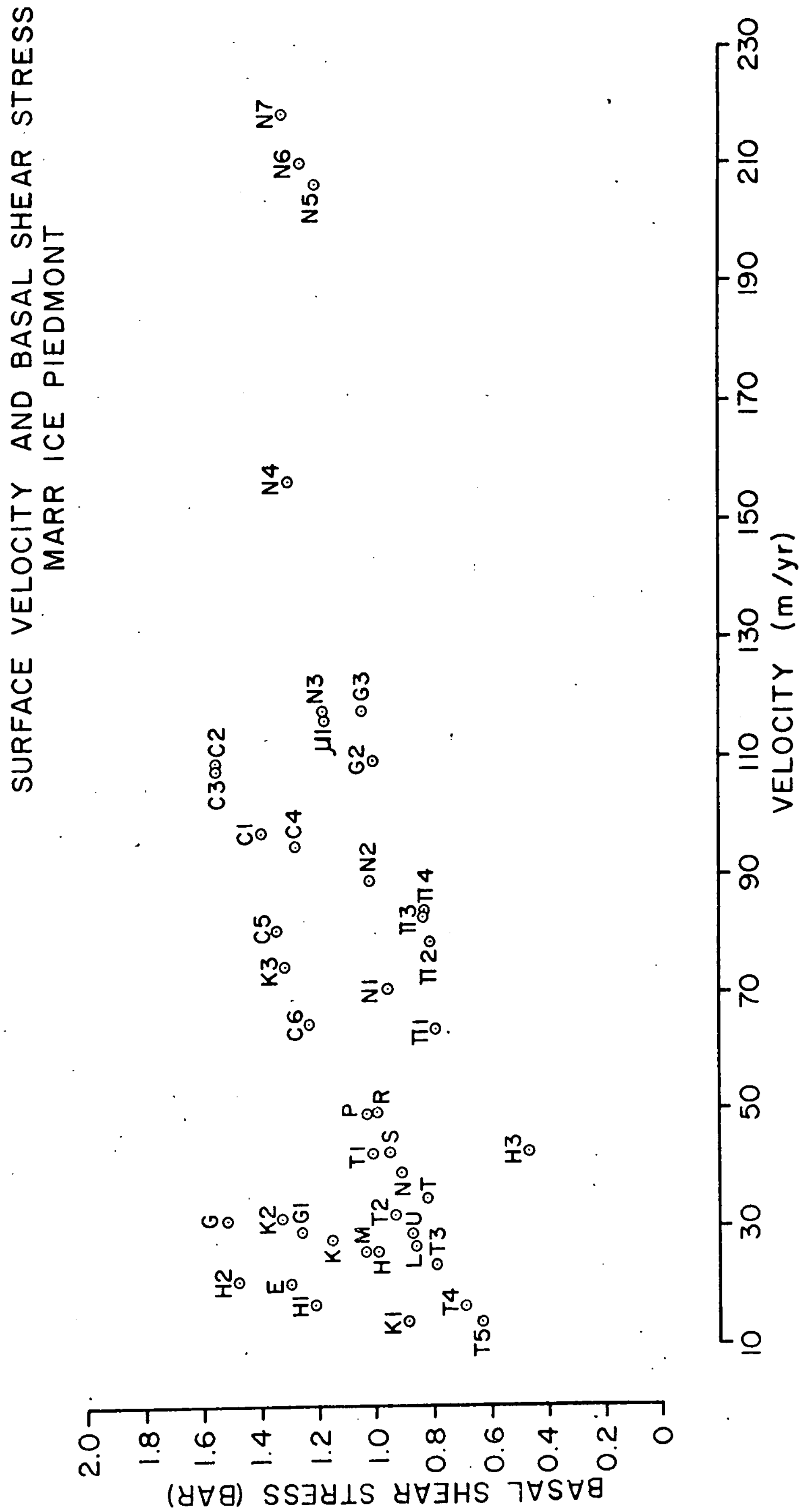

SURFACE VELOCITY AND BASAL SHEAR STRESS
MARR ICE PIEDMONT
BASAL SHEAR STRESS (BAR)
VELOCITY (m/yr)

Fig. 64.

Bay ice stream and its very high rate of movement, under relatively low basal shear stress, is the result not only of a high rate of basal sliding but that the sliding itself may be enhanced by basal melting. The same explanation is suggested for the lower parts of the Wylie Bay ice stream marked by stations Mu1, G2 and G3. Thus the "lubricating" effect is greatest below the ice streams than other parts of the piedmont and the streams are behaving very much like temperate valley glaciers. In zone II, the marginal condition, it can be reasonably inferred that most of the observed surface velocity is made up of internal deformation velocity but that a small amount of basal sliding is taking place. In this zone, a basal sliding mechanism similar to the "slip-and-stick" mechanism proposed by Weertman (1957; 1964) is envisaged. In zone III it is barely possible to reconcile the low surface velocity with those which would be explained by the basal shear stresses and it is suggested that in this zone, the piedmont is frozen to the bed. Station H3 (and H4, H5 and H6, if velocity values had been derived) whose total movement apparently is made up almost exclusively of basal sliding, lies over the sub-sea level glacier bed and thcugh the ice there may not be actually floating, the bed is significantly lubricated by the sea.

