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GRAVITY AND MAGNETIC SURVEYS IN GRAHAM LAND

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ABSTRACT

BETWEEN 1959 and 1965 gravity and magnetic surveys were undertaken in Graham Land by geophysicists of the British Antarctic Survey. A Worden gravimeter, Askania vertical-field magnetometers and proton total-field magnetometers were variously used for the observations. Several thousand kilometres of traverses have been completed in regional investigations in addition to more detailed local surveys. In the final data analysis, available and relevant aeromagnetic and marine geophysical profiles have been incorporated, and radio echo-sounding techniques have provided essential ice thicknesses. Two- and three-dimensional computer techniques have been applied in the interpretation of the residual gravity and magnetic anomalies.

The majority of the magnetic anomalies over the Trinity Peninsula area are attributed to the underlying, sometimes exposed, intermediate to basic representatives of the Andean Intrusive Suite, here believed to form an extensive system of batholiths. The presence of the Andean Intrusive Suite is again suggested for magnetic anomalies south of lat. 68° S., where the overland traverses are supplemented by certain of the Project Magnet aeromagnetic profiles. There is a possible north-west to south-east correlation between the profiles, indicating a sub-surface structural continuity. Provisional profile interpretation shows a sub-ice topography similar to that recorded at the southern end of the Antarctic Peninsula. Normally magnetized Tertiary olivine-basalt lavas and dykes provide the source rocks for the distinctive magnetic pattern observed at the Seal Nunataks. All of the nunataks visited have a similar associated magnetic pattern, and vertical-field measurements in excess of 6,000 gamma have been recorded. Adjacent to the west coast of Graham Land the marine and aeromagnetic profiles indicate a consistent change in the magnetic pattern. The magnetic intensity is much greater and in sharp contrast to the short wave-length anomalies found on the mainland, and to the magnetically quiet Larsen Ice Shelf. Basic igneous rocks exposed on the offshore islands and proposed faulting off Cape Legoupil are both reflected in the geophysical results.

Graham Land is associated with a negative Bouguer anomaly reaching a minimum value around lat. 68° 15′ S., and crustal thickening is the expected cause of this. Beneath the Larsen Ice Shelf, localized negative gravity anomalies, indicative of areas of glacial overdeepening, are associated with many of the inlets; sea depths approaching 900 m. have been calculated in Trail Inlet. A large negative anomaly over Duse Bay has yielded depths to the sea floor of over 2,000 m., a figure far in excess of any previously encountered around Graham Land. Positive gravity anomalies adjacent to promontories and peninsulas have been attributed to areas of grounded ice. Low-density volcanic rocks are suggested sources for the negative anomalies over Jason Peninsula and Tabarin Peninsula, whereas the more dense Trinity Peninsula Series sediments may be the cause of the positive anomaly over eastern Joerg Peninsula. Geological and

geophysical evidence indicates that the anomaly at the northern end of Churchill Peninsula is due to an underlying gabbro–microgranite laccolith.

Structural implications from the geophysical data support the theory that Neny Trough is a major tectonic feature with much of the Gibbs Glacier bedrock lying below sea-level. Elsewhere in Graham Land, removal of ice would isolate Jason Peninsula from the mainland and show much of Churchill Peninsula to be submerged. Isostatic evidence is at present confused due to geological imbalance coupled with the relatively recent glacial recession, and to insufficient data coverage.

CONTENTS

I. INTRODUCTION

THE Antarctic Peninsula lies in western Antarctica and can be described as the area of the mainland curving northward for 1,200 km. from a line between Cape Adams and lat. 73° 25′ S., long. 72° 00′ W. to the vicinity of Hope Bay (lat. 63° 24′ S., long. 57° 00′ W.). It is further subdivided by a line, governed by a major topographical discontinuity between Cape Jeremy and Cape Agassiz, which separates Palmer Land to the south from Graham Land to the north. The geophysical investigations described in this report are almost wholly concerned with Graham Land (Fig. 1), though the surveys are continuing in Palmer Land. Graham Land itself consists of an ice-covered plateau 40–50 km. in width and with heights of 2,000 m. The west coast is characterized by a group of offshore islands paralleling the mainland and stretching from Adelaide Island in the south through the Biscoe Islands to the Palmer Archipelago. In sharp contrast, the concave east coast facing the Weddell Sea has a much less severe topography and, apart from the extreme northern area east of Trinity Peninsula, it is fringed by the Larsen Ice Shelf. To the north-west across Bransfield Strait are the South Shetland Islands, the southernmost island group of the Scotia Ridge, which extends to South America by a sinuous course through the South Orkney Islands, South Sandwich Islands and South Georgia.

Geological mapping and surveying have predominated in the initial exploration of the Antarctic Peninsula since work was commenced by members of the Falkland Islands Dependencies Survey in 1943. However, in 1959 geophysical investigations of the Scotia arc, including the Antarctic Peninsula, South Shetland Islands, South Orkney Islands, South Sandwich Islands and South Georgia, were commenced by the British Antarctic Survey in collaboration with the Sub-Department of Geophysics, University of Birmingham. The initial exploration of the area, using magnetic methods supplemented by land gravity measurements, was conducted from the Survey's ships, R.R.S. *John Biscoe* and R.R.S. *Shackleton*. The survey, which was directed by Professor D. H. Griffiths of the University of Birmingham, developed over the years to a fuller geophysical coverage and incorporated seismic reflection and refraction work. Results of the marine work have been given by Cox (1964*b*), Griffiths and others (1964), Allen (1966*b*), Ashcroft (1972) and Barker and Griffiths (1972).

Coupled with the geological work on land, the marine geophysics in the Scotia Sea began to add to the knowledge of the geological history and structure of the region, especially in regard to its association and affinities with the Scotia arc and South America. It was therefore essential that land geophysical investigations should be commenced in order that a greater understanding of this complex region could be achieved.

In 1959, a magnetic survey was commenced from the British station at Hope Bay and it was intended that this survey would ultimately tie in with the marine work already done in the vicinity, and in addition form the basis of a much wider investigation of the Antarctic Peninsula. Between 1959 and 1962, the magnetic surveys were continued systematically (Ashley, 1962; Cox, 1964*a*; Allen, 1966*a*). Subsequent to this, the land geophysical surveys, both from Hope Bay and Stonington Island (since 1963), have used magnetic and gravity methods for the investigations over the length of Graham Land (Renner, 1971). The present report deals primarily with this land work.

A. Survey Procedure

Geophysical investigations elsewhere in Antarctica have variously incorporated seismic, gravity, magnetic and topographical methods. From these, crustal information has been obtained either directly or from certain empirical relationships (Worzel and Shurbert, 1955; Woollard, 1959, 1962*a*; Frolov, 1965*a*, *b*, *c*; Demenitskaya, 1960). Crustal studies involving seismic methods in Graham Land have so far been confined to minor extensions of the marine investigations of the Scotia arc. Because the overland survey has been involved entirely with gravity and magnetic investigations, the control has not been available for deeper crustal studies nor has the quantity of data yet permitted the application of any empirical formulae. The present research is therefore primarily concerned with interpretation of near-surface geological features.

The gravity and magnetic traverses have been carried out using dog-sledge transport with navigation methods involving compass–sledge-wheel techniques, and resection from previously determined trigonometrical points. Where the existing 1:200,000 topographical maps proved to be inadequate, plane-table surveys were completed in order that station positions in areas of detailed survey could be more accurately

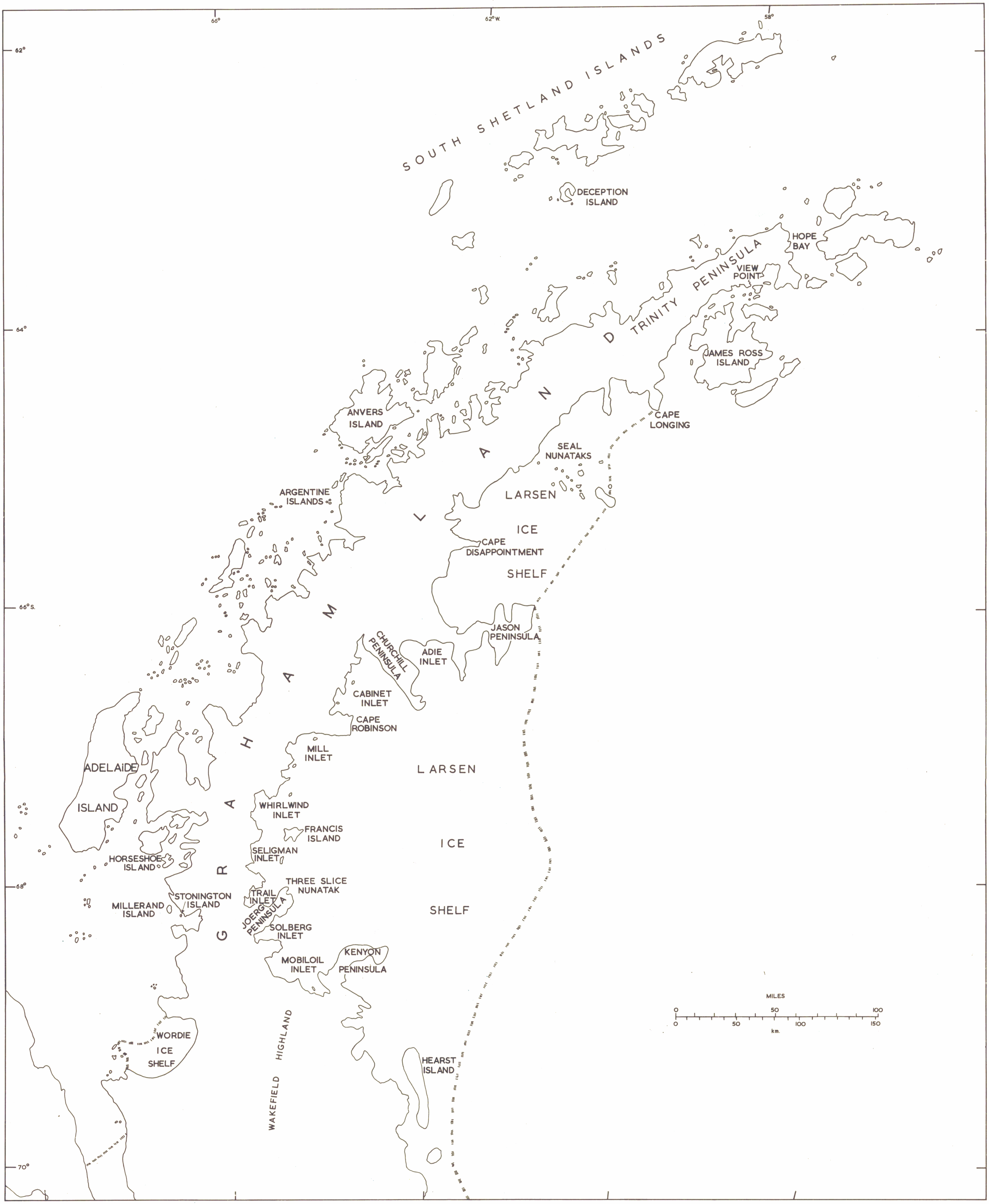

FIGURE 1
Location diagram for Graham Land.

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located. Station elevations, which provide an almost constant problem in Antarctic geophysical surveys, were determined by using a minimum of three Walker aneroid barometers, a method so far found to be the most practicable (Kennett, 1965*c*). Due to logistic problems, it was not possible to use either leap-frog or closed-loop techniques to improve the elevation accuracies.

Wherever possible, gravity and magnetic traverses have been run together though in several instances the instrumental difficulties and distribution of personnel have prevented such a simple generalization. For regional surveys, the average station spacing was 6 km., whilst for the detailed work a closer network was involved which was dependent on the problem being investigated. In all over 600 gravity and 1,500 total-field magnetic stations have been occupied on a regional basis. In addition, the greater part of Trinity Peninsula has been covered with a systematic vertical-component magnetic field survey.

Standard procedures in Bouguer, free-air and terrain corrections have been adopted in the reduction of the gravity data. In some cases, however, it has been necessary to infer ice thicknesses from field estimates together with information provided by radio echo-sounding surveys over the plateau and Larsen Ice Shelf (Renner, 1969; Smith, 1972). In Tabarin Peninsula, ice-thickness values indicated by Ashley (1962) were used in data reduction. Fortunately, where ice-thickness information is at a minimum, namely on the west–east plateau traverses, the anomalies are of such a magnitude that the errors do not affect the general anomaly trend.

The geophysical programme completed by the author from Stonington Island (1964–65) consisted of two field journeys. The first of these took place during April–June and lasted for 58 days, of which 45 days were spent working. The longer summer journey of 98 days duration over October–January included 31 lie-up days, the remainder being spent covering 2,100 km. of geophysical traverses; this compares with 670 km. covered on the earlier autumn journey.

Insufficient supply depots coupled with the time factor restricted the autumn journey to traverses radiating from Three Slice Nunatak. Average temperatures of $-22°$ C meant that magnetometer readings could only be taken at overnight stops when the instrument and cable could be warmed. The Worden gravimeter, on the other hand, was read as and when required, although below $-30°$ C the penlight batteries required intermittent warming. On the summer journey, working conditions were much more favourable, except during October when average temperatures of $-20°$ C again severely limited the efficiency of the magnetometer. For the remainder of the journey the temperatures fluctuated around $0°$ C, enabling magnetometer readings to be taken regularly. North of the Seal Nunataks, total-field magnetic measurements had previously been taken by J. Mansfield operating from Hope Bay; thus, in order not to duplicate this earlier work, the magnetometer was operated farther south by E. Thornton in conjunction with the geological parties. During this period the gravity traverses were continued north of the Seal Nunataks, where they were partially governed by the necessity of visiting the supply depots. Where possible, route deviations were undertaken, especially where the opportunity arose to run traverses at right-angles to the axis of the peninsula.

During the months of July, August and September, field work was operated locally from Stonington Island. The programme was two-fold and consisted of a series of magnetic diurnal measurements, later compared with those recorded at the Argentine Islands observatory, and to the evaluation of the gravimeter drift rates.

B. Previous Geophysical Results from the Antarctic Peninsula

Behrendt (1963, 1964) has completed seismic, gravity and magnetic investigations in Ellsworth Land and at the southern end of the Antarctic Peninsula, to the east and north of Eights station. Seismic reflection and refraction methods were applied (Behrendt, 1963) but between reflection stations free-air anomalies were used to calculate bedrock elevation. From the formula for the gravitational attraction of an infinite slab, and by using densities of 0·9 and 2·67 g. cm.$^{-3}$ for ice and rock, respectively, a theoretical value of 13·5 m./mgal is obtained and from this variations in bedrock elevation can be calculated. However, by applying a least-squares fit to a plot of differences in anomaly to known seismically determined variations in bedrock elevation, a value of 20 m./mgal was found to be a more realistic figure. This contrast with the 13·5 m./mgal was attributed to severe topographical variation below the ice and was not due to the use of incorrect densities. The results of the seismic investigations showed:

i. If the ice were removed, the Antarctic Peninsula would be an island separated from the Eights Coast by a narrow sub-sea-level channel, and from the Sentinel Range to the south by a broad channel over 1,000 m. below sea-level. Behrendt believed that this latter depression, after isostatic rebound in the event of the ice melting, would form a connection between the Ross and Weddell Seas.

ii. The discovery of a low-velocity layer 450 m. thick at the base of the 2,100 m. thick ice near Eights station. The low-velocity basal layer is thought to be the result of 13±3 per cent of included morainic material.

iii. About 300 km. east of Eights station an 18 km. refraction profile was recorded (Fig. 2).