The considerations put forward by Weertman in his (1961) "Freezing Model" hypothesis are only partially helpful in envisioning this zonation of the piedmont, because the temperature data particularly are somewhat inadequate. In that hypothesis, an argument against the shear hypothesis for the formation of the moraines of Barnes Ice Cap, Baffin

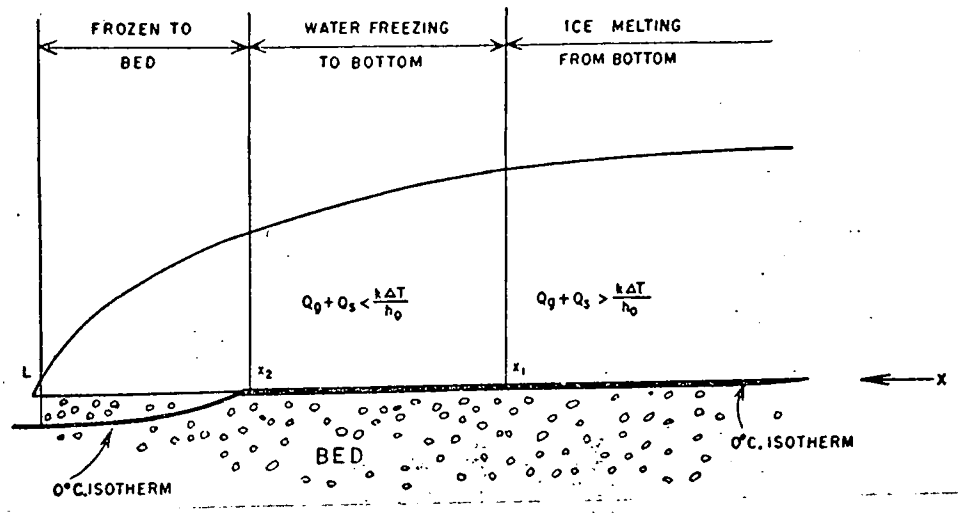


Fig. 65. (From Weertman 1961, p. 971). Schematic diagram of proposed geophysical zonation of an ice cap according to the "Freezing Model" hypothesis.

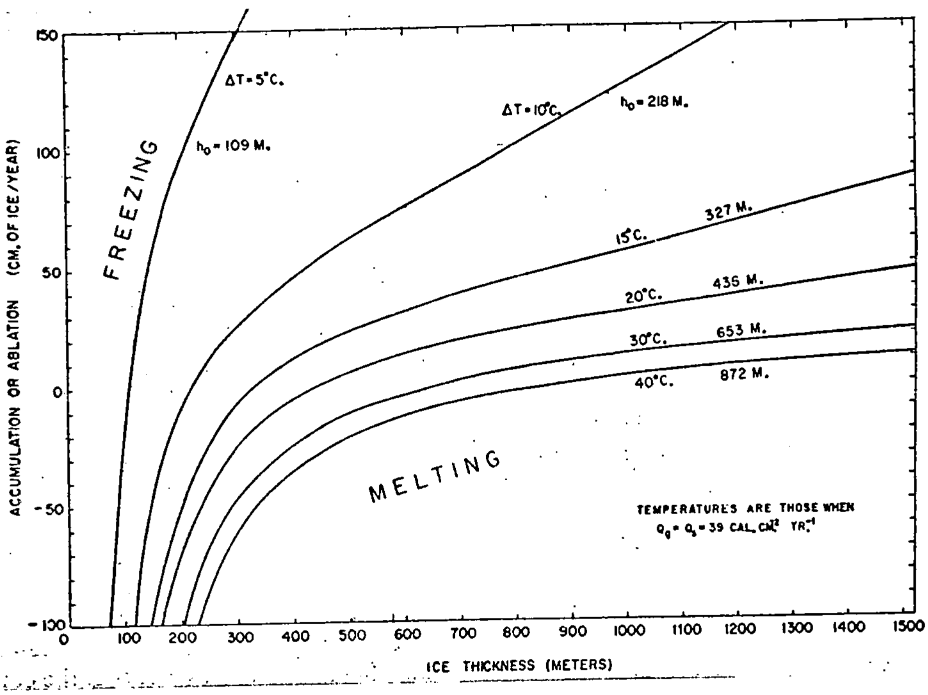


Fig. 66. (From Weertman 1961, p. 975). Graphical summary of proposed basal condition with differing values of ice thickness, annual accumulation rates and internal ice temperature. ΔT is difference between melting point of ice and upper surface temperature of ice sheet. h_0 is ice thickness at which base of ice is at pressure melting point.

Island, Weertman (1961) suggested that a cold ice cap whose edge is frozen to bedrock should have a geophysical structure as shown diagramatically in Figure 65. The scheme assumes that inland from the edge the bottom is at the pressure melting point and that this inland section can be divided into two geophysically differing parts. In the furthermost inland part the combined geothermal heat and the heat of sliding is greater than can be conducted down the temperature gradient and the basal ice is melted to water. The meltwater is forced outward toward the edge of the ice cap. As it moves outward it enters a section where the temperature gradient can conduct away more than the geothermal heat and heat of sliding. In this section the water freezes to the bottom and rejoins the ice cap. The basal ice will still be at the pressure melting point because the extra heat required to maintain this temperature comes from the latent heat of freezing. Figure 66 shows Weertman's graphical summary of the relationship between ice thickness and temperature and the predicted basal condition.

According to the curves in Figure 66, in the higher parts of the Marr Ice Piedmont, the situation is marginal in the context of that theory. The ice is about 600 m. thick and the 10 m. ice temperature is about -5 °C. Thus it is conceivable that in the high interior the ice could be frozen to the bottom but the limitations of the data (and the theory) do not allow this to be a definite conclusion. Also, to have the peripheral ice frozen to the bed, one has to invoke lower ice temperatures than those which were actually observed. This could plausibly be done by considering that in this zone the formation of (superimposed) ice at depth prevents the downward percolation of summer meltwater to appreciable depth and that the resultant warming effect is not operative at the base of the ice in this area. To some extent the temperature measurement of -0.8 °C at the base of the ice in the cave behind Palmer Station supports this idea as does the fact that neither this author nor Honkala's party observed meltwater issuing directly from beneath the peripheral ice. Thus it is conceivable that the upper layers of the peripheral ice are at a slightly higher temperature than the lower ones.

However, one objection to the "Freezing Model" theory is that it assumes that the heat of sliding is equal to the geothermal heat and that the geothermal heat is equal to 39 cal/cm^2/yr. If this is an overestimate and less heat is passing through the base of the ice cap than the theory presumes, then a supposedly melting situation might be transformed to one of basal freezing. However, according to the limited geothermal data available, there seems little cause to doubt that 39 cal/cm^2/yr is a reasonable value to take. In a review of the data Bullard (1954) found a worldwide average of 1.23 x 10^{-6} cal/cm^2/sec or 38.79 cal/cm^2/yr.