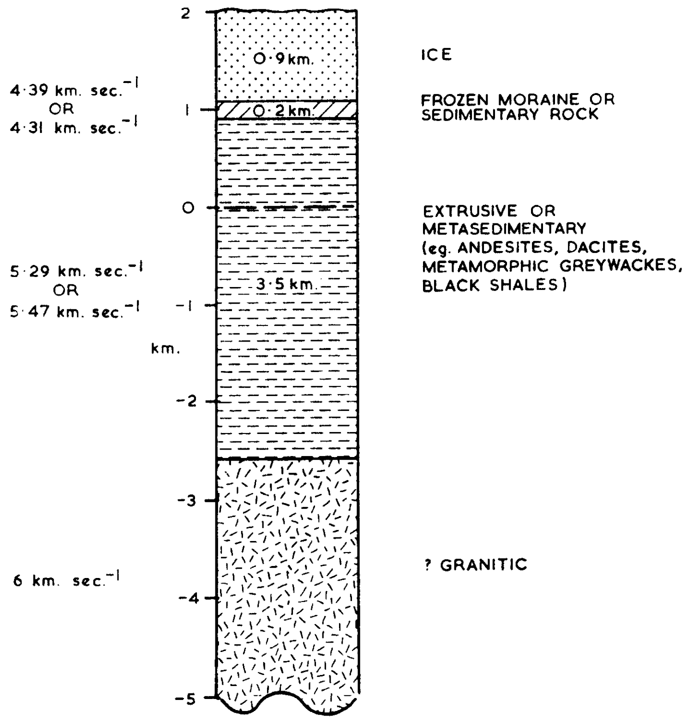

FIGURE 2

A representative geological section compiled from a seismic refraction profile (Behrendt, 1963) at the southern end of the Antarctic Peninsula.

Several samples of exposed country rock were collected in the area with representatives of Cretaceous sandstones, black shales, arkoses, conglomerates, argillites, dacites, andesites and metamorphosed greywackes (Laudon and others, 1964). As velocity measurements on these were all similar, it was not possible to distinguish between an extrusive or metasedimentary origin for the 5·29 km. sec.$^{-1}$ layer (Fig. 2). East of Byrd station and in the Sentinel Range 1·3 km. of rock with a velocity of 5·20–5·30 km. sec.$^{-1}$ is present (Bentley and Ostenso, 1961).

In the paper by Behrendt (1964), gravity and magnetic surveys over Ellsworth Land and the southern Antarctic Peninsula are discussed. Over Ellsworth Land, west of Eights station, the average free-air anomaly is only +11 mgal, which suggests the area is isostatically compensated. In sharp contrast is the area at the base of the Antarctic Peninsula east of Eights station where the average free-air anomaly is +60±30 mgal. The large scatter is indicative of an area of rough subglacial topography which is more probably compensated on a regional than on a local basis. Theoretical crustal models for this area support this conclusion. From the average Bouguer anomaly of +22 mgal in Ellsworth Land and a mean crustal thickness of 31·7 km. at sea-level, a crustal thickness of 31 km. was calculated using 2·86 and 3·31 g. cm.$^{-3}$ for average densities of crust and mantle, respectively. This contrasts sharply with the approximate value of 36 km. depth to the Mohorovičič discontinuity calculated for the southern part of the Antarctic Peninsula. Variations in crust and mantle densities could affect the estimated crustal thicknesses.

From the land magnetic and Project Magnet flight traverses completed over the Antarctic Peninsula, Behrendt (1964) produced residual total-field magnetic profiles showing numerous anomalies, several exceeding 1,000 gamma in amplitude. The regional field varied between 50,000 and 55,300 gamma. Using Peters' (1949) half-slope method, approximate maximum depths to many of the Antarctic Peninsula anomalies were calculated showing that 76 per cent had a source shallower than 1 km. below bedrock surface, while 52 per cent had their source shallower than 0·5 km. Apparent susceptibilities were calculated using the method of Vacquier and others (1951) which gave susceptibilities ranging from 0·0006 to 0·0147 c.g.s. units with a mean of $0 \cdot 0036 \pm 0 \cdot 0018$ c.g.s. units. Depth estimates indicate that the source of the anomalies may deepen to the south from an elevation of $+1 \cdot 5$ to $-5 \cdot 5$ km. The shallow "magnetic basement" over eastern Ellsworth Land at the base of the Antarctic Peninsula is thought to have its origin in the magnetite content of the dacites and andesites (Laudon and others, 1964), which are found in association with the steeply dipping metasedimentary rocks believed to extend throughout the area, and which may be correlated with the rocks of the Antarctic Peninsula (Laudon, 1971). Geological correlation along the length of the Antarctic Peninsula has also been provided by Williams and others (1971). The magnetic anomaly pattern favours the theory that the mountains of the Antarctic Peninsula and eastern Ellsworth Land are not necessarily continuous with those of the Ellsworth Mountains, as the anomalies associated with the latter are infrequent and with sources deeper than those associated with the Antarctic Peninsula.

The gravity and seismic investigations completed over the Filchner Ice Shelf (Behrendt, 1962) have primarily been concerned in delineating the marked topographical features of the bedrock. There is a deep trough bordering the eastern margin and an area of grounded ice over the more centrally located Berkner Island and Malville Peninsula. Over the western part of the ice shelf the ice is again grounded at Korff Island, but otherwise calculated depths to bedrock in the vicinity are about 500 m. An echo-sounder profile from Bowman Peninsula to Moltke Peninsula shows that immediately east of the Antarctic Peninsula depths are slightly greater than 500 m. but they gradually decrease to the east before the trough is encountered. Thiel and Behrendt (1959) have published vertical magnetic intensity field data from the Filchner Ice Shelf and this is referred to elsewhere (p. 57).

Crustal thicknesses over the Antarctic Peninsula have largely been extrapolated from surrounding areas, or they are based on the indirect approach of incorporating empirical relationships. From the study of dispersed Love and Rayleigh waves, Evison and others (1960) calculated an average crustal thickness in Marie Byrd Land of 25 km., but using the same data Bentley and Ostenso (1962) amended the value to 30 km. Using similar techniques, a depth to the Mohorovičič discontinuity of 35 km. was given by Dewart (1961) along the axis of the peninsula. Woollard (1962*a*), using normal empirical relationships and equivalent elevations, produced a map of crustal thicknesses beneath Antarctica with values approaching 36 km. over the central southern parts of the peninsula. At the junction with the Filchner Ice Shelf and the Bellingshausen Sea the depth given was 32 km. These estimates compare with the values of around 36 km. given by Behrendt (1964) for the mainland and 31 km. (Behrendt, 1962) beneath the Berkner Bank on the Filchner Ice Shelf. Both Frolov (1965*a*) and Demenitskaya (1960) have extrapolated crustal thicknesses over the Antarctic Peninsula, the values of Demenitskaya being considerably greater than those postulated by other authors. Assuming crust and mantle densities of 2·84 and 3·27 g. cm.$^{-3}$, respectively, Worzel and Shurbet (1955) calculated a standard crustal thickness of 33 km. for the low-lying continental areas. This was used as a basis by Griffiths and others (1964) to calculate crustal thicknesses from Bouguer anomalies over Bransfield Strait and northern Graham Land. Thicknesses of 28 km. were obtained for Bransfield Strait, while the tentative value of 37 km. determined beneath the Antarctic Peninsula was based on the minimum of field data. With additional data, Davey (1971) computed crustal thicknesses beneath northern Graham Land and obtained values of up to 32 km.

C. The Larsen Ice Shelf and Continental Shelf

In a previous paper, Renner (1969) discussed the need for information regarding surface elevations and ice-thickness values for the Larsen Ice Shelf. Radio echo-sounding methods (Smith, 1972) have provided limiting depth values but there still remains the problem of the depth and lithology of the sea floor beneath the ice shelf. In the absence of seismic traverses across this area, it was decided to compare the available bathymetric data from the Larsen Ice Shelf with that recorded by seismic and gravity methods from other Antarctic ice shelves. If known depths to the continental shelf off the Larsen Ice Shelf were found to be typical of those elsewhere, then sea depths could be extrapolated with some degree of confidence. By so doing, a more accurate Bouguer correction could be applied and a more representative regional gravity gradient determined.

Investigations on the physical characteristics of ice shelves and the continental shelf around Antarctica have been carried out from ship traverses in the Southern Ocean and also by geophysical traverses completed on Antarctic ice shelves themselves. Geophysical methods employed include both seismic and gravity, the latter being used to interpolate sub-surface topography between the seismic stations where depths have been accurately determined. Reflection and refraction techniques have given the ice–water and water–rock interfaces as well as providing evidence of rock type beneath the ice shelves, e.g. morainic deposits. Where no reflections were obtained between ice and water, the ice thicknesses were determined from surface elevations on the assumption that hydrostatic equilibrium had been achieved for the floating ice shelf.

The ice shelves of Antarctica have been estimated by Thiel and Ostenso (1961) to cover an area of 1,400,000 km.^{2}, which is 10 per cent of the surface area of the continent. The largest of these, the Ross Ice Shelf, has a surface area of 540,000 km.^{2} and an average thickness of 450 m., though ice-shelf thicknesses may vary from about 200 m. at the ice front to over 1,300 m. at the junction with the mainland (Zumberge and Swithinbank, 1962). Thiel (1962) estimated an overall average thickness of 380 m. for all Antarctic ice shelves.

The continental shelf surrounding Antarctica has been found to be very different from that of the rest of the world. Woollard (1962*b*) has stated that soundings in the southern oceans have shown the shelf break to lie at a depth of about 400 m. compared with the world average value of 132 m. given by Shepard (1963). Thiel (1962) has given a mean depth for the Antarctic continental shelf margin of 720 m. below sea-level. This was obtained by averaging depths from the bathymetric charts through each 10° of latitude and taking the mean of the 36 values thus obtained. Lazarev and others (1965) gave an average depth of 700–800 m. beneath the Shackleton Ice Shelf, while Tsukernik and others (1963) estimated the bedrock at 400–500 m. below sea-level on the "West Ice Shelf". On the Maudheim Ice Shelf (Robin, 1958), seismic reflection methods gave depths to the sea bed averaging between 500 and 600 m. Grushinsky and Frolov (1967) have given a depth of 400–500 m. with a width extending outward in places for 500 km.; the same authors brought attention to a feature which appears particularly in the eastern Antarctic, that the entire peripheral zone is characterized by tectonic structures. These trend both parallel and at right-angles to the continent and can take the form of faults, depressions and areas of block relief. Lisitzin and Zhivago (1960), Tsukernik and others (1963), Frolov and Koryakin (1960) and Behrendt (1962) noted specific cases.

The Ross Ice Shelf is one of the most accessible to working parties and this has resulted in a fairly comprehensive study of its characteristics. Crary and others (1962) have given a detailed picture of the depths to the ocean floor beneath the ice shelf. They found that the western end of the Ross Ice Shelf overlies water to a depth of 1,400 m., though relatively deep water is to be found on all sides of the ice shelf where it is adjacent to high plateau areas. Elsewhere on the Ross Ice Shelf, depths to the ocean floor vary from 400 to 700 m. Covering a wider area, Zhivago (1962) gave a geomorphological map of the Southern Ocean which included the structures of the continental shelf off the east coast of the Antarctic Peninsula. He described the shelf in this vicinity as "hillocky shelf", and of a zone of recent dislocation. There is no evidence concerning sea-floor structures beneath the Larsen Ice Shelf though bathymetric data are available on the west coast. Between lat. 65° and 68° S., Nichols (1960) gave the continental shelf depth at 425–489 m., while in the Marguerite Bay area soundings suggest the presence of a north–south-trending depression extending northward from George VI Sound. The depression can be traced for over 300 km., reaching a depth of over 1,200 m. at lat. 68° 30′ S. where the average width between the 600 m. isobaths is about 32 km. A narrower trench reaching similar depths can be traced northward from Marguerite Bay running between the mainland and the offshore islands; north of lat. 66° 30′ S. its presence has not been identified possibly due to the lack of bathymetric data. Linton (1964) suggested this inner tectonic trough opens into Bransfield Strait, whereas its southern end may continue into George VI Sound. Insufficient bathymetric data are available at the southern end of Marguerite Bay to allow further comment on this. Linton regarded the group of offshore islands comprising the Palmer Archipelago, the Biscoe Islands and Adelaide Island as a mountain arc feature.

In discussing the results of geophysical and glaciological studies over the Filchner Ice Shelf, Behrendt (1962) noted that the eastern part of the ice shelf overlies a trough more than 1,500 m. in depth, whereas the western part overlies a shallow morainic bank. Behrendt suggested that both the trough and the bank could have been formed simultaneously by glacial action at a time when the ice was considerably thicker. Adjacent to the southern end of the Antarctic Peninsula depths to bedrock have been given as 500–600 m., the 600 m. isobath approximately paralleling the existing coastline. Between lat. 72° and 78° S. at the edge of the Filchner Ice Shelf, Behrendt showed depths to bedrock as determined by *Endurance* in 1915, *Deutschland* in 1912 and between 1956 and 1959 by the U.S. Navy ships *Wyandot*, *Staten Island* and *Edisto*. Depths range from 200 to 1,000 m. below sea-level; immediately off the ice-shelf edge they average 200–300 m., while 300 km. offshore they approach 400–500 m. and trend almost parallel to the coastal contours. A preliminary bathymetric chart (Dale, 1968) shows the 500 m. isobath coincident with the eastern limit of the continental shelf around Graham Land. At lat. 68° S. the shelf width is about 430 km.

Little knowledge is available concerning the unconsolidated sediments of the Antarctic continental shelf. Frolov and Koryakin (1960) commented on the character of Antarctic continental shelf regions, describing their surfaces as being at depths greater than 500 m., as being irregular and covered by a thin layer of detritus, the sediments increasing in thickness seawards. Beneath the Maudheim Ice Shelf (Robin, 1958), the moraine is of a reasonably uniform thickness with depths averaging around 65 m. not uncommon. Lisitzin (1962) explained the maximum deposition towards the continental slope as due to iceberg transport, while deposition near the coast is slight and the bottom topography more rugged. When present, the shelf sediments include angular stones, sands, silts and clayey muds. In discussing the glacial marine-sediment distribution around Antarctica, Hough (1956) stated that the unsorted glacial sediments extend as far north as the maximum limit of the pack ice, where they gradually merge into the next sedimentary zone. Poulter (1947) and Crary (1961), using seismic methods over areas on the Ross Ice Shelf, calculated sediment thicknesses. Crary also took cores which showed that the upper parts of the sediment consisted of coarse and fine glacial till probably deposited by ice. Both investigations were on the eastern side of the Ross Ice Shelf, in the "Little America" area. Poulter determined that near the ice front moraine thickness averaged around 110 m., the deposits being quite uniform except at places of maximum currents. In many instances, he claimed that the ice shelf is grounded on moraine, which may reach thicknesses up to 457 m. This grounding is sufficient to account for the very existence of the ice shelf. From seismic soundings, Crary published the following results regarding sediment thicknesses beneath the eastern Ross Ice Shelf:

i. In the vicinity of "Little America" station:

	Depth (m.)	*Velocity* (km. sec.$^{-1}$)
Ice	238	3·45
Water	409	1·44
Sediments*	1,325	2·4
Rock†	645	4·24
Rock		6·38

ii. 35 km. south-east of "Little America":

	Depth (m.)	*Velocity* (km. sec.$^{-1}$)
Ice	363	3·45
Water	293	1·44
Sediments	754	2·4
Rock		

Depth to bedrock 1,354 m.

* The sediments thin at a low rate of 1 m./200 m. to the south-east. Depth to bedrock was calculated as 1,933 m.
† Tentatively equated to the Beacon Supergroup.

Similar thicknesses of sediments were expected by Crary to cover the remainder of the Ross Sea–Ross Ice Shelf area.

From seismic investigations over the continental shelf north-west of the South Shetland Islands, Ashcroft (1972) estimated that unconsolidated to semi-consolidated sediments may reach thicknesses of 2 km. Consideration of all sediments increases this to 6 km. with derivation from the South Shetland Islands since the Carboniferous. The present extent of the shelf in this area is attributed to such prolonged sedimentation. In contrast, the shelf north-west of Trinity Peninsula, which probably originated as an erosional surface during a period of reduced sea-level, is not expected to reveal much unconsolidated sedimentary material. Beneath Bransfield Strait up to 2 km. of unconsolidated sediments (seismic velocity 1·8–3·3 km. sec.$^{-1}$) are estimated as having been derived from marine erosion and Tertiary–Recent volcanicity. For post-Jurassic sediments on the shelf south-west of Clarence Island, Watters (1971) also calculated a thickness of about 2 km.

The continental shelves bordering glaciated land masses are thus seen as amongst the most distinctive shelf types. They are generally wide with deep basins and troughs trending both parallel and transverse to the coastline, and the shelf depressions often connecting with inland fjords. The origin of the shelf appears to have been greatly influenced by the lowering of sea-level, estimated to be 110–165 m. during maximum glaciation. This would have resulted in large areas of grounded ice which would then have greatly contributed to the shaping of the shelf structures. Shepard (1963) preferred a glacial origin for the troughs in preference to a faulted one, though older fault lines may have influenced the margins of some of the trough directions. Recent movements due to ice removal may have contributed to some of the observed tectonics.

Glacial activity must have been of prime importance in the history of the Antarctic continental shelves. Voronov (1960) estimated on the basis of the present equilibrium profile that during maximum glaciation the edge of the continental shelf formed a natural boundary for the ice sheet. As a result, the 2,000 m. contour over the Antarctic Peninsula would have passed outside of Alexander Island, swinging across the mainland at about lat. 66° S. before following the present line of the east coast southwards. The 1,000 m. contour was located about 50 km. inside the edge of the continental shelf. Nichols (1960) produced evidence that in the vicinity of Stonington Island ice thicknesses of 610–915 m. existed in areas now deglaciated, and that the whole of the Antarctic Peninsula was covered with a more or less continuous ice cap with extensive coverage over Marguerite Bay. In the Trinity Peninsula area, at least 300 m. of ice have been removed since maximum glaciation (Koerner, 1964) and substantially more from Prince Gustav Channel. Adie (1964*a*) gave estimates that at glacial maximum the ice over Graham Land was 370 m. more than at present, bringing the level of the ice on the plateau to over 2,000 m., which is in accordance with that given by Voronov. Thiel (1962) argued that on a simple isostatic equilibrium principle 1 km. of ice will depress the underlying rock 256 m., thus assuming an average increase of 1,000 m. of ice during the Pleistocene maximum (Goodell, 1965) and a depth to the shelf of −720 m.; this would place the continental shelf limit at 1,000 m. below present sea-level. As previously discussed, opinion appears to favour the edge of the shelf to be at a depth of less than 720 m., therefore indicating a Pleistocene shelf depth of about 800 m. below present sea-level. Allowing a drop of 150 m. (Goodell, 1965) for a Pleistocene sea-level due to ice accumulation, this would give a continental shelf edge of −650 m. during that period. Considering that some ice thicknesses approach this value on the Larsen Ice Shelf (Renner, 1969), it therefore seems more than likely that during the Pleistocene the ice sheet was grounded on the continental shelf out to a considerable distance, if not to the edge of the shelf. On the Antarctic Peninsula this would have resulted in a more or less continuous ice sheet engulfing Alexander Island and the South Shetland Islands, while on the east coast the Larsen and Filchner Ice Shelves would have merged into one. Behrendt (1962) estimated that at present 400 m. of additional ice would be necessary to ground the Filchner Ice Shelf at its deepest, but much less would be required on the shallower parts. The trough and Berkner Bank were suggested as having been formed simultaneously by glacial action, when the extent of the ice was coincident with the 500 m. isobath in the Weddell Sea. Hollin (1962) estimated that during the Würm glaciation maximum the ice shelves of Antarctica could have extended on average for a further 90 km. That the ice sheet did have an effect in Graham Land and the Scotia arc was demonstrated by Adie (1964*b*), who compared the depths to the continental shelf over the whole region. Here a direct correlation between latitude and depth is evident, with South Georgia having a shelf depth of less than 130 m. compared with a 500 m. depth in Graham Land. The difference has been attributed to differential depression coupled with the delay in isostatic rebound due to the thickness and removal of the overlying ice.

Exceeded only in size by the Ross and Filchner Ice Shelves, the Larsen Ice Shelf, which is situated on the east coast of the Antarctic Peninsula, extends in an arcuate belt from the Seal Nunataks (lat. 60° 03′ S., long. 60°18′ W.) for about 850 km. southward to Cape Mackintosh (lat. 72°53′ S., long. 60°03′ W.). At its broadest, near lat. 67° 40′ S., it is 230 km. in width, and its surface area has been estimated (Thiel, 1962) as 92,000 km.^{2}; however, the author considers on the basis of information now available that an area of 75,000 km.^{2} is a more realistic value. North of the Seal Nunataks the ice shelf continues without any obvious change for 75 km. to Cape Longing (lat. 64° 33′ S., long. 58° 50′ W.), where it appears to reach its northernmost limit as a true ice shelf. On the northern side of Cape Longing, an apparently thinning ice shelf stretches for a further 30 km. before merging with the fast ice in the vicinity of Persson Island at the southern end of Prince Gustav Channel. The boundary between the thinning ice shelf and the fast ice is extremely difficult to determine because of the pressure ridges in this area; their origin, at least in part, must be attributed to pressure from the Sjögren Glacier ice tongue flowing from the west. One indication of the transition from the ice shelf to the sea ice was the increased oscillation observed on the gravimeter, due possibly to ocean waves entering the area and to disturbances by local winds (LeSchack and Haubrich, 1964). Reece (1950) could see no great difference between the ice shelf in the Persson Island–Cape Longing area and that south of Cape Longing.

Until recently, surface-elevation values on the ice shelf have been restricted to visual estimates together with more objective determinations (Mason, 1950; Fleet, 1965; Kennett, 1965*c*; D.O.S. 1:200,000 sheet W6864). A more comprehensive determination of surface elevations (Renner, 1969) was undertaken using radio echo-sounding measurements. These initially gave readings of the total ice thickness, ranging from 130 m. at the ice front to 650 m. at the heads of the inlets. The thicknesses provided some guidance as to the minimum depths to the sea floor and, by assuming hydrostatic equilibrium, enabled the derivation of surface elevations. The radio echo-sounding traverses were located on that part of the Larsen Ice Shelf between Cape Disappointment (lat. 65° 33′ S., long. 61° 45′ W.) and Mobiloil Inlet (lat. 68° 35′ S., long. 64° 45′ W.), corresponding to the area where most of the geophysical traverses have been made. North of Cape Disappointment, the only ice-shelf elevations available were made by the Argentines near their station "Teniente Matienzo" on Larsen Nunatak, one of the Seal Nunataks. R. N. M. Panzarini (personal communication) has given the height of the barometer zero at "Teniente Matienzo" as 32 ± 2 m. a.s.l., and the height of the ice shelf at its edge near Christensen Nunatak as 14 m., varying up to 17 m. in the section between Christensen Nunatak and Robertson Island.

The available bathymetry beneath the Larsen Ice Shelf area (Fig. 3) gives some indication of the depth and extent of the associated continental shelf; it also illustrates certain discrepancies within the data. The first is the placement by Shepard (1963) of the 100 fathom [183 m.] line. The source of the data used for this compilation is unknown, but from the radio echo-sounding programme ice-thickness values of over 550 m. below sea-level have been obtained for the floating ice shelf in the area between Churchill Peninsula and Three Slice Nunatak. The soundings (personal communication from R. N. M. Panzarini) completed by the ice-breaker *General San Martín* present the second misleading phenomenon. "While moored to the ice shelf between Christensen Nunatak and Robertson Island the *General San Martín* sounded 150 fathoms [275 m.]; at 0·5 nautical miles [0·9 km.] off Christensen Nunatak the depth is less than 25 fathoms [46 m.]; and at a station located 0·3 nautical miles [0·55 km.] off Robertson Island, bearing E 53° S. true from the 'San Roque' shelter, a depth of 50 fathoms [92 m.] was sounded."

At first these soundings appear to be far too shallow for depths to the continental shelf but a closer examination of them reveals a plausible explanation. The three depth values were all obtained close to rock outcrops whose surface gradients could continue gradually beneath sea-level. In addition, the Seal Nunataks are formed of volcanic rocks belonging to the James Ross Island Volcanic Group of Miocene or younger age, and thus sea-floor irregularities could partly be due either to the erupted material or to denudation products from the nunataks. Lisitzin (1962) has described the sediments which cover the bases of sub-Antarctic islands as having been formed from the erosion of the late Tertiary–early Quaternary volcanic rocks of the islands.

Soundings by R.R.S. *John Biscoe* and R.R.S. *Shackleton* have been made over the ice-free parts of Duse Bay and Prince Gustav Channel. Recorded depths were variable but with maximum values of 970 and 890 m., respectively. However, such values are not considered as being typical of Antarctic shelf areas, and their depths are therefore not regarded as being representative of the east coast of the Antarctic Peninsula as a whole.

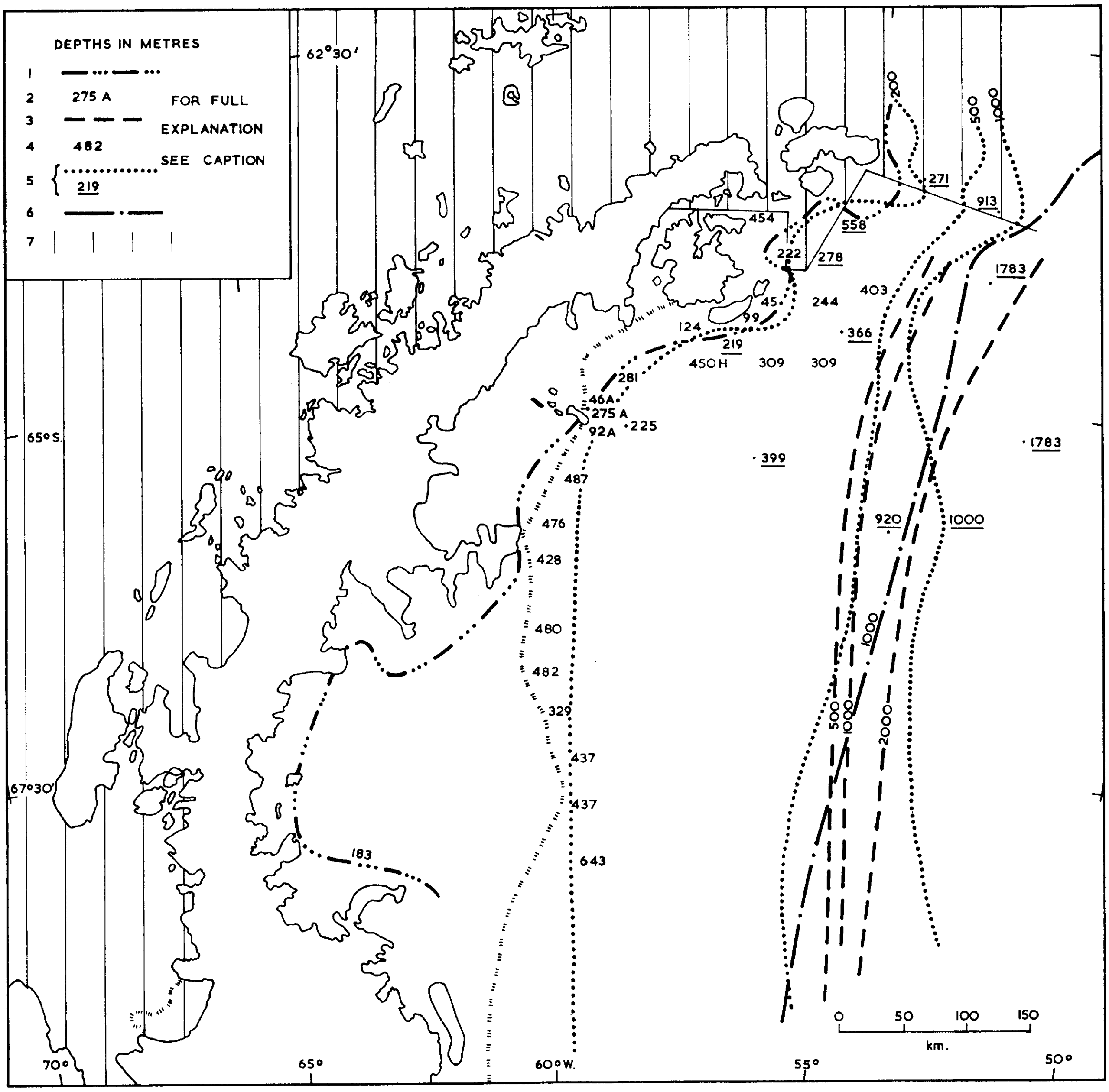

FIGURE 3

Available bathymetric data over the continental shelf off the east coast of Graham Land.

1.	——· · ·——· · ·	Shepard (1963).
2.	275A	R. N. M. Panzarini (personal communication).
3.	—— —— ——	Dale (1968).
4.	482	Admiralty chart 3200.
5.	 225	Carte Générale Bathymétrique des Océans, Feuille B'I, 3rd édition, publiée par le Bureau Hydrographique International, 1952.
6.	——·——·——·	D.O.S. 1:2,000,000 sheet 813 (series 3203).
7.	| | | |	Information from Admiralty charts 3205, 3570 and 3571.