Without raising the issue of glacier surges and their possible causes, the proposed zonation of the Marr Ice Piedmont can best be discussed in the following terms. Consider that over the piedmont the conditions of temperature and rate of ice accumulation, in the absence of movement, require that the ice be everywhere frozen to the bed, because if no sliding occurs, no heat of friction will be produced by this mechanism. But once the piedmont begins to slide the heat of friction will maintain the basal temperature at the pressure melting point and the sliding will continue. Thus the three-fold situation on the Marr Ice Piedmont is that in zone I the low sliding velocities are producing just sufficient heat of friction to maintain the basal ice temperature at the pressure melting point to allow the "slip-and-stick" melt-regelation mechanism to operate. In the ice streams the sliding velocity is so great that the heat of friction is high enough to lead to basal melting which regenerates the ice's ability to keep on sliding continuously. If the rate of sliding here were somehow reduced, the melting condition might be halted and the melt-regelation mechanism instituted. In zone III sliding has not been induced and no heat of friction is available to raise the basal temperature to the pressure melting point. If, somehow, the peripheral ice were induced to slide, the heat of friction would keep the basal ice at the pressure melting point and the sliding would continue.

CONCLUSIONS

A limited theoretical analysis of the field data has been made. The calculated basal shear stresses are similar to those obtained from other glaciers and are acceptable. The relative distribution of ice velocity due to internal deformation and that due to basal sliding has been deduced and though there are certain inconsistencies in the results it has been possible to hypothesise that the basal condition of the piedmont is not everywhere the same. Parts of the peripheral ice appear to be frozen to bed while in much of the interior, limited basal sliding has been inferred. It is possible, though not confirmed, that melting at the base of the ice streams may be occurring. The configuration of these basal conditions apparently does not conform to the "Freezing Model" proposed by Weertman (1961) for the Barnes Ice Cap, which is a plausible theory for basal erosion by that ice cap. It remains therefore, to consider the possibility of basal erosion by the Marr Ice Piedmont in the light of its basal condition.

CHAPTER TWELVE

BASAL EROSION BY THE ICE PIEDMONT

EVIDENCE OF EROSION BY THE PIEDMONT

One of the problems posed at the outset of this thesis concerned the erosive activity of the piedmont and it has been emphasised elsewhere in this text that visible evidence of erosion is quite lacking. Nowhere do the ice cliffs contain rock debris and there is a remarkable absence of moraines of any kind. Yet in direct conflict to these observations, numerous masses of debris-rich ice, contained in floating icebergs, were observed floating offshore of the piedmont. One such sample, shown in Figure 67, was taken near Arthur Harbor in March 1966 and later studied at Palmer Station. The exact location of origin of this specimen is unknown but from observations of sea currents, local winds and the drift pattern of sea ice at the time, it is believed that it originated in the Wylie Bay area to the north of Palmer Station. The specimen was studied both macroscopically and microscopically and within the limits of the little that is known concerning the exact mechanisms involved in basal erosion, the study suggests a tentative hypothesis on basal erosion by the Marr Ice Piedmont.

Physical Characteristics in Hand Specimen

a) The original iceberg specimen can best be described as a distinct, richly loaded dirt band enclosed in a mass of clean ice (figure 67).

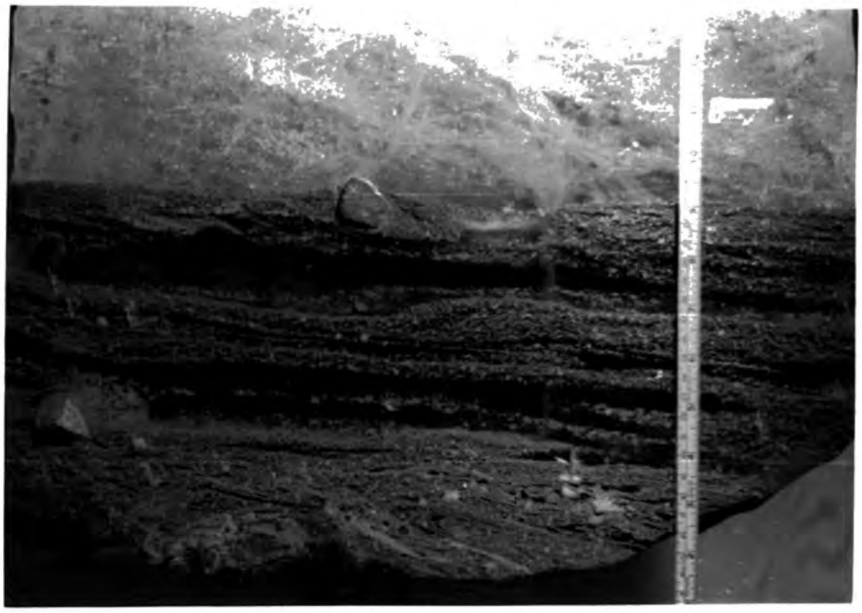
b) The dirt band was divided into a series of roughly parallel dirt layers, each layer being separated from the next by clean, completely

Fig. 67. Dirt band enclosed in clean ice. Thickness of the band about 1 meter. Scale: Climbing rope is 3/8" diameter.

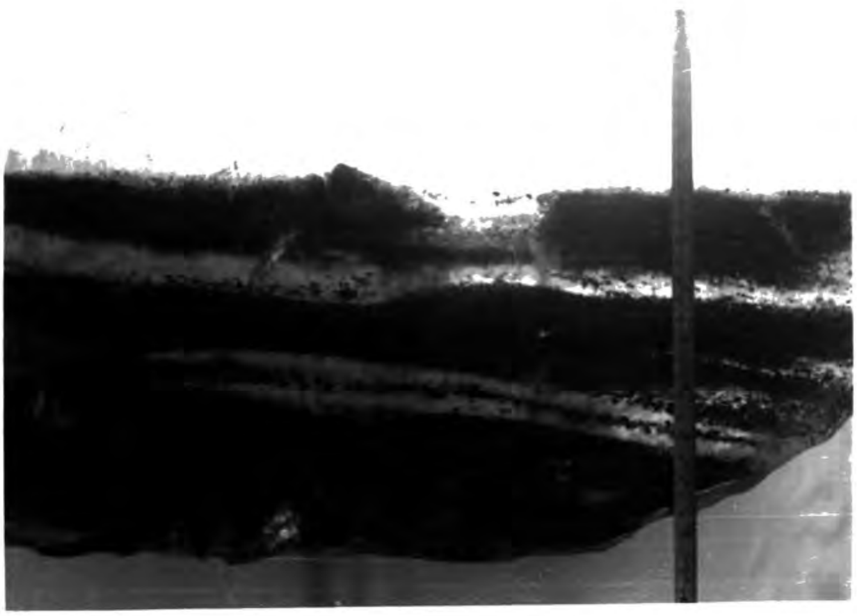
bubble-free ice (figure 68).

c) There was a slight difference between the "upper" and "lower" boundaries of the main dirt band; one was more regular than the other. The real orientation of the original specimen within the glacier is, of course, unknown but the more regular boundary has been taken as the lower one for descriptive purposes.

d) The clean ice "above" the dirt band was extremely bubbly, there being a slight banding to the bubble foliation. The bubbles were generally spherical or nearly so, and varied greatly in size. There were no significant elongated or tubular cavities. The clean ice"below" the dirt band showed similar characteristics and the only significant feature in connection with this portion of the ice was that the contact



a)



b)

Fig. 68. Ice sample showing dirt band and clean bubbly ice "above". Band is divided into dirt layers with clean, bubble-free ice between. Photo a) in reflected light. Photo b) in transmitted light.

with the debris ice was extremely weak. All attempts to cut sections which included both sections of ice were unsuccessful due to the failure of this contact.

e) The only absolutely bubble-free ice was that between the individual dirt layers within the main dirt band.

f) The dirt layers themselves were composed of "planes" of spherical, or nearly so, concentrations or "globules" of dirt material when viewed in a three dimensional sense. When viewed in section they formed lines of dirt globules. Undulation of the planes of globules was evident when viewed in sections cut at right angles to each other.

g) Over twenty samples of varying size were chopped from the original specimen from random locations for determination of the dirt content. This proved to range from 13% to 19% by volume.

Specimens in Thin Section

Extreme difficulty was encountered in working with this material. Thin sectioning was almost impossible and the few sections which were thin enough to be studied between crossed polaroids were small and fragmented. A single usable thin section was finally obtained from the sample shown in Figure 69 and is shown in Figures 70 and 71.

The structural characteristics of this specimen are intriguing but are difficult to understand or explain within the limits of present-day theory on glacial erosion processes.

Synthesis of Erosion Theory

The exact mechanism by which a glacier erodes its bed remains to be found and the overall process may well consist of several interacting

Fig. 69. Debris-rich ice sample. Arrows indicate the plane of thin sections shown in Figures 70 & 71. Note also rock inclusions within two dirt layers. Alignment of rock axes suggest ice movement into the plane of the paper or parallel to its long axis.

mechanisms. Weertman's "Freezing Model" discussed above, page 238, envisages an "en masse" pickup of basal debris as a result of shifts in the position of the zero degree isotherm at the base of the ice and the subsequent incorporation of debris by freezing to the base. Boulton (1970; 1972) also favors a form of Weertman's hypothesis for the incorporation of basal debris in Makarovbreen, Svalbard. From their direct observations of basal sliding, Kamb and LaChappelle (1964) described a "regelation layer", the formation of trains of bubbles and the inclusion of small amounts of fine rock material at points where the pressure-melt-refreeze mechanism was observed. The regelation layer was structurally and texturally distinct from the ice above and was composed of very fine grains. The report leaves no doubt about the existance of a

Fig. 70. Thin section of ice with dirt layer photographed in transmitted light.

Fig. 71. Same thin section photographed between crossed polaroids. Note the extremely fine crystal texture of the ice with the dirt layer.

basal pressure-meltwater layer and suggests that erosion may well be associated with the sliding phenomenon. However, the amount of debris actually incorporated through this mechanism is small and any debris layer created will be regularly destroyed, together with the regelation ice, by melting against further downstream obstacles. Therefore, the development of highly concentrated debris loads by this mechanism seems unlikely.

Erosion by cold sliding ice has been discussed by Lister (1972), who in laboratory experiments considered that rock wear by sliding ice is a function of friction. However, the amount of wear (erosion) was small and was quickly curtailed by the early formation of a deformation layer of crystallographically oriented ice, with c-axis normal to the surface. Ice so oriented has a very low coefficient of friction (Lister 1972). Frequently it was found that the sliding motion left ice adhering to the rock, producing an ice to ice sliding surface. Though the cold sliding ice showed lines of rock particles analogous to glacier dirt bands the production of rock debris was small. However, in an extension of the experiment, the shattering of pebbles was readily observed after as few as seven cycles of freeze-thaw. Thus, sliding cold ice, without the intervention of the liquid phase, may have appreciable erosive ability. However, Lister's conclusions were that freeze-thaw processes are at least as effective in erosion as cold sliding ice but that a freezing process seems more effective in the production of rock debris.

HYPOTHESIS RELATING TO THE MARR ICE PIEDMONT

In the light of observational and experimental evidence the main question surrounding the problem of erosion is the mechanism by which very substantial amounts of debris are incorporated into the basal ice and to establish the relationship between basal erosion and the basal sliding phenomenon. The observations of Kamb and LaChappelle (1964) appear to relate erosion and sliding by the freeze-thaw mechanism but the amount of debris incorporated is small. Weertman's (1961) hypothesis remains to be proven though Lister (1972) did find that layers of sand and gravel in an insulated container were disturbed and buckled after several freeze-thaw sequences. Thus, the weight of evidence at present seems to point to a relationship, complex though it might be, between sliding and erosion and that at least an intermittent liquid phase and a refreezing cycle are necessary for substantial pickup of basal debris.