From the evidence available, it is suggested that the Larsen Ice Shelf and the continental shelf in this area share certain features common to other Antarctic regions. Thus, provided that the thickness values assigned to the Larsen Ice Shelf are within those ascribed to others, then the accuracy of data reduction can be considered enhanced. Outside of an approximate limit to the inlets, where total ice thicknesses of more than 650 m. have been recorded, the ice-thickness values are all less than 500 m., a figure shown to be not uncommonly quoted for depths to bedrock beneath Antarctic ice shelves. It was therefore decided in the first instance to use 500 m. as the standard depth to bedrock beneath the Larsen Ice Shelf except, of course, in those areas where ice-thickness values showed otherwise when the thickness value itself was used as a limiting depth. Areas where incorrect depths were assumed soon became apparent through their residual gravity anomalies, and they are discussed later (p. 70–73). Whether or not the bedrock beneath the Larsen Ice Shelf is free from significant structures is yet unknown but the gradual thinning of the ice eastward and the almost complete lack of severe variations in the ice-thickness values imply that the bottom topography is not a rugged one. The thickness, composition or density of any overlying shelf sediments cannot be discussed as neither measurements nor samples have been taken, although the magnetic survey does contribute (p. 57) slightly towards an understanding of the nature of the bottom topography. There are anomalous ice-thickness areas, as exemplified to the east of Three Slice Nunatak, but these are regarded as only local irregularities.

D. Geology

The Antarctic continent may be sub-divided into two distinct geological provinces by a line between the Filchner Ice Shelf and the Ross Ice Shelf. East Antarctica, or the Gondwana Province, has a stratigraphical and tectonic history reminiscent of a stable continental platform but, in contrast, the Andean Province of west Antarctica has a relatively young geological history exhibiting marked similarities to that of the South American Andes. The Antarctic Peninsula forms a considerable part of the Andean Province and there have been numerous publications concerning its geological background. Adie (1964*a*) has reviewed the geology of Graham Land and the Scotia arc, and in this section it is intended to give only a brief outline of the geology of the mainland. Ashley (1962) and Allen (1966*a*) have previously discussed the geology of north-eastern Trinity Peninsula and its relation to their geophysical interpretations. Similarly, reference is made throughout this report to local geology, and where such information has been used the relevant source is given.

A stratigraphical succession for Graham Land is given in Table I (Adie, 1964*a*), which serves to illustrate the major geological divisions. The Basement Complex of (?) Precambrian–Cambrian age, which is restricted to the southern part of Graham Land, south of lat. 65° 45′ S., consists of metamorphic rocks of igneous and sedimentary origin. Subsequent earth movements and igneous activity have resulted in a succession of highly foliated schists and gneisses exhibiting a north-east to south-west regional strike. During the early Palaeozoic there occurred the first identifiable volcanic episode with representatives again limited to, and to the south of, the Marguerite Bay area. Unconformably overlying these earlier rocks are the Trinity Peninsula Series whose type area and most extensive development so far recognized is at the northern tip of Graham Land. These sediments, at least 12,000 m. of which have been recorded at one locality in Trinity Peninsula (Aitkenhead, 1965), were deposited in a major geosynclinal trough during the (?) Carboniferous and they are almost devoid of fossil evidence. Since their deposition, they have undergone considerable folding along a south-west to north-east axis, while intense thermal metamorphism occurs in those areas intruded by the younger Andean Intrusive Suite. Lithologically similar greywacke-facies sediments have been identified in South America, South Georgia and the South Orkney Islands.

The Middle Jurassic provided an abundance of fossil fauna and flora in the sediments which lie unconformably on the Trinity Peninsula Series, the latter having been uplifted during the early Jurassic. Again confined to the north of the Antarctic Peninsula, these beds are exposed at Hope Bay as the lacustrine Mount Flora plant beds. Bordering Prince Gustav Channel at Church Point, a similar fossil record has been discovered but, despite the sequence again being of a shale and conglomerate facies, it is believed that marine transgressions influenced the succession. The close of the Middle Jurassic heralded the most extensive of the volcanic episodes that occurred in Graham Land. Along its entire length, volcanic rocks belonging to a calc-alkaline andesite-rhyolite suite are exposed with thicknesses of up to 3,000 m. There appears to be a systematic distribution within this suite, since the more basic representatives, such as the andesitic lavas, are far more predominant along the west coast of the peninsula and the more acid rhyolitic and dacitic lavas crop out along the east coast. The volcanic rocks of the Hope Bay area show an upward gradation from the Middle Jurassic plant beds but elsewhere they covered the Basement Complex and

TABLE I
STRATIGRAPHY OF THE MAINLAND OF GRAHAM LAND (ADIE, 1964*a*)

Recent	Sea-level changes
Pleistocene	(Glaciation)
Pliocene	*Pecten* Conglomerate
M. Miocene	(Osterrieth Range volcanic rocks and associated dykes) James Ross Island Volcanic Group
L. Miocene	Seymour Island Series
Late Cretaceous to early Tertiary	Andean Intrusive Suite
U. Cretaceous (L.–M. Campanian)	Snow Hill Island Series Cape Longing Series
U. Jurassic	Andesite-rhyolite volcanic group
M. Jurassic	Mount Flora plant beds Church Point plant beds Basal conglomerate
(?) Carboniferous	Trinity Peninsula Series
(?) Early Palaeozoic	Intrusive suite Volcanic rocks
Precambrian	Basement Complex

Trinity Peninsula Series sediments which by that time possessed a topography at a fairly mature stage. Intruded into the volcanic rocks and almost contemporaneous with them are porphyries and white granites.

Despite its widespread exposure on Alexander Island, the Lower Cretaceous has not yet been positively identified on the mainland of Graham Land, whereas the Upper Cretaceous is represented by over 5,000 m. of fossiliferous conglomerates, sandstones and shales. These rocks are gently folded about a north–south axis and they are mainly confined to north-eastern Graham Land, where they are known as the Snow Hill Island Series and the Cape Longing Series.

The very existence and present relief of the Antarctic Peninsula can largely be attributed to the igneous and tectonic activity resulting from the emplacement of the Andean Intrusive Suite. Extensively distributed along the length of the Antarctic Peninsula, these rocks occur as a series of batholithic bodies which intrude (often causing intense thermal metamorphism) all pre-Upper Cretaceous rocks. This intrusive suite has a wide range of chemical compositions as is demonstrated by the presence of gabbros, diorites and granites. Outside the Graham Land region, they have been recognized in the South Shetland Islands, appear as an equivalent suite in South Georgia and are exposed in the Patagonian Cordillera of South America.

Relics of any Tertiary sedimentation are not abundant in Graham Land, being represented by only 150 m. of the Seymour Island Series. As yet, these deposits have not been accurately dated within the stratigraphical column, as their exact relationship to the Upper Cretaceous Snow Hill Island Series is still questionable. The most significant geological feature of the Tertiary, however, was a volcanic event which reached its climax during the Middle Miocene. Olivine-basalt lavas, agglomerates and tuffs form part of the 1,500+ m. thick James Ross Island Volcanic Group, which was extruded at a stage when tectonic activity was at its peak. Two related, yet distinctive volcanic areas resulted, though recent K-Ar dating by Rex (1971) has suggested that there is a possible time interval between them. The James Ross Island group of islands appears to have been the result of volcanic plug eruptions, whereas the members of the Seal Nunataks group apparently owe their existence to eruptions along a west-north-west fissure. Also attributed to this period are basaltic dyke swarms and a small lava infilling in the Osterrieth Range of Anvers Island. It is thought that by the onset of the Tertiary Prince Gustav Channel must have already been partially in existence though tectonic movement must have been renewed during this latter period of volcanism. The Tertiary is also well represented by a volcanic succession in the South Shetland Islands, where lavas and pyroclastic deposits are interbedded with fossiliferous horizons.

The geological events since the undetermined close of the Tertiary volcanic episode have been confined to the deposition of an apparently thin *Pecten* conglomerate off north-eastern Graham Land and to the Pleistocene–Recent volcanic episode of Deception Island. However, this volcanism is regarded as a continuation of the Tertiary activity in the South Shetland Islands.

Passing reference has only been made in the above summary to the islands which are the visible physical evidence of the feature known as the Scotia Ridge. There are geological similarities between Graham Land, the South Shetland Islands and South Georgia, which indicate the important tectonic movements that have taken place. Barker and Griffiths (1972) have discussed more fully the geological and geophysical implications of the Scotia arc and how a tectonic history is gradually emerging from their conclusions. Neither is it proposed to comment on the structural or stratigraphical significance of Alexander Island which lies off the south-west coast of the Antarctic Peninsula. Both gravity and magnetic traverses have recently been completed over this area, and it is hoped that in future publications it will be possible to give a more complete account.

E. Interpretation Methods

1. *Magnetic interpretation*

The interpretation of the magnetic anomalies has been more straight-forward than those involving gravity methods. This is due to their more direct reflection of the source rocks in there being much less error introduced in the data reduction.

Initially, from the inspection of the residual-anomaly maps the individual anomalies could be broadly classified into groups. An immediate outcome of this was to reveal the different magnetic character exhibited by the rocks of the flat non-magnetic Larsen Ice Shelf and those of the more magnetically active mainland of the Antarctic Peninsula, both of which constrast sharply with the belt of high magnetic activity off the west coast. Over Graham Land, and superimposed upon the broad low-amplitude anomalies which reflect variations in the magnetic basement, are anomalies of greater amplitude and smaller wave-length. These areas are often coincident with the outcrops of the Andean Intrusive Suite. Where the Trinity Peninsula Series sediments are present, the anomalies show a corresponding decrease in amplitude. More isolated magnetic anomalies with larger amplitudes are also found which in places could be associated quite definitely with the outcropping basic dyke and intrusive rocks of the James Ross Island Volcanic Group.

After the preliminary examination of the magnetic maps, the more quantitative aspect of individual anomalies required investigation. Wherever possible, representative profiles were examined but this was somewhat dependent on the field coverage because in a few areas, e.g. the Seal Nunataks, the anomalies were traversed by single profiles only. Provided a representative profile had been obtained, then the interpretation methods applied to any particular anomaly depended on the quality of data available and the character of the anomaly itself. Certain basic conditions were assumed in these methods. The body was taken to be two-dimensional with a horizontal upper surface, its sides parallel and often extending to great depths, and with a uniform magnetization throughout. The empirical method of Peters (1949) gives estimated maximum depths to the top surface. Smellie (1956) devised a series of type curves based upon selected parameters for a limited number of theoretical bodies ranging through a pole and dipole source to a line of poles and dipoles. By this method the observed profiles are equated to idealized geological forms from which certain factors can be derived and these are then readily convertible to depth values. Complete curve-matching techniques have been developed by Gay (1963), who published a set of standard curves for all values of dip, strike and inclination over a thin infinite dyke. This idea was developed further by Grant and Martin (1966), allowing a larger number of curve-matching alternatives. The susceptibility values used in the interpretation have been selected from different sources. In some instances, particularly over Trinity Peninsula, measured values were available (Ashley, 1962; Allen, 1966*a*), and where such type exposures were found in other areas the corresponding measured susceptibility was used as a first approximation. During the geological programme many of the specimens collected were analysed modally and reference was made to the relationship of magnetite content to susceptibility as given by Mooney and Bleifuss (1953). The susceptibility can also be determined by reference to Gay (1963) and Grant and Martin (1966), whereas that of Vacquier and others (1951) is more applicable to bodies of limited length. In discussing the susceptibilities calculated from the magnetic profiles, no individual account is taken of any remanent magnetization which may be present. The susceptibility is therefore an apparent susceptibility based on the total magnetization of the rock body as a whole. In cases where there is much variation in rock composition within a body, this may be at least as accurate as laboratory susceptibility measurements on questionable "representative" samples.

All of the significant magnetic anomalies were subjected to at least one of the profile methods of interpretation. Where sufficient field data had been obtained to warrant a more rigorous investigation, then the information derived from the profiles was used as a basis for computer analysis. Computer programmes were compiled to satisfy both two- and three-dimensional bodies. It was partly because of these more advanced techniques that some of the vertical-field anomalies of Ashley (1962) and Allen (1966*a*) were re-interpreted, these having been originally interpreted by graticule methods.

The following are brief descriptions of the computer programmes used in the interpretation.

a. *Two-dimensional interpretation.* The programme is based upon the work of Talwani and Heirtzler (Heirtzler and others, 1962) for the computation of magnetic anomalies due to two-dimensional bodies of polygonal cross-section. The data input included body coordinates and a facility for a variable surface topography which was necessary when the body was known to approach within a few metres of the ground level. Values of magnetic inclination were required as were the strike of the body, the total ambient magnetic field and the susceptibility. Vertical, horizontal and total magnetic fields were given in the output together with a graphical print-out which enabled a more rapid comparison with the observed profile. Varying body shapes and susceptibilities were tried until the observed and calculated fields showed an acceptable correlation.

b. *Three-dimensional programme.* This programme, which originated from a method described by Bott (1963), was primarily used in the interpretation of the anomaly associated with Bull Nunatak, one of the Seal Nunataks. Data information required included reference to the magnetic inclination, the intensity of the Earth's total magnetic field and the susceptibility contrast, while the method also has the facility to accommodate any remanent magnetization. The computation depends on the representation of the body on a two-dimensional platform by a series of plane polygonal surfaces. The process involves the transformation from a volume integral to a surface integral, thereby permitting the components of magnetic intensity to be evaluated at any prescribed observation point. A total-field anomaly was calculated for each station position and body coordinates were altered where necessary in order to derive the most satisfactory fit to the field data.

2. *Gravity interpretation*

It became increasingly apparent that the interpretation of the gravity data would have to be severely limited because of the uncertainty of the ice thicknesses and sea depths. However, the radio echo-sounding techniques have greatly improved interpretation by providing greater control in data reduction. Although no quantitative interpretation has been attempted for the west–east crustal sections due to insufficient control, reference is made to the relative crustal variations observed along the length of the Antarctic Peninsula and to the proximity of two traverse sections around lat. 68° S.

Certain qualitative observations can be made from the data, the most obvious of which concern the areas of gravity "lows" coincident with many of the inlets along the east coast of Graham Land. These can be readily interpreted as areas of glacial overdeepening. Almost equally apparent are those areas, often adjoining promontories and peninsulas, where increased gravity indicates areas of a rise in bedrock with ice thinning and possibly subsequent grounding. Where the anomalies appear to have a more direct association with geological causes, quantitative estimates of depths and extents have been incorporated. The Churchill Peninsula anomaly is an example of where certain body parameters were initially estimated using the methods described by Nettleton (1942) and Skeels (1963). In general, however, the majority of the quantitative work has been concerned with depths to bedrock whether beneath the ice of the mainland, the Larsen Ice Shelf or the sea ice of Duse Bay. The computer programmes used have been varied to suit the individual needs, and it is hoped that when ice-thickness information becomes available greater importance may be placed on the geological aspects.

a. *Two-dimensional interpretation*. The programme for the model-fitting solution was based on that of Talwani and others (1959). In this method a theoretical anomaly is calculated for a specified two-dimensional polygonal body in a direction perpendicular to the strike. The body is defined by a series of coordinates and therefore the accuracy of the calculated anomaly is governed by the number of representative sides. For the present work this programme was essentially used to calculate the relative variations in bedrock topography on the sea floor beneath the Larsen Ice Shelf. Removal of a regional gradient did present difficulties as the data are not yet adequate to enable a thorough analysis of the regional pattern. In many cases, the variation in bedrock beneath the Larsen Ice Shelf can be more simply estimated by using the two-dimensional Bouguer slab formula. This is particularly applicable over the areas away from the mainland, where, provided thick sedimentary deposits are absent, a change in free-air anomaly of 1 mgal would theoretically represent a change in bedrock elevation of 14·6 m. The presence of unconsolidated sediments could increase this to values approaching 25 m. However, in all of the gravity interpretation it has been assumed that thick sedimentary deposits are absent. Therefore, all depths to the sea floor are minimum depths provided the two-dimensional conditions are fulfilled.