Taking this as a basis for discussion, it is possible to re-examine the Anvers Island debris-ice samples described above and to present a hypothesis, albeit tentative, concerning the formation of the debris sample and to extend this hypothesis to the present erosive activity of the Marr Ice Piedmont.

Several features of the debris-ice specimen are of particular interest. The first impression is that the dirt band as a whole is completely bubble-free and that the clean and bubbly ice surrounding the dirt band is unaltered glacier ice. Secondly, is the problem of the alternation of the dirt layers and the completely bubble-free ice (see figures 68 & 69). Thirdly, is the extremely weak contact between the dirt band and the clean ice "below". Finally, there is the remarkable difference in grain size and texture between the dirt layer-ice and the clean ice surrounding it (figures 70 & 71). These characteristics are strongly reminiscent of the observations of Kamb and LaChappelle (1964) in Blue Glacier. It is hypothesised here that the dirt layers are in fact bubbly but every bubble is filled with dirt! The clean ice between the layers and which is bubble-free and is in weak contact with the dirt layer, represents the "spicule" ice described by Kamb and LaChappelle (1964). Within this hypothesis, these characteristics suggest that the debris-ice specimen is the result of a freeze-thaw pressure-melt sliding mechanism similar to that proposed by Weertman (1957;1964) and Llyboutry (1968a; 1968b) and that the very fine grained dirt layer is regelation ice, with the fine lineation of (dirt filled) bubbles or "globules" having been caused by pressure-melt regelation ice formation. Each single line of bubbles represents a single pressure-melt-event location and the planes of bubbles constitute an areal extension of these event locations. Perhaps a major flaw in this hypothesis is the relatively high concentration of dirt, up to 19% by volume as compared with no more than 10% reported by Kamb and LaChappelle (1964).

Weertman's (1961) "Freezing Model" concept also emphasises the importance of a basal water layer in the erosion of the bed but it is difficult to reconcile the "en masse" concept of debris pickup with the actual appearance of debris-laden ice. Perhaps a reworking, as a result of movement, and Weertman's inference of a "sludgy" bottom in some parts of the glacier could have led to the structure of the Anvers Island specimen. A similar or related hypothesis was put forward by Schytt (1963) in a discussion of fluted moraines in Norway. He suggested the presence of fluid moraine at the pressure melting point which freezes to the base of the ice when it enters a subglacial cavity and the pressure is released. This could certainly lead to an enhanced debris load in the basal layers, which include large rock fragments (see figure 69).

Thus there seems to be a strong possibility that a basal water layer must exist if there is to be substantial debris pickup at the base of the ice. This may take the form of a continuous water layer under at least a part of the ice as Weertman (1961) has suggested or it my be a transient phenomenon at specific locations in conformity with the qualitative principles of the sliding theory.

One important fact seems evident. If the basal condition is constant over all the glacier bed then there will be no substantial debris pickup. This applies equally to a base which is frozen at all locations or a base which is melting at all locations. The liquid water must refreeze at some point at the base if it is to incorporate dirt material.

Accepting this premise as the foundation of a hypothesis, the distribution of basal condition within the piedmont, suggested in Chapter 11, implies that some basal erosion is taking place. There are two zones within the piedmont where basal sliding has been inferred. In zones I and II basal erosion could be taking place but the conflict between visible and theoretical evidence needs to be resolved.

In zone II, in the highest parts of the piedmont, it is suggested that the erosion process is active in the pressure-melt-regelation mechanism and small amounts of debris are being taken into the basal ice. The erosion at individual locations will be small and considerable reworking will be required before the basal debris load becomes significant.

In zone I, the ice stream, basal sliding has also been inferred but here under different conditions from zone II. In this zone where the sliding velocities alone reach to well over 100 m. per year in places, the heat of friction could possibly be great enough to lead to basal melting. Under these conditions it is not possible for the basal ice to incorporate debris, therefore no erosion is taking place. This however, could be an overestimate and the possibility that zone I is in the same state as zone II, with some erosion taking place, should not be overlooked.

In zone III, no sliding is taking place, therefore there is no erosion because the ice is frozen to the bed.

The conflict between these conclusions and the visible evidence of the clean ice cliffs, which denies erosion, can be readily resolved. Consider that the three zones do exist and that zone I is not in the same state as zone II. The basal debris picked up in zone II in the regelation ice is transported, with the possible exception of station T5, in a southwesterly and southerly direction, following the vectors shown in Figure 30, p 99 , and discussed in Chapter 5. This ice ultimately enters zone I. It may do so in the vicinity of station C4 if zone II terminates at this point or at about station C1 if zone II extends southwards, as indicated by the dashed line in Figure 61. In any case, the fate of the ice is the same. It enters the rapidly moving ice stream and the basal ice begins to melt and at least some of its basal debris load is lost. Even the ice passing through stations P and N, for example, ultimately enters zone I, even though it may have been in close proximity to zone III and entered zone I near station K2. The remaining debris-laden ice is then channeled over the sub-sea level glacier bed (see figure 63) where basal melting may be further enhanced. Whatever debris-ice survives and reaches the coastal ice cliff, obviously must be below sea level.

Even if zone I is not a zone of basal melting and is in fact actively eroding the subglacial valley floor, the visible evidence is still below sea level when it reaches the coast. Under this set of conditions the ultimate debris load at the cliffs could possibly be considerable. In the northwestern part of the study area, similar arguments apply to the Wylie Bay ice stream, from which it is postulated the debris-ice specimen originated. The ice immediately behind Arthur Harbor may also fall into zone I, especially if, or where, the subglacial bed is below sea level.

Discussion

Though there is a general paucity of information regarding the debris load carried by various glacier types, Boulton (1970), though contested somewhat by Andrews (1971; 1972) suggested the generalisation that many polar and sub-polar glaciers, such as those in Svalbard, Baffin Island, Greenland etc carry very considerable amounts of englacial debris derived from the glacier bed, whereas temperate glaciers such as those of Norway, Iceland and the Alps carry very little basally derived debris.

However, the sub-polar category (where the sub-polar classification is derived from temperature observations which do not reach to great depth) seem to vary in their erosive ability which may be a direct reflection of differences in their basal condition. The Barnes Ice Cap is known to be frozen to its bed around the periphery but it has produced the Thule-Baffin moraines. It also has a rich dirt content as seen in the cliffs and so strikingly portrayed in the photographs in Ward (1952, p. 19).

A liquid phase at the base in the interior parts of this ice cap seems plausible but has yet to be proven. Weertman's "Freezing Model" is based on this ice cap.

The ice piedmonts of the western coast of the Antarctic Peninsula present a slightly different situation and from the observations on Anvers Island, it is obvious that the Marr Ice Piedmont is very different, geophysically, from the Barnes Ice Cap. The same might apply, in fact probably does, to the Fuchs Ice Piedmont on Adelaide Island. Whether or not the Freezing Model is supported by evidence from the Barnes Ice Cap, it certainly does not exist in the Marr Ice Piedmont. The high rates of accumulation, the subglacial topography, the ice streaming effects and the inevitable intervention of basal sliding to maintain equilibrium, raises the temperature of the ice streams to a degree where they are probably temperate. The sub-polar classification is probably applicable to the highest interior parts of the piedmont where the 10-meter temperatures range around -5°C. The peripheral ice standing on the subglacial saddle, on the basis of temperature data available, apparently lies between the two classifications. However, more observations, to greater depth are needed to fully understand this part of the piedmont. Thus, though there are zones of melting and freezing at the base, their position, relative to each other, is adjacent rather than concentric, as supposed for the Barnes Ice Cap. Consequently, the Marr Ice Piedmont does not pick up the massive amounts of basal debris seen in some sub-polar glacial features, even though it may, in parts, be truly sub-polar. The ice streams "shear" alongside the zone of basal freezing rather than flow towards it. Consequently, the ice over the subglacial saddle can remain frozen to the bed because velocity enhancement by basal sliding is not needed to maintain equilibrium.

CONCLUSIONS

The argument has been put forward that basal sliding and basal erosion are related interacting phenomena and taking this as the basis for hypothesis, it is concluded that the Marr Ice Piedmont is selectively eroding its bed. This process is active over the high interior subglacial plateau, but as a result of ice streaming outward from this area it is possible that intensive erosion at least, of the subglacial valley floors, is prevented by basal melting. This leads to the conclusion that englacial debris derived from the interior plateau may be redeposited along the valley floor or slowly transported outward towards the coast by subglacial melt streams. On the other hand, it should not be overlooked that though the sliding velocity of the ice streams is high, a pressure melting-regelation mechanism could be active leading to erosion of the valley floors. In any case, any visible evidence of erosion in the coastal ice cliffs may be small and is certainly below sea level.

The field evidence leads to the conclusion that over the subglacial saddle between the two valleys, there is little or no erosion because the ice here is frozen to the bed. Certain inconsistencies in the data however, indicate that there might be isolated areas of basal sliding leading to localised erosion over the saddle. Information on the subglacial topography in this area lacks detail and is inadequate to confirm localised sliding and erosion.

There is no evidence, either from the geomorphological appearance of the subglacial platform or from the preceeding discussion, that the Marr Ice Piedmont is a "strandflat glacier" within the context of Holtedahl's (1929) description. The piedmont is not at present cutting, and may never have been cutting, a planed surface at a level controlled by the sea. Indeed, it is more likely that it is in the process of destroying the planed surface by selective erosion which is accentuating the topography of the platform.

CHAPTER THIRTEEN

GENERAL SUMMARY AND CONCLUSIONS WITH SUGGESTIONS FOR FUTURE WORK

As a pilot study to elucidate the basic glaciological characteristics of the Marr Ice Piedmont, the present investigation was sufficiently comprehensive to be adequate. However, like many investigations which preceded it in the Antarctic Peninsula, it was a localised study, in many ways unrelatable to the major, contemporary geographical patterns in the region. It has learned much about the Marr Ice Piedmont and is the first major undertaking of its kind in the peninsula area. It has prompted considerable speculation regarding many aspects of its findings and in many cases mere speculation and suggestion is all that is possible because of the general lack of relatable information from elsewhere in the peninsula. The Antarctic Peninsula is a distinct geographic entity and because most of the immediate problems in the peninsula have a geographic connotation, they have a profound regional as well as local importance. Therefore, future work, as revealed by the present study, should not be restricted to Anvers Island alone. Above all, the present study has set a firm foundation upon which future work can be developed.

From the mass balance investigations, it is concluded that the Marr Ice Piedmont is in equilibrium or very slightly positive and evidence from the ice ramps tentatively suggests that the piedmont may have been experiencing a prolonged state of slightly positive regime, after a period of recession. Certainly there is no evidence, either glaciological or geological, to suggest a present state of recession. Thus, the Marr Ice Piedmont appears to be typical of the glaciers of the peninsula's west coast.

These measurements were adequate as a preliminary study but, as with meteorology, the longer the record, the more satisfactory the result. This is very significant when there is the likelihood of marked annual variations in the balance factors. It is suggested that mass balance studies should be resumed at Anvers Island and continued for several years. Such studies should be intimately coupled with glacier heat balance and energy exchange studies, which would not only lead to a better understanding of the physical characteristics of the piedmont, but would also help in elucidating the many accumulation and ablation factors in the mass balance equation and for which, admittedly, this study has not been able to account. As the evidence points to the ice ramps as important mass balance indicators, a resumed study should direct specific attention to them. The question of a recent and continuing positive regime should be further investigated at Anvers Island and similar evidence should be sought elsewhere.

With the foundation set by the present study, and despite the arduous field-working conditions, the Marr Ice Piedmont is a feasible long-term mass balance study area. Certainly the precedent for long-term studies, under equally arduous conditions, has been set in the Arctic, for example, Müller 1963; Hattersely-Smith & Serson 1970; Iken 1972; Müller & others 1972.