A reverse procedure was adopted for the iterative solution compared with that of the model-fitting approach. A regionally corrected gravity profile was used and a two-dimensional body was obtained from it by a method of successive approximation. It was particularly applicable to some circumstances found over the Larsen Ice Shelf, Prince Gustav Channel and some of the inlets where the anomaly was attributed to a simple density contrast between sea-water and bedrock. Originally applied by Bott (1960) for use with sedimentary basins, the method first approximates the depth to bedrock beneath selected observation points on the assumption that the body can be represented by a horizontal infinite slab. For each observation point, a similar procedure is adopted and then the total contribution for all of the blocks is obtained by summation. The resulting anomaly is compared with the observed profile, any differences being obtained by subtraction to give a residual value. A gradual reduction of the residual value follows by the successive adjustment of the individual blocks, this process continuing for a pre-determined number of iterations or until the desired accuracy has been reached. The accuracy is determined in the first instance by the number of observation points selected. Each point represents a given strip width of the anomaly and hence an equivalent width in the calculated two-dimensional block. The greater the number of strips, the more realistic will be the bedrock profile.

b. *Three-dimensional interpretation*. Talwani and Ewing (1960) gave an expression for the gravity anomaly caused by a horizontal lamina. Assuming that an irregular three-dimensional polygonal body can be accurately represented by a system of thin horizontal laminae, the gravity anomaly for each can be calculated at any external point. By interpolation and integration, the anomaly caused by the entire body can be evaluated. In practice, the body under investigation is divided into a series of layers by depth contours. These are then approximated by polygonal faces, the number of which governs to some extent the accuracy of the method. The coordinates of the faces and the density contrast form part of the input to a computer programme, by means of which the gravity anomaly over the whole body is computed.

In northern Trinity Peninsula, part of the gravity survey involved a systematic coverage of Duse Bay. This feature has been partly depth sounded but the presence of semi-permanent sea ice over most of the area regularly prevents access by ship. It was therefore decided to combine the known bathymetry with the gravity data in attempting a three-dimensional iterative interpretation of the feature. This was again on the understanding that the bathymetry alone was responsible for the anomaly. The provisional interpretation showed that the sea bed lay at considerably greater depths than the extrapolation of the known bathymetry suggested. Therefore, to simplify the problems of repeated body-fitting, a three-dimensional iterative solution was incorporated.

The basic ideas of this method are similar to those described by Bott (1960) and used in the two-dimensional interpretation mentioned above. Namely, that from the gravity anomaly input an approximate model is derived whose calculated gravity effect is then compared with the observed data. The residual anomalies are then successively modified by an iterative process until the desired accuracy is achieved. Cordell and Henderson (1968) published an account of the method on which the present programme was based. A rectangular grid system was superimposed on the gravity anomaly contour map and the values were interpolated at the grid intersections. The coordinates of the grid system, together with a uniform density contrast, formed part of the data input. During the calculation it was assumed that each grid square represents the upper surface of a uniform vertical prism and that the body as a whole is composed of a large number of these. In the Duse Bay area, the upper surface of the anomaly-producing body was sea-level and this was therefore chosen as the necessary horizontal reference plane.

II. MAGNETIC SURVEYS

IN 1959, the British Antarctic Survey effected a plan for a magnetic survey of Graham Land when Askania vertical force variometers were acquired and measurements were commenced in the Hope Bay area. By 1962, a vertical magnetic field survey had been completed west to Cape Roquemaurel and south to Pitt Point (Ashley, 1962; Allen, 1966*a*), with a small vertical-field survey also completed at the Argentine Islands (Agger and others, 1963). The following season saw an extension of the vertical-field survey, N. Y. Downham completing traverses southward from Pitt Point to Cape Longing (June–September 1963) and to the west from Russell West Glacier to Charcot Bay (April 1963). In October 1963, R. C. Robson and R. M. Wilkinson carried out a local survey over Evensen Nunatak, one of the Seal Nunataks which protrudes through the Larsen Ice Shelf. Fig. 4 illustrates the areas covered by vertical magnetic field surveys in Graham Land.

In 1962, two Elsec proton magnetometers became available for work on the Antarctic Peninsula, thus giving a means of recording absolute values of the total magnetic field which could be directly compared with similar measurements being made in the Scotia Sea and Bransfield Strait (Griffiths and others, 1964). Since 1962–63, the vertical-component survey has been gradually superseded by measurements of the Earth's total magnetic field.

A proton magnetometer was used for a regional traverse of north-east Trinity Peninsula and also for a local survey of Hope Bay in 1962 (Cox, 1964*a*). The regional survey was extended the following year by J. Mansfield. This extended south to Cape Disappointment, the majority of the readings being confined to that area of the Larsen Ice Shelf between Cape Disappointment and Cape Longing. During the 1963 season similar work was being carried out in a total-field survey based on Stonington Island. A local survey (Kennett, 1966*b*) requiring sea-ice travel was accomplished on the west coast but longer regional traverses (Kennett, 1966*a*, *c*) extended to the east coast then northward to Cape Disappointment where a link was made with Mansfield's survey. At the close of the 1963–64 summer season the station at Hope Bay was vacated, leaving the field magnetic work to be based on traverses from Stonington Island. During the 1964–65 season, the author with the assistance of E. Thornton continued the total-field survey on the east coast, in addition to recording diurnal variation measurements at Stonington Island. Traverses extended from Mobiloil Inlet to the Seal Nunataks and thus overlapped both of the earlier surveys by Mansfield and Kennett. Until 1966 most of the total-field magnetic work had been confined to the east coast of Graham Land, though it was supplemented by smaller local traverses on the west coast in the vicinity of Stonington Island. During the 1966–67 season, J. Ross carried out further west coast traverses both northward to Pourquoi Pas Island and southward to the Wordie Ice Shelf, as well as making a closed plateau traverse

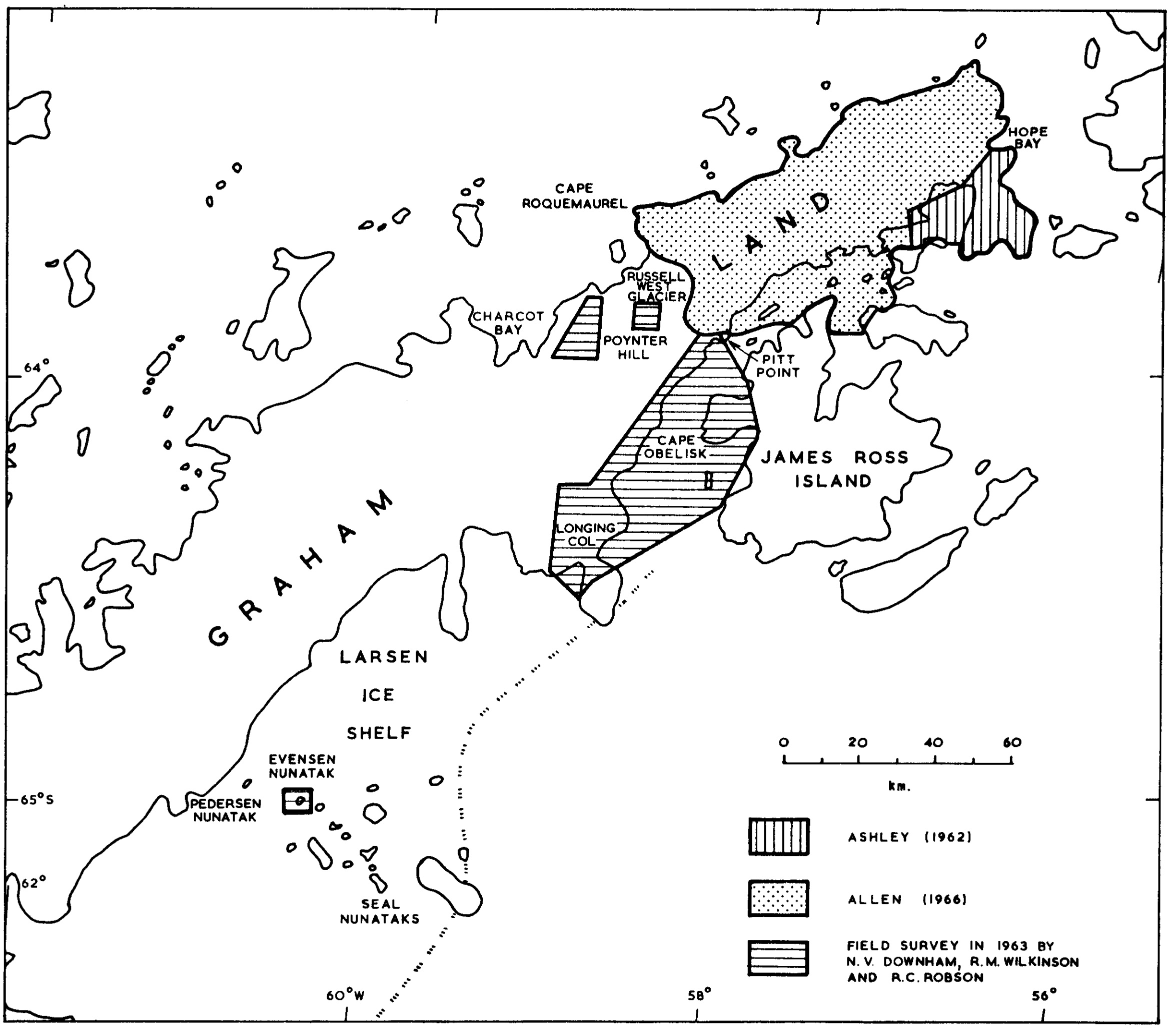

FIGURE 4
Areas covered by the vertical-field magnetic surveys.

from Stonington Island to lat. 70° S. Fig. 5 summarizes the total magnetic field traverses carried out by the British Antarctic Survey until 1967.

Consideration is given in this report to the total and vertical magnetic field surveys carried out since 1963 from Hope Bay, together with the total magnetic field surveys completed since 1964 from Stonington Island. During these surveys continuous recording of the vertical and horizontal components of the Earth's magnetic field was carried out at the Argentine Islands geophysical observatory, and at Hope Bay prior to its closure in 1964.

The aims of the total and vertical magnetic field surveys were as follows:

i. To extend the limits of the previous vertical magnetic surveys and to produce a total-field magnetic map which would form the basis for further magnetic studies in Graham Land.
ii. To determine the pattern of the regional magnetic gradient and compare it with the gradient determined in the Scotia Sea.
iii. To determine as far as possible the geological significance of the residual magnetic field, especially of areas where the rocks are largely hidden by sea or ice.

Any geological interpretation based solely on magnetic anomalies can only be ambiguous. However, the degree of ambiguity of interpretation may be decreased by considering all the available geological evidence and the magnetic properties of the rocks in the area of investigation. Ashley (1962) and Allen (1966*a*) have shown that the magnetic survey method is capable of providing much useful information in the Trinity Peninsula area when used in conjunction with the available geological information and rock-magnetism measurements. In particular, it can often delineate boundaries between rocks of the Andean Intrusive Suite and the relatively non-magnetic Trinity Peninsula Series, and reveal structural trends not obvious from geological mapping alone. It may also distinguish highly magnetic basaltic lavas and dykes of the James Ross Island Volcanic Group from the weakly magnetized tuffs and tuffaceous agglomerates of the same group, or from the adjacent Cretaceous sediments.

A. INSTRUMENTATION

Three instruments have been used in the course of the magnetic survey of Graham Land since 1963:

i. Askania Gfz torsion magnetometer—used for the vertical magnetic field survey.
ii. Askania Gf6 magnetometer—essentially used as a diurnal variation recorder at Hope Bay though previously (Allen, 1966*a*) it had supplemented the Gfz torsion magnetometer in field survey.
iii. Elsec proton magnetometer—two of these were in operation for the total magnetic field surveys, one based at Hope Bay and the other at Stonington Island.

1. *Askania Gfz torsion magnetometer*

Survey and instrumentation details of the magnetometer have been given previously by Ashley (1962) and Allen (1966*a*). At the close of the survey period in 1962, it was returned to the manufacturers for an overhaul before being used in the Antarctic during the following season. Calibration of the instrument was carried out before each period spent in the field, individual results differing by up to $\pm$ 6 gamma/scale degree from the manufacturer's calibration of 226·5 gamma/scale degree. The mean of the calibration factors, however, lay within the readability of the magnetometer and gave sensitivities to within $\pm$1 gamma/scale degree of the manufacturer's value.

2. *Askania Gf6 magnetometer*

This instrument, together with a Hartmann and Braun dotted line recorder, continued to be used for recording the diurnal variation of the Earth's vertical magnetic field at Hope Bay. Apart from a short period (Allen, 1966*a*) when it was used as a field instrument, the Gf6 magnetometer had been housed for several years in a non-magnetic hut at Hope Bay. Ashley (1962) described its installation and operation, and discussed the accuracy of the diurnal variations recorded. Provided the instrument was functioning correctly, a most important consideration in the accuracy of the diurnal records was the effect of temperature variation within the magnetometer hut. Unlike the Gfz magnetometer, which showed errors of less than 1 gamma over the temperature extremes encountered, the Gf6 magnetometer had a much greater temperature sensitivity with a temperature coefficient of up to 3 gamma/°F [5·4 gamma/°C]. Despite efforts to maintain a constant temperature, there were times when the methods employed proved most unsatisfactory and resulted in temperature variations of up to 20° F [11° C]. Mansfield (personal communication) believed that because of this the diurnal records from Hope Bay are not totally reliable.

3. *Elsec proton precession magnetometer*

These precession magnetometers have been used for the total magnetic field survey. They are absolute instruments which, in optimum conditions, are capable of measuring the total magnetic field intensity to an accuracy better than 1 part in 50,000, or $\pm$1 gamma in median latitudes. The principle of this type of magnetometer has been described by Waters and Phillips (1956).

The main advantage of the proton magnetometer over other types is that it requires minimal time to operate, gives an absolute reading and is drift-free. As the instrument measures the total magnetic field intensity, the data may be readily compared with other total-field measurements made in the Scotia Sea (Griffiths and others, 1964; Kroenke and Woollard, 1968).