A resumption of studies at Anvers Island together with similar supporting mass balance studies at selected locations in the Antarctic Peninsula, with special attention to the little-known eastern side would be ideal. The overall objective of such a regional study, performed perhaps on an international basis, would be to determine the state of the balance equation as it applies to the peninsula as a whole, to describe and explain local and regional variations and to catalog the glaciological transition from the northern temperate to the southern cold-polar environment. Such studies would not only be complementary to those in the Arctic and elsewhere, but with a regional approach would have particular value and significance in the light of the objectives of the International Hydrologic Decade.

Of all measurements during this study, those of ice temperature were the least adequate and temperatures from greater depths are needed before any firm conclusions can be made regarding the geophysical state of the piedmont. On the basis of temperature data now available, the piedmont lies, geophysically, between temperate and sub-polar.

Precipitation over Anvers Island is high, amongst the highest on record for the Antarctic Peninsula but is not anomalous in this central-southern section. Over the Marr Ice Piedmont, precipitation increases rapidly with elevation and varies locally with topographic irregularities. At 850 m. elevation precipitation exceeds 250 g/cm^2/yr. There is no significant variation in precipitation with change in latitude except that the southern lee slope receives the least because the principal snow-bearing winds are north-northeast and easterly. These winds and the physiographic configuration of the island appear to create a precipitation shadow, which may be effective over the entire piedmont.

Pronounced annual variations in precipitation are evident and are probably related to geo-climatic factors which this study has not been able to determine. Future work, particularly if it is on a regional basis, should direct specific attention to the interrelationship between locally determined mass balance factors and the regional distribution of geo-climatic factors.

The piedmont ranges in thickness from 60-80 m. at the coastal cliff to more than 600 m. inland at the foot of the mountains. It rests on two almost horizontal low-level platforms, one peripheral at approximately 50 m., the other at about 200 m. elevation. In places these surfaces are dissected by valleys, resulting inpronounced ice streaming at the surface. It is concluded that this subglacial topography is not a strandflat according to the classical definition, though the amplitude across the surfaces appears to be no more than about 40-50 meters.

Surface ice velocities were obtained using open traverse survey techniques with electronic and optical equipment. It can be concluded that with sufficient observational control, this method of movement determination is satisfactory and to be recommended for such studies when solid rock control is not available. Surface ice velocities range from 14 m/year in the high interior to 218 m/year in the ice streams.

To reconcile the distribution of measured surface velocities and the distribution of calculated basal shear stress and inferred sliding velocities, differences in basal condition from one part of the piedmont to another has been invoked. It is believed that a slip-and-stick sliding mechanism is operating in the higher interior parts and that much of the peripheral ice is frozen to bedrock. In the ice streams the inferred sliding velocity exceeds 100 m/year in places suggesting the possibility of basal melting. Alternatively, an enhanced slip-and-stick mechanism may be operating in the ice streams.

With the conclusion that there is a direct relationship between basal sliding phenomena and glacial erosion, the zonation of basal conditions in the Marr Ice Piedmont has been related to its erosive capability. Basal erosion is active in the interior by the slip-and-stick sliding mechanism while no basal erosion is occurring below much of the peripheral ice because it is frozen to bedrock. The basal condition of the ice streams is uncertain but if melting occurs beneath them, they are not erosive. Alternatively, a slip-and-stick mechanism accounting for a sliding velocity of as much as 100 m/year leads to the conclusion that the ice streams could be extremely erosive. Whereas the remarkably clean ice of the peripheral cliffs suggests no erosion, it is further concluded that the evidence of erosion does exist but is, by virtue of the subglacial topography, below sea level when it reaches the coast.

On the basis of this information, the present study finds insufficient evidence to support the belief that the present ice cover has eroded its subglacial surface in the way described by previous observers. Thus, if the Marr Ice Piedmont, as it now exists, is not wholly responsible for the construction of its subglacial surface, its origin by other mechanisms must be considered. It is beyond the scope of this thesis to pursue this problem but this author has suggested elsewhere that the surface beneath the Marr Ice Piedmont, and possibly the surfaces beneath other glaciers and piedmonts throughout the Antarctic Peninsula, are preglacial and of initial marine origin, later modified by glacial action and that there may be a regional occurrence of elevated surfaces, similar to that upon which the Marr Ice Piedmont rests (Rundle 1971). This is an interdisciplinary problem but basically one of geology and geomorphology. Certainly it is a regional problem and forms an important part of future work in the Antarctic Peninsula.

The subglacial surface of the present study area should be further investigated in detail by an intensive gravity survey, supported by seismic control, and extended throughout Anvers Island to determine whether the known subglacial topography is a local or widespread feature. The form of the inland subglacial continuation of the Cape Monaco promontory should be determined and the subglacial morphology of the area between Cape Monaco and Hamburg Bay should be investigated. From this, a more complete knowledge of the subglacial geomorphology will be available and a more concise evaluation of its history will be possible. The morphology of the subglacial surfaces of other piedmonts and glaciers throughout the peninsula should be studied through gravity and seismic investigations to examine the possibility of a regional occurrence of elevated surfaces, particularly at about 200 to 300 meters a.s.l. and to further explore the question of their marine origin and age. Such a regional study would be helpful towards a better understanding of the structural history of the Antarctic Peninsula.

Elsewhere in this thesis this author has expressed slight dissatisfaction with the ice temperature measurements taken during the study and that future work should concentrate heavily on this problem. The internal distribution of temperature from the surface to bedrock should be fully and accurately investigated on the piedmont. Ideally, this should be done at at least three locations; one in each of the three suggested basal zones. These studies would determine the true geophysical character of the piedmont and determine whether or not there are thermal differences from one part of the piedmont to another. Such a study, coupled with what little is understood concerning the relationship between glacier thermal characteristics, the basal erosion process and the amount of basally derived debris carried by temperate and non-temperate glaciers, would be a most logical extension of the present study and would be an important test of the main thoughts expressed in this thesis. Additionally, the bore holes resulting from such an investigation would be of considerable value for deformation studies, especially in the ice streams, where activity is high, and would lead to more precise values for the constants in the deformation equation as it applies to this particular glacier morphology. Cores taken during the drilling operation would lead to valuable information on the recent history of the piedmont and ice samples from the basal layers would attest to the erosiveness of the piedmont and would permit further, detailed study of the basal erosion process.

From this study it is not possible to date the present ice cover. It is probably a survivor from the most recent glacial recession, though in its present highly active state it seems quite capable of independent existance, and theoretically could be merely centuries in age. Assuming that the ice streams are transporting debris, marine sediment studies, close inshore where the streams reach the coast, would be helpful toward establishing the age of the ice and could possibly shed considerable light on the recent history of the area.

An interesting problem, not touched upon in the main text of this thesis, for lack of more than circumstantial evidence, concerns the ice ramps. Why does the piedmont terminate as ramps at such places as Norsel Point and Bonapart Point? Is there a delicate equilibrium situation which is maintaining the ramps in their present morphology? Or, ultimately, will they develop into land-based cliffs?

The ramp studies described in Chapter 7 indicate that in some years the budget is slightly positive while in other years it is slightly negative. In Chapter 8 the question was raised as to the relative frequency of positive and negative budget years, and it was suggested that there had been a recession of the ramps at some time in the past and that subsequently there had been a general tendency towards a higher frequency of positive budget years, reflected by the accumulation of superimposed ice on the old dusty ice surface. On the basis of these observations two hypotheses can be put forward.

1) The frequency of annual budget variations are approximately equal and this coupled with velocity factors is the controlling element in the maintenance of the ramp morphology. Stake number 5 on Norsel Point ramp (see figure 44) was surveyed by tape trilateration and showed a barely significant forward movement of 9 cm over a two-year period. Thus, the ramp at this point is virtually stagnant and probably is so

for a further 150-200 m. inland. Thus in negative budget years total net ablation reduces the lower ramp as it did in 1964 and 1965 but the loss is replaced during the positive budget years when the remaining winter accumulation forms very consolidated iced firn or superimposed ice which again is removed in subsequent negative budget years. Thus the surface level and the profile shape of the ramp fluctuate slightly year by year but forward motion is too small and positive or negative budgets do not occur consecutively over long enough periods to cause significant growth or recession of the ramps.

The virtual lack of forward motion is not only the cause for the ice not reaching the coast at this point but also is one reason why a land based cliff is not formed. The positive balance over most of the upper ramp is not brought forward to this point. The position and lineation of the crevasses on the ramp (see figure 44) indicate shearing of the ice and the channeling of it into Lowdwater Cove and Arthur Harbor. The foot of the ramp therefore, is maintained almost as a separate entity, somewhat divorced from the general activity of the main piedmont. The significant fact which emerges is that a ramp-shaped feature and not a cliff is apparently being maintained.

2) A second hypothesis is presentable in view of the present surface profile of Norsel Point ramp (see figure 45). This profile was obtained by precise level and staff survey by L.E. Brown in 1966. The survey reveals a very steep gradient from the edge of the ramp inland to the vicinity of stake number 8 where a sharp break of slope occurs. This is the point where the superimposed ice deposit begins. From this point the gradient of the ramp surface inland is much less.

Though most of the ice on the upper ramp shears into Arthur Harbor and Lowdwater Cove, and though the forward motion of the foot of the ramp is only 4.5 cm/yr. there may be forward motion generally parallel to the main axis of the ramp in this area and if this is the case, a possible interpretation of the gradients on the lower ramp is that the initial steeper gradient, from the edge of the ice to the first break of slope, represents a very shallowly inclined cliff face. Possibly on a long-term basis, the combined effects of the superimposed ice formation and the small but persistent forward motion, will exceed the overall effects of ablation and the steep ramp foot will increase in inclination to be restored to a true cliff configuration.

There are no direct field measurements to support this second hypothesis and Brown's survey is merely suggestive. However there is limited circumstantial evidence, but from simple field observations alone, that it could contain some credibility.

The cliff face surrounding Skua Lake behind Palmer Station is inclined at an estimated 30° from the vertical and has been observed by both this author and I.M. Whillans to fracture and collapse periodically indicating that some movement is occurring. Whillans' attempts to measure the movement along a vertical profile on the cliff face failed due mainly to the hazards created by the collapse. However, from his direct observations he feels that over a 13-month period, the face of the cliff became slightly steeper. This leads to the conclusion that the features of the cliff are becoming more strongly emphasised. Unfortunately, this author did not direct any special attention to the cliff in 1965 or 1966.

This is an intriguing problem and has implications concerning the present regime of the piedmont and its recent history and also the questions surrounding the problem of recent climatic changes in this central-southern part of the Antarctic Peninsula. A resumption of long-term mass balance studies on the ramps should include precise profile surveys and direct attention to the questions raised by these two hypotheses.

Finally, a major regional question in the Antarctic Peninsula relates to its history and particularly to the sequence of events during the Pleistocene. These problems have recently been approached by Clapperton (1971) and Everett (1971) but so far only one tentative glacial chronology has been published (John & Sugden 1971). Widespread evidence establishes that the ice has been at least 300 m. thicker than at present and it is probable that during the maximum glaciation the ice was much more extensive. It may have been entirely land based, during an eustatically lowered sea level, or it may have taken the form of a locally anchored ice shelf. Much of the evidence of a former ice cover, and this certainly applies to any previous extension of the Marr Ice Piedmont, is hidden due to the recent submergence of the land. Submarine geomorphological investigations and sediment studies would be instrumental in determining the former extent of the ice and would provide valuable information for the formulation of the regional glacial chronology.

In the Antarctic Peninsula there are many problems to be solved. They are not isolated problems and they all are enormous in their scope and magnitude. In many ways the region is but little known to science. This thesis has attempted to scratch the surface of some of these problems and in so doing it has raised even more. It is this author's hope that the work of his collegues and himself has formed a worthwhile contribution toward a better understanding of the Antarctic Peninsula and that it is a firm foundation for the work which remains.

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