One disadvantage of the proton magnetometer for Antarctic use is that a supply of electrical power is needed. The instrument operates from a 12V, 8A/hr. accumulator bank. The re-charging of batteries may present a problem if long periods are spent in the field with only dog-sledge transport. If tractors are used,

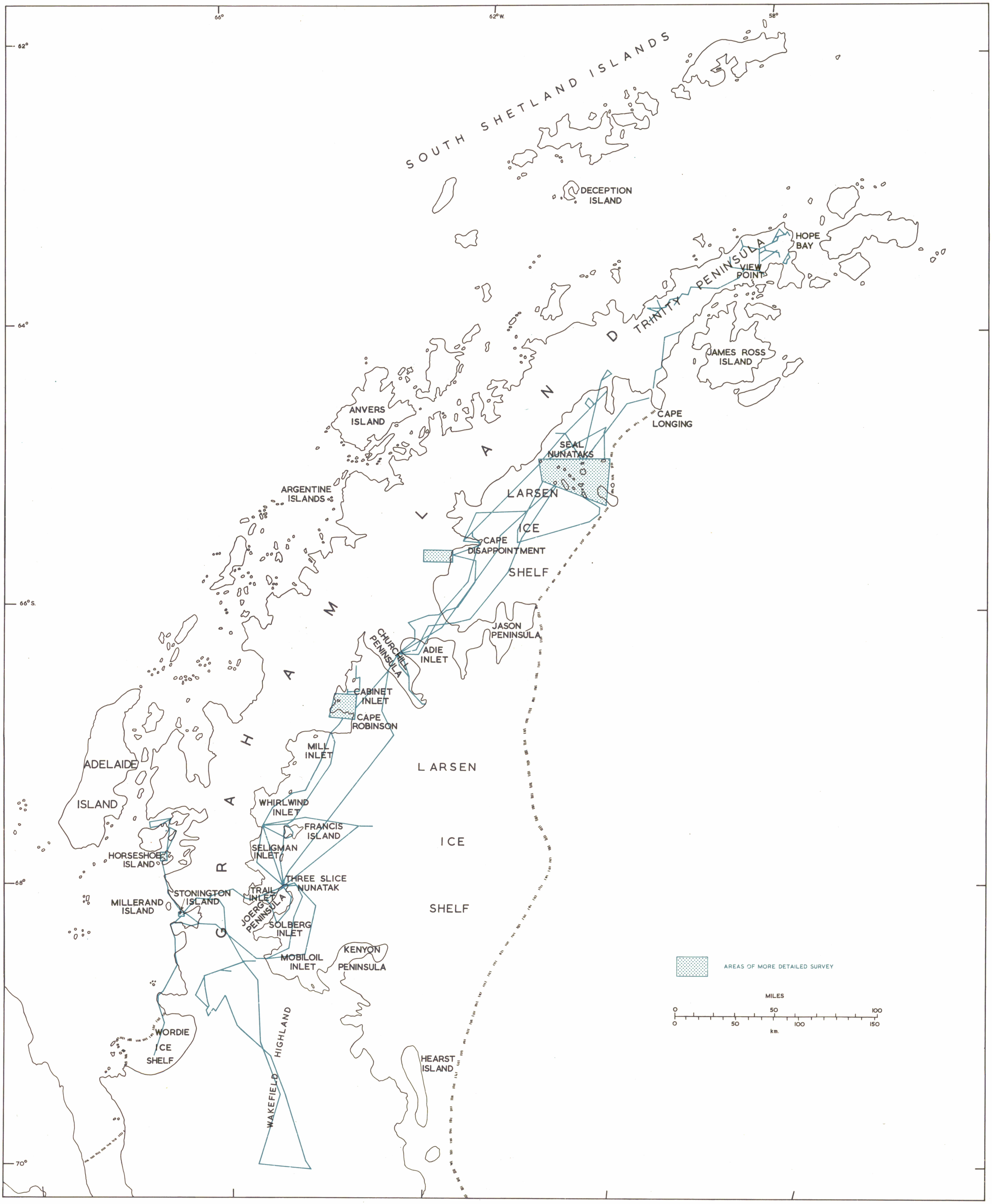

FIGURE 5
Total-field magnetic traverses in Graham Land.

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batteries can be re-charged from starter batteries or a portable charging generator can be carried; on one occasion it was possible to have re-charged batteries flown in by aircraft. It was found that one fully charged battery pack was sufficient for a month's normal work or about 150 stations provided the instrument was switched off after each reading. Spare battery packs were always carried but they lost charge gradually while not in use. Fortunately, it was possible to call at the Argentine station "Teniente Matienzo" at Larsen Nunatak on several occasions during the surveys and the batteries were re-charged there.

The magnetometers were carried on dog sledges for nearly 4,800 km. in the course of the survey. The failures that occurred due to rough travel were minimal and mainly attributable to the Venner plug-in units working loose in their sockets and causing circuit breaks. Where necessary, pieces of plastic foam were used to wedge the units in place. A number of circuit failures were caused by broken wires or joints but no components within the circuitry failed. One very persistent source of trouble was the failure of the external cable joining console and detector. At normal indoor temperatures the cable was flexible but at temperatures below 0° C the plastic insulation became brittle and cracked, leaving the conductors to take all the stress. The outcome was that on handling the conductors would break or short-circuit.

During part of the survey, in an attempt to reduce the frequency of cable failure and avoid the routine of uncoiling and coiling the cable for each station, the detector was towed on a small sledge 9 m. behind the dog sledge. A terylene line was used for the tow and the cable was loosely taped to the line so that no towing stress fell on the cable. The detector was built into an octagonal wooden frame which fitted into a square box on the sledge in such a way that the detector could be rotated horizontally in 45° steps to gain a near east–west orientation no matter what the traverse direction. The box was mounted on two semi-elliptic ash slats, which acted as springs, and was protected from damage in overturns by aluminium angle strips screwed to the lid. All materials used in the sledge and box were of course non-magnetic. Rope brakes were applied to the sledge in the normal manner to prevent overtaking on steep descents.

Towing the detector unit halved the time taken to read a station and there was a slight decrease in the frequency of cable failure. However, the cable was in a very poor state when towing started so no great improvement could have been expected. Because of the difficulties encountered, it was recommended that for future work a more suitable cable be found. Experiments with a rubber-sheathed cable capable of flexing down to − 50° C proved highly successful and this is now incorporated as standard equipment by the Survey.

The Elsec magnetometers used in these surveys were designed for use between temperatures of − 10° and + 40° C. At temperatures below − 10° C, the output of the crystal-controlled oscillator may vary due to changes of the crystal properties, causing errors in the count. Also, the efficiency of the amplifier transistors is impaired at low temperatures with consequent loss of signal amplification. Unfortunately, much of the survey was conducted in temperatures well below − 10° C so it was found advantageous to take the instrument into the tent during non-travelling hours in order that it could be warmed before the start of the following day's work. The instrument was transported in an insulated box, and with temperatures above − 10° C full-scale deflections and slow signal-decay rates could be observed. During colder periods, or as the instrument cooled through the day's work, signal amplification declined and at ambient temperatures below − 20° C the instrument was almost inoperative, only spurious readings being obtained. The majority of readings were repeated several times and, while the instrument was functioning properly, there was generally a maximum scatter of 2 or 3 gamma at a station.

B. Survey Techniques, Data Correction and Errors

1. *Magnetic survey station positions*

The ideal reconnaissance magnetic survey would entail measurements along regularly spaced traverse lines controlled by reference to a network of accurately determined base stations. With absolute field readings, the need for a base network is not too critical provided diurnal variations are available from a not too distant station. For vertical field determinations, however, where only relative values are measured, it becomes an important consideration. In the Antarctic, even assuming a suitable terrain, many factors influence the accurate establishment of a base-station network. These include mode of travel and poor weather which may cause long delays between the occupation of successive base stations.

For the vertical-field survey between Pitt Point and Longing Gap, the development of a base-station network, outward from an established base, was precluded by the late formation of sea ice in Prince

Gustav Channel. The survey was therefore commenced on the mainland at Longing Gap and the network was extended northward to Pitt Point as the survey proceeded. Six magnetic base stations (Fig. 6) were occupied and these were linked to that of previous surveys by the common base at Pitt Point (base station 24 of Allen (1966*a*)). Wherever possible, double links were made between base stations, reading first at station A then at station B, and immediately returning to make a second reading at station A. This allowed

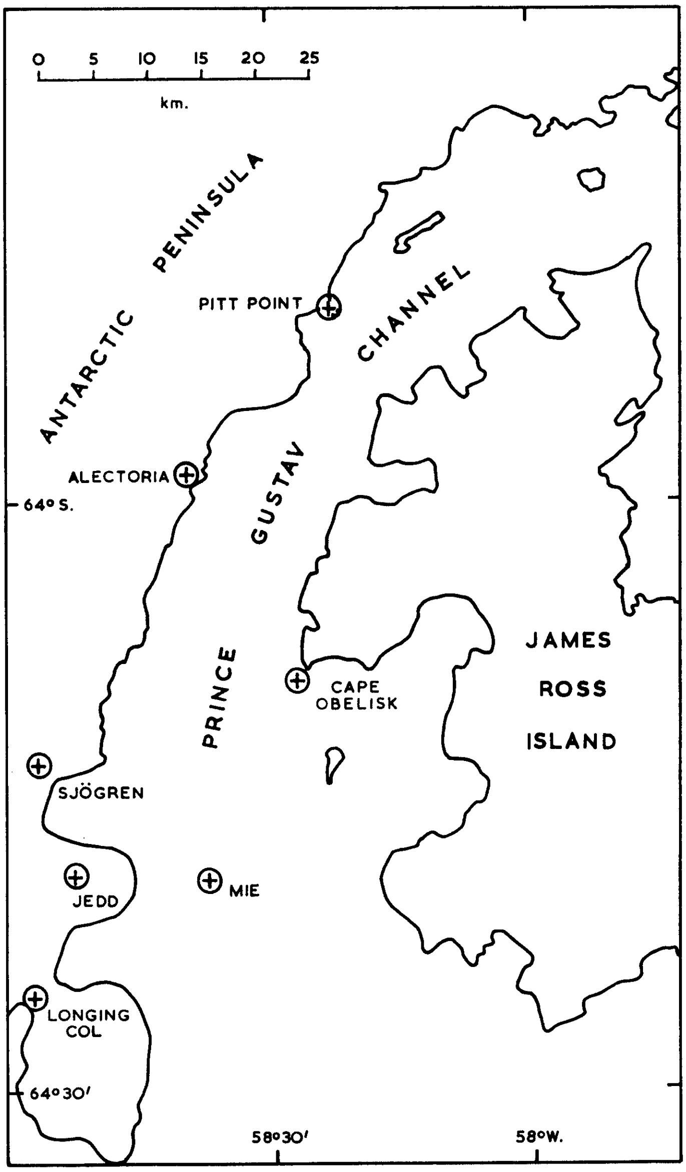


FIGURE 6
Network of vertical magnetic field base stations between Pitt Point and Cape Longing.

drift corrections to be made to the observed differences with consequent improvement in accuracy. Where single links only were made, they were repeated if possible at later dates and the mean difference was calculated.

Because of the adverse sea-ice conditions, fewer links were achieved than had been originally planned, but sufficient were completed to form a series of closed loops between the base stations and along which the magnetic differences could be calculated.

Clearly, the algebraic sum of the magnetic field differences round a closed loop must be zero, any residual remaining after such a summation being designated as the "closure error". The residual can generally be attributed to errors in the instrumental measurement or to drift assessment and can be eliminated by appropriately adjusting the individual links of the loop. This was accomplished by using the graphical least-squares adjustment method described by Smith (1951). Fig. 7a shows the base-station network, the ringed numbers in the centres of the triangles representing the closure error between the relevant base-station differences. The closure error of 12 gamma around the periphery of the whole network has also to be reduced to zero during the least-squares adjustment. Adjustability factors were assigned to each link in the network according to the quality of the link. During the least-squares adjustment these determined the proportional corrections made to each link of a closed loop. Double links were given an adjustability factor of 1 and non-repeated single links a factor of 3. Repeated single links were given an adjustability factor of

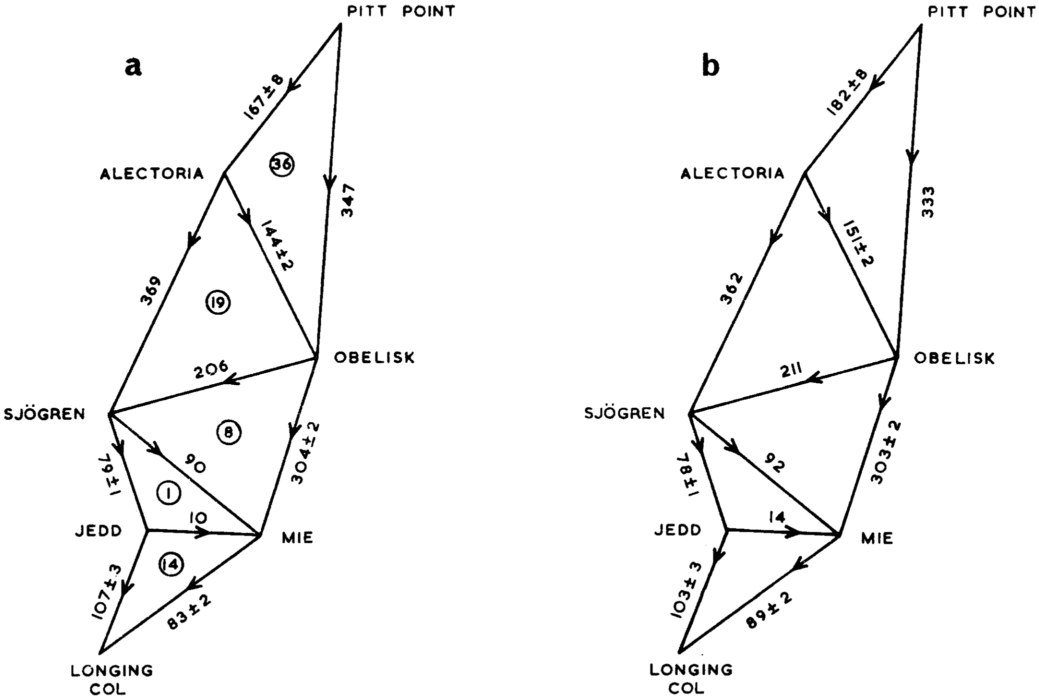


FIGURE 7

a. Magnetic field differences (gamma) between individual base stations. The residual errors are circled.
b. Adjusted vertical magnetic field differences between individual base stations.

2 or 3 depending on the deviation of the measurements. The adjusted base-station differences, together with the standard error of the individual links, are shown in Fig. 7b and the base values themselves are listed in Table II. Within this base-station network the intention was to establish a grid of intermediate stations with a nomal 3·2 km. spacing. The larger unsurveyed areas in the northern part of Prince Gustav Channel are due to the absence of sea ice for most of the working season.

Because the establishment of base stations was not necessary, the programme for total-field magnetic survey was more ambitious. Regional traverses were undertaken, as well as some local detailed work, though there still remain large areas where the survey is incomplete. At its best, between Cape Longing and Cape Disappointment, the regional traverse lines were spaced up to 9·6 km. apart with stations observed at 3·2 km. intervals. South of Cape Disappointment, traverses were far less frequent with station spacing up to 6·4 km. but on the plateau traverse J. Ross obtained readings every 1·6 km. The majority of stations were plotted on the D.O.S. 1:200,000 maps of the area using sledge-wheel readings together with compass bearings taken between pre-determined localities. Unfortunately, trigonometrical stations were not always easily distinguishable even when theoretically visible. In some cases, it was only possible to take bearings on prominent features which at some later date were more accurately fixed by plane-table surveys.

An absolute error of the order of 1·6 km. is estimated for those station positions on parts of the Larsen Ice Shelf which are distant from trigonometrical stations. However, most stations on the ice shelf are expected to be within 1·0 km. of their plotted positions. Stations between Pitt Point and Longing Gap, where

TABLE II

VERTICAL MAGNETIC FIELD BASE-STATION VALUES USED IN THE SURVEY FROM PITT POINT TO LONGING COL

Base-station reference number		*Base-station locality*	*Lat. °S.*	*Long. °W.*	*Vertical magnetic field values* (gamma)
Allen 24	=	Pitt Point	63°50′30″	58°23′00″	+ 629±10
Mansfield 29	=	Alectoria	63°58′30″	58°40′00″	+ 796±13
Mansfield 30	=	Obelisk	64°09′00″	58°28′00″	+ 947±13
Mansfield 31	=	Sjögren	64°13′00″	58°57′00″	+ 158±14
Mansfield 32	=	Jedd	64°19′00″	58°53′00″	+1,236±14
Mansfield 33	=	Mie	64°19′00″	58°37′00″	+1,250±14
Mansfield 34	=	Longing Col	64°25′30″	58°58′00″	+1,339±15

trigonometrical stations are more frequent, are probably accurately plotted to within 0·2 km. While the absolute station errors may be large over some of the regional traverses, the relative positions of the stations along individual traverse lines are much better due to sledge-wheel mileometer control. With a station interval of 3·0 km. along traverses, the relative station-to-station error is estimated to be less than 0·2 km.

2. *Errors in Antarctic magnetic surveys*

The diurnal and secular variations are features of the Earth's magnetic field which must be considered in most magnetic surveys. The secular variation may be neglected in isolated surveys over a limited period of time but larger accumulative errors are introduced if no allowances are made for the diurnal changes. Ideally, a systematic base-station network covering the survey area would give the necessary control but this is not always possible in the Antarctic, especially where regional traverses are being made. An alternative is to use available base-station data and to extrapolate over the intervening distances, though Bentley (1964) stated that little correlation in diurnal variations can be seen over distances greater than 100 km. However, Adams and Christoffel (1962) obtained diurnal correlation in the Ross Sea area out to 192 km. from a magnetic base station. By reading two proton precession magnetometers at intervals along a traverse at the southern end of the Antarctic Peninsula, Behrendt (1965) was able to make diurnal corrections which gave a relative accuracy of ± 7 gamma between intermediate stations. During part of that traverse the field diurnal measurements were compared with the total intensity diurnal variations computed from horizontal and vertical intensity magnetograms recorded at Eights station. A standard deviation of ± 15 gamma was found between the two sets of data over a 20 day period. Although there were marked differences between field and observatory diurnal data even within distances of 100 km. of one another, it was considered that when only one field instrument was in operation base diurnal variations could justifiably be applied to field stations up to 400 km. away. Aeromagnetic traverses partly overcome the problem of diurnal variation, as the flights can be completed on magnetically quiet days. Behrendt and Wold (1963) adopted an anomaly minimum of 200 gamma from their investigations. Field data were not corrected for diurnal variations and, as fluctuations of around 200 gamma had been recorded on the base magnetogram during the flight, it was decided to set this minimum anomaly value. For aeromagnetic work in the vicinity of Mirny, Glebovskiy (1959*a*) anticipated an accuracy in results of ± 50 gamma though a lower order of accuracy was encountered by the same author for an aeromagnetic survey a short distance to the west (Glebovskiy, 1959*b*). Access to suitable base-station records is also inherent to marine magnetic surveys, though Griffiths and others (1964) estimated the error in the Scotia Sea due to lack of diurnal control to be no greater than ± 50 gamma.

A second significant source of error results from navigational uncertainties. In ground surveys, positions can be fixed to the same order of accuracy as for seismic and gravity work, namely to less than 3 km., but with marine and aeromagnetic surveys the accuracy of navigation is much more questionable. For aeromagnetic work, Behrendt and Wold (1963) determined possible errors of up to 50 km. The problem therefore becomes critical where networks of traverses are being considered. Instrumental errors, even if amounting to ± 10 gamma in total-field proton magnetometers, are not of significance in the overall accuracy. For land traverses in Ellsworth Land and the southern Antarctic Peninsula, Behrendt (1965) gave a standard deviation in the absolute value of ± 25 gamma with a relative accuracy between intermediate stations of ± 7 gamma. Over Trinity Peninsula, the vertical-field surveys of Allen (1966*a*) produced maximum overall errors of ± 36 gamma but mean values of ± 15 gamma are probably more representative for the area.

3. *Diurnal and secular variation*

At any given observation point in low and middle latitudes, the diurnal variation in the Earth's magnetic field depends primarily on the latitude and local time. At higher magnetic latitudes, particularly those within the auroral zones, the two-fold dependence is further complicated since the longitude is also of significance. The magnetic survey carried out on the Antarctic Peninsula, although reaching a geographical latitude of 70° S., lies at a lower geomagnetic latitude and which is outside the main disturbed areas of diurnal variation. For ideal control in a magnetic survey, the diurnal correction applied to the field readings should preferably be that recorded as near to the field locality as possible. The survey over Graham Land covers over 6° of latitude and it has not been possible to occupy continuous recording magnetic stations in sufficient numbers to give ideal coverage. However, the magnetic observatories at Hope Bay and the Argentine Islands (lat. 65° 15′ S., long. 64° 16′ W.) have been used by the British Antarctic Survey to record variations in the Earth's magnetic field, and the results have been variously applied to the field data.

The observatory at Hope Bay, which used an Askania Gf6 magnetometer for recording variations in the vertical component, has since been dismantled but the one at the Argentine Islands still continues to record variations in both the horizontal and vertical components of the Earth's magnetic field. (All magnetic records are housed at the British Antarctic Survey's headquarters, Madingley Road, Cambridge CB3 0ET.) The observatory at the Argentine Islands is approximately 400 km. from and midway between the stations at Hope Bay and Stonington Island. Ashley (1962), working in the vicinity of Hope Bay, used the Askania Gf6 diurnal records with an estimated minimum error of ± 4 gamma being attributed to a combination of errors from temperature effects, calibration and incorrect reading of the records. The field survey by Allen (1966*a*) was partly completed using the base Askania Gf6 and it was therefore necessary to obtain some diurnal records from the Argentine Islands. Allen found that the smallest daily closure to the field readings was obtained by the direct use of the Argentine Islands values without any correction to either amplitude or to phase changes. For his total magnetic field survey in the immediate vicinity of Hope Bay, Cox (1964*a*) was able to compare the records obtained at Hope Bay and at the Argentine Islands. Examination of the respective variations revealed that, although the Hope Bay records were always the more disturbed, the main peaks and troughs were approximately (to within 12 min.) synchronous, with the ratio of the amplitudes varying from 1·5:1 to 3·0:1. The diurnal variation used (that of the total-field variation calculated from the Argentine Islands and supplemented with field evidence) was estimated to have an accuracy of ± 3 gamma. A local vertical-field survey completed by Agger and others (1963) at the Argentine Islands also had an accuracy of this order. The geophysical surveys from Hope Bay in 1963 included both vertical-field and total-field measurements, reaching as far south as Cape Disappointment about 340 km. from the base station and approximately in the same latitude as the Argentine Islands. The diurnal charts from the latter station were required for the total-field survey but it was also possible to examine the vertical components recorded at both base observatories. Over the 17 days examined, good correlation was shown between the two (personal communication from J. Mansfield) but again those of Hope Bay exhibited greater disturbance with the amplitude ratios varying from 1·0:1 to 4·0:1. Because of the encouraging correlation shown, the continuity of the Argentine Islands records, the questionable temperature effects mentioned previously on the Gf6 magnetometer at Hope Bay, and the field areas approaching the latitude of the Argentine Islands, it was decided to use the diurnal records from the more southerly observatory.

Only total-field magnetic surveys have been completed from Stonington Island and, in the absence of local base diurnal records, it was necessary to refer again to the Argentine Islands records. After the completion of a local survey in Marguerite Bay, Kennett (1966*b*) compared the diurnal variation observed at a local field station with that measured at the Argentine Islands. A very poor correlation was found which did not justify the use of the observatory data as a source for diurnal corrections beyond that of indicating general magnetic conditions. Kennett estimated that because of this the expected maximum error in this survey was $\pm$ 25 gamma. During the survey on the east coast of the Antarctic Peninsula, Kennett (1966*a*) carried out two further diurnal checks. One of these was on a magnetically disturbed day at Mill Inlet, 190 km. south of the Argentine Islands, and the other was on a magnetically quiet day at Churchill Peninsula, 120 km. south of the Argentine Islands. Comparison with the observatory values shows a correlation with a standard deviation of $\pm$ 10 gamma on the disturbed day and $\pm$ 2 gamma on the quiet day. Since the magnetic events in both of the areas occurred simultaneously and were of the same amplitude, it was decided to apply the Argentine Islands diurnal values directly to the east coast field data. During the 1963–64 field season, and within 6 weeks of one another, the base station at Cape Disappointment (130 km. east of the Argentine Islands) was occupied on independent occasions by P. Kennett and J. Mansfield. After applying the relevant Argentine Islands values for both diurnal and secular variation, the field values agreed to within 1 gamma.

In 1964–65, the proton magnetometer was used to record the diurnal variation in the Earth's total magnetic field. Four such measurements were made at Stonington Island by the author, and several were made on the east coast of the Antarctic Peninsula. It would have been desirable to run more such checks but several factors, including the following, prevented this:

i. Throughout the autumn journey (April–June 1964) and in the early part of the main field journey (September–October 1964) low temperatures severely affected the performance of the magnetometer.

ii. The condition of the batteries was always a critical factor as the possibilities of having them re-charged in the field were limited. This restricted their use to essential magnetic stations, permitting diurnal measurements only when the battery condition allowed.

iii. Over much of the period, geological work had precedence over the magnetic survey which often limited traverses to those of a local nature.

After returning from the Antarctic, magnetograms from the Argentine Islands observatory were obtained for days corresponding to those on which field diurnal measurements had been taken. Fig. 8a and b illustrates the comparison. It is evident that those recorded at Stonington Island and on the east coast show a strong similarity to those from the Argentine Islands, the magnetic events correlating both in the time and amplitude dimensions. From these results, it was again considered justifiable to apply directly the observatory diurnal variation records for the correction of the field results. An estimated error of $\pm$ 2 gamma for a magnetically quiet day is in accordance with that estimated by Kennett (1966*a*). Total-field magnetic readings were taken on more than 70 days in 1964, 9 days of which could be classed as magnetically very disturbed and one of which was stormy. Disregarding these very disturbed days, when the maximum correction applied was -76 gamma, the diurnal corrections varied between -40 and $+40$ gamma.

In addition to the correction for the diurnal variations, the readings were also subjected to a further correction for the secular variation of the Earth's magnetic field.

Nagata (1962), reporting on the geomagnetic secular variation for 1955–60, has given the annual rate of secular variation at the Argentine Islands for the X, Y and Z components over the period 1956–58 as -34, -48 and -131 gamma, respectively. Behrendt and Bentley (1968, pl. 4) have given an annual change of about -115 gamma in the total magnetic intensity at the Argentine Islands. From the monthly mean values measured at the Argentine Islands observatory by the British Antarctic Survey, the secular variation has averaged $-9{\cdot}8$ gamma/month over the period January 1961–December 1964. During the period of the total-field survey, however, the variation has shown a slightly smaller decrease of $9{\cdot}5 \pm 2$ gamma/month. All of the total-field results on the Antarctic Peninsula have been reduced to 30 September 1963, using the data supplied by the Argentine Islands observatory.

The vertical-field survey, which measured relative values, has been tied in to the earlier surveys in north-eastern Trinity Peninsula through the re-occupation of a common base station.

C. Regional Magnetic Field

1. *Total magnetic intensity*

After corrections for both the diurnal and secular variation, it was necessary to derive and remove the regional field in order to produce the residual anomalies. A first approximation to the geomagnetic field was

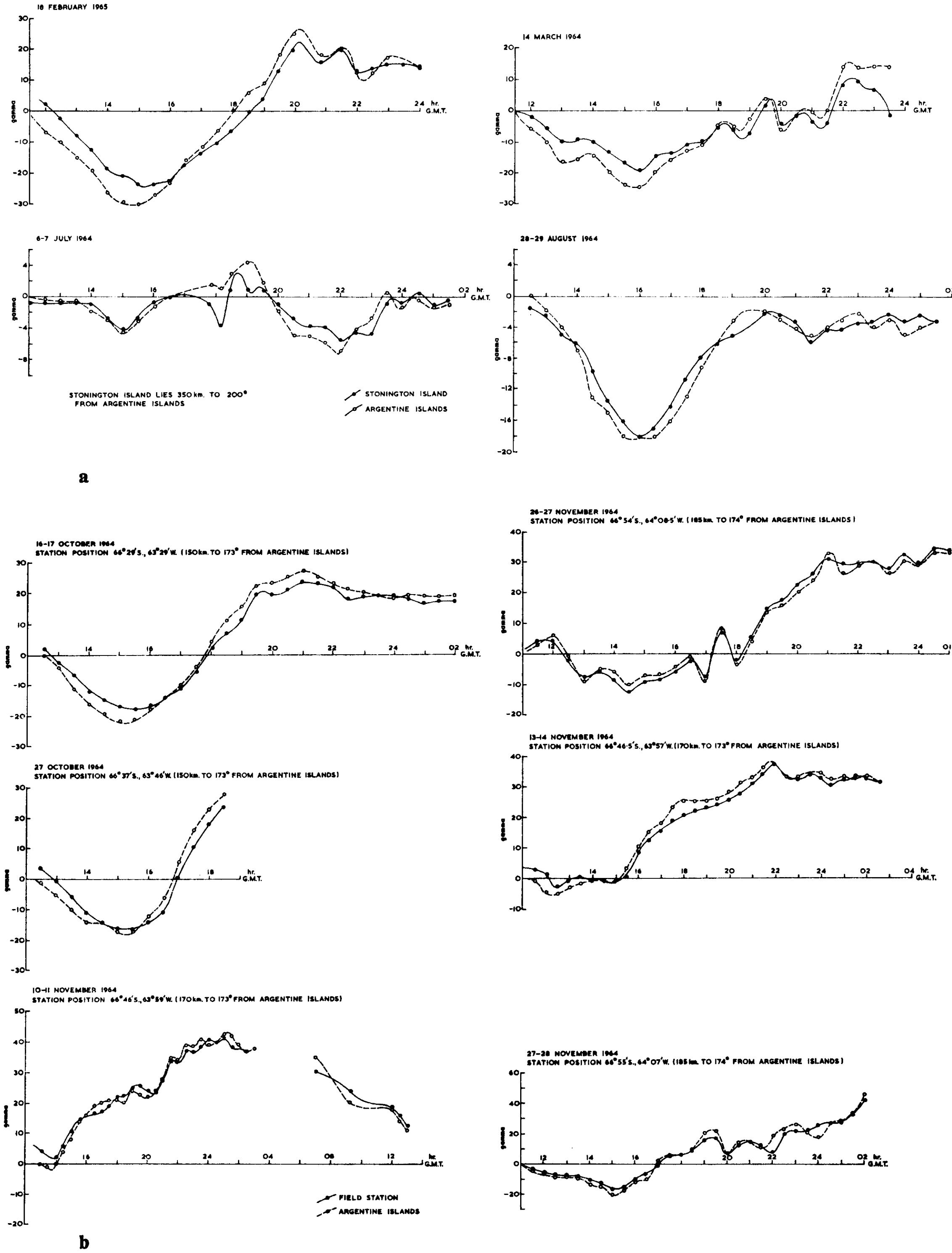


FIGURE 8

a. Comparison of the diurnal variation recorded at the geomagnetic observatory at the Argentine Islands with that observed from measurements at Stonington Island.
b. Comparison of the diurnal variation recorded at the geomagnetic observatory at the Argentine Islands with that observed at field stations on the east coast of Graham Land.

made visually but this was superseded by the values obtained from the application of a polynomial expression. P. F. Barker (personal communication) had previously calculated the regional field from the total-field magnetic traverses across the Scotia Sea by means of a computer programme which incorporates a third-order three-dimensional polynomial:

$$Z = B_0 + B_1 X + B_2 X^2 + B_3 X^3 + B_4 Y + B_5 X Y + B_6 X^2 Y + B_7 Y^2 + B_8 X Y^2 + B_9 Y^3 + Ct + R,$$

where Z is the observed field, B is a coefficient, X is the component of latitude, Y is the component of longitude, C is the secular variation, t is the time (relative to reference date), and R is the residual field strength.

The same computer programme was applied to over 700 of the total-field magnetic measurements taken on the east coast of Graham Land between Cape Longing and Mobiloil Inlet. Over 100 stations were omitted because it was considered that they would affect the true regional gradient; these included readings from significantly large anomalous areas. The least-squares criterion was used to assess the accuracy of the polynomial field. The standard deviation of the residual values for the first-order field was $\pm$ 171 gamma; this includes both data variations and errors. The second- and third-order standard deviations gave values of $\pm$ 151 and $\pm$ 149 gamma, respectively. However, when plotted, the second- and third-order fields showed an undesirable edge effect, which was not consistent with the overall regional gradient. Because of this, it was decided to concentrate only on the first-order field which has the form of a linear gradient.

Previous estimates of the total magnetic regional gradients over Graham Land have been largely extrapolated from the data recorded at geomagnetic observatories, or from scattered traverses near to or over the area. On the Antarctic Peninsula, two observatories have been used: that of the Argentine Islands and "Base Gabriel González Videla" (lat. 64° 49′ S., long. 62° 52′ W.). In addition, continuous magnetic measurements have been made from ship-borne traverses across the Scotia Sea and from three Project Magnet air-borne traverses over Graham Land. The accuracy of the Project Magnet data is, however, questionable due to possible navigational errors. Fig. 9 summarizes the estimated regional field gradients for Graham Land. Secular variation corrections have been applied by the author to bring all values relative to September 1963. The compilation of this diagram from the respective publications necessitated the transfer of the relevant information from maps of varying scales, accuracy and projection. In consideration of this, and from the sparsity of data sources, there is an acceptable correlation particularly in strike and gradient between the regional field intensities given by the various authors. At the northern end of Graham Land, extrapolation of the regional field shows a marked correlation with that computed from the measurements made over the Scotia Sea (Kroenke and Woollard, 1968; personal communication from P. F. Barker). The International Geomagnetic Reference Field (IAGA Commission 2 Working Group 4 Analysis of the Geomagnetic Field, 1969) has been computed by Fabiano and Peddie (1969) as grid intersection tables of latitude and longitude. Over the area considered here, a high level of agreement is exhibited with that determined by the author. Differences in absolute values become more marked to the south of this area due to a poor field coverage and therefore for future surveys outside its immediate vicinity it is suggested that the International Geomagnetic Reference Field be used in preference to any extrapolation of the regional field computed by the author.

The land measurements indicated that the Earth's total magnetic intensity increases from 42,000 to 46,000 gamma between Cape Longing and Mobiloil Inlet. It shows an approximate increase to the south and west of 7·7 and 3·4 gamma/km., respectively, the lines of equal magnetic intensity trending 113° E. of true north.

2. *Vertical magnetic intensity*

The distribution of the vertical magnetic stations and their associated anomalies did not permit the application of a polynomial expression in the calculation of the vertical intensity. The method decided upon was to use the calculated total-field intensity together with the corresponding magnetic inclination, which varies over the area from 56·5° to 58·5° (Behrendt and Bentley, 1968, pl. 1). Between Pitt Point and the Seal Nunataks the vertical-field intensity was found to increase from 34,600 to 36,000 gamma. The strike over the area was very similar to that of the total magnetic field but the southern and western gradients averaged + 8·8 and + 3·3 gamma/km., respectively. These values compare with those of + 8·1 and + 3·0 gamma/km. calculated by Allen (1966*a*) over north-eastern Trinity Peninsula, and which were based on extrapolation of the total-field magnetic gradient as determined on the southern side of Drake Passage by Griffiths and others (1964).

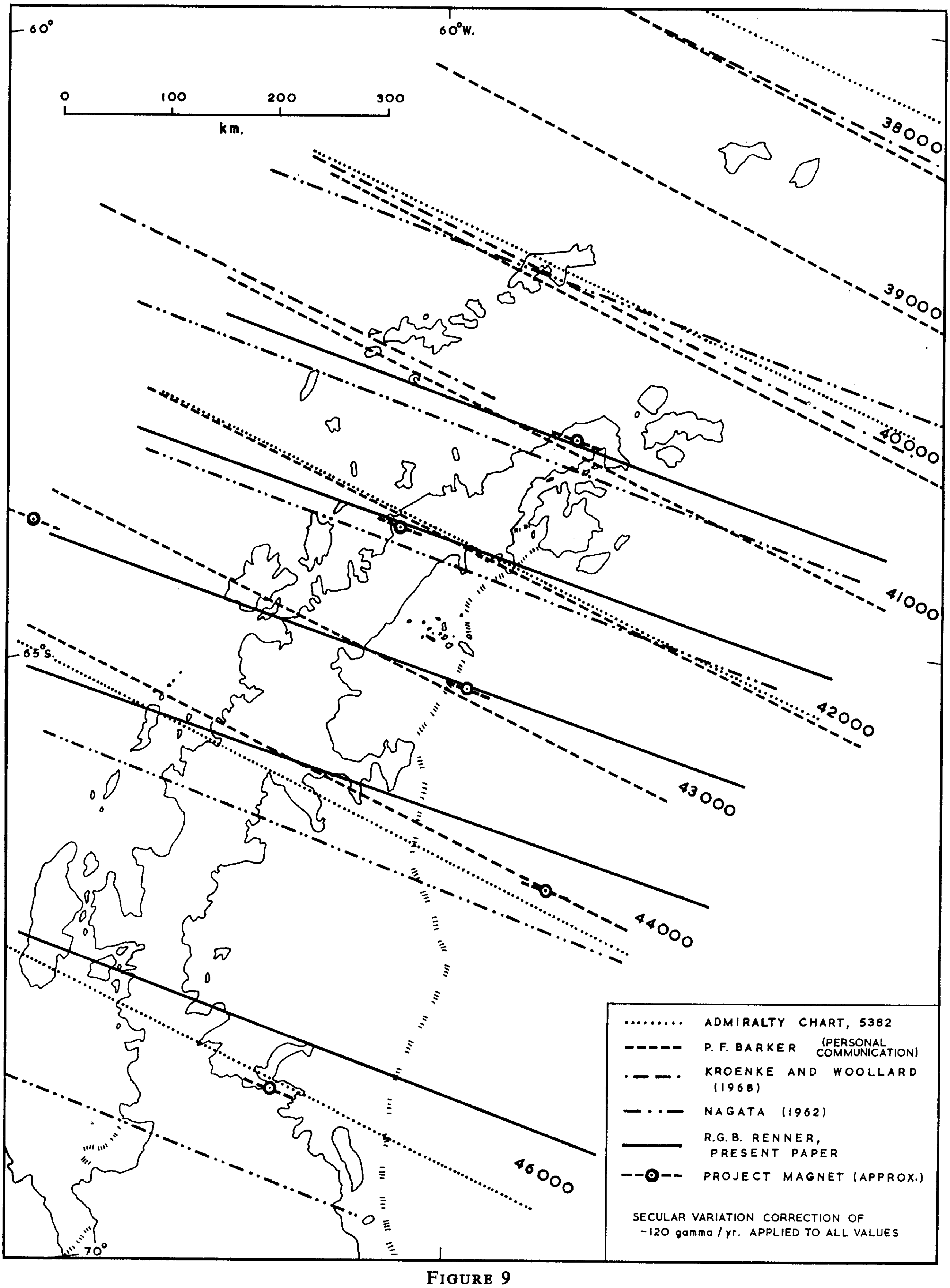


FIGURE 9

Estimates of the Earth's regional total magnetic field over Graham Land.

D. Magnetic Properties of Rock Specimens from Graham Land

The induced magnetization, I per unit volume, imparted by the Earth's magnetic field to any rock containing ferromagnesian minerals is given by $I=kH$, where k is the volume susceptibility of the rock and H is the magnitude of the Earth's magnetic field. However, in many rocks, particularly lavas, the remanent magnetization is dominant and plays an important part in governing the characteristics of associated magnetic anomalies. Because of this, the remanent and induced magnetizations must be combined vectorially to give the resultant magnetization (Green, 1960). Division of the latter by the Earth's magnetic field gives the equivalent susceptibility, the value of which can be considerably greater than the measured volume susceptibility.

Thermo-remanent magnetization is the commonest variety of remanent magnetization likely to be found in the igneous rocks of Graham Land. It is imparted to the rock body by the ferromagnetic minerals acquiring a magnetization during their cooling from the original magma. Detrital or chemical remanent magnetization may also contribute to the overall natural remanent magnetization but this is considered insignificant in the rock bodies studied. Ashley (1962), Allen (1966*a*) and Blundell (1962) made a series of measurements on specimens from Graham Land and determined the intensity, direction and stability of their magnetizations. From the published results for the Andean Intrusive Suite and later rocks, it would appear that the direction of remanent magnetization lies close to that of the Earth's present axial dipole field. Therefore, it was assumed that where sample measurements were not taken it was justified to use a magnetic dip similar to that of the Earth's present magnetic field. Reversed directions of magnetization have been recorded but such anomalies usually become apparent during the field survey or in the subsequent data reduction.

With the exception of the isolated values given by Kennett (1966*a*) and those in Table III, there are no measurements of magnetic properties of rocks for the east coast south of Trinity Peninsula. Interpretation of anomalies has, therefore, been initially guided by using approximate susceptibility values based on the known geology and from the resultant magnetizations calculated for similar rock types in north-eastern

TABLE III

VALUES OF VOLUME SUSCEPTIBILITY

Rock type	*Locality*	*Number of samples measured*	*Volume susceptibility*	
			Range of values (c.g.s. units/cm.$^3\times10^{-6}$)	*Mean value* (c.g.s. units/cm.$^3\times10^{-6}$)
Upper Jurassic volcanic rocks				
Tuff*	Astro Cliffs, Cape Alexander	9	340–520	430±70
Rhyolite*	Nunatak, 8 km. west of Gemini Nunatak	6	230–950	640±29
Lava†	Tower Peak	2	36–39	38
Andesite†	Tower Peak	2	450–530	490
Andean Intrusive Suite				
(?) Diorite* (described by Marsh (1968) as gabbro)	Peak 1040, Churchill Peninsula	8	2,400–4,700	3,600±900
Granodiorite* (? age)	Gulliver Nunatak	2	130	130
Sediments				
Trinity Peninsula Series†	lat. 64°18′ S., long. 58°58′ W., 3·2 km. west-north-west of Jedd trig. station	11	18–25	21±2

* Specimens collected and measured by Kennett (1966).
† Specimens collected and measured by J. Mansfield.

Trinity Peninsula. Recently, susceptibility measurements (Table IV) have been completed on rocks collected by P. J. Rowe along a section on the west coast of Palmer Land. Although they have no direct bearing on the interpretation, they provide an interesting comparison with similar rocks found elsewhere on the Antarctic Peninsula.

TABLE IV
SUSCEPTIBILITIES OF ROCKS FROM THE WEST COAST OF PALMER LAND BETWEEN LAT. 70°10′–70°42′S. AND LONG. 66°08′–67°38°W.

Rock type	*Number of measurements*	*Range of volume susceptibility* (c.g.s. units/cm.$^3 \times 10^{-6}$)	*Mean volume susceptibility* (c.g.s. units/cm.$^3 \times 10^{-6}$)
Syenite	8	3,076–4,423	3,531±382
(?) Dolerite	8	265–398	321±52
Dolerite	11	2,990–3,649	3,159±293
Diorite	5	4,000–4,505	4,301±213
Basalt	8	3,222–3,796	3,529±102
Andesite	5	3,500–7,000	4,402±403
Basalt	5	5,473–6,244	6,106±422
Xenolithic dyke	6	2,252–3,017	2,704±268
Gabbroic gneiss	6	2,566–2,973	2,759±205
Basalt	9	2,545–3,077	2,889±143
Volcanic ash	9	1,415–1,769	1,607±179
Agglomerate	1	4,112	4,112
Banded tuff	3	1,283–1,944	1,607±269
Trachytic lava	5	3,800–5,387	4,426±548
Volcanic ash	4	94–163	128±26
Tuff-agglomerate	5	584–1,400	968±258
Dioritic gneiss	7	1,684–2,120	1,834±132
Basalt	9	2,222–3,628	2,877±383
Altered basalt	9	212–506	307±92

E. Interpretation of Magnetic Anomalies

1. *North-east Trinity Peninsula*

In addition to the vertical-field surveys undertaken by Ashley (1962) and Allen (1966*a*), a total-field magnetic survey (Fig. 10) was run over part of the same area by J. Mansfield. Although the station network was not as close-spaced as that of the vertical-field survey, it did have sufficient coverage over two sections of Trinity Peninsula to warrant an independent interpretation.

The two areas (Fig. 11) investigated were:

i. Mott Snowfield—this includes a traverse from Mount Bransfield to Cugnot Ice Piedmont via Magnet Hill, Camel Nunataks, Theodolite Hill and Stepup Col.
ii. The northern part of Tabarin Peninsula, including a traverse from Hope Bay via Summit Pass (lat. 63° 27′ S., long. 57° 02′ W.) and Last Hill (lat. 63° 28′ S., long. 57° 05′ W.) to Duse Bay.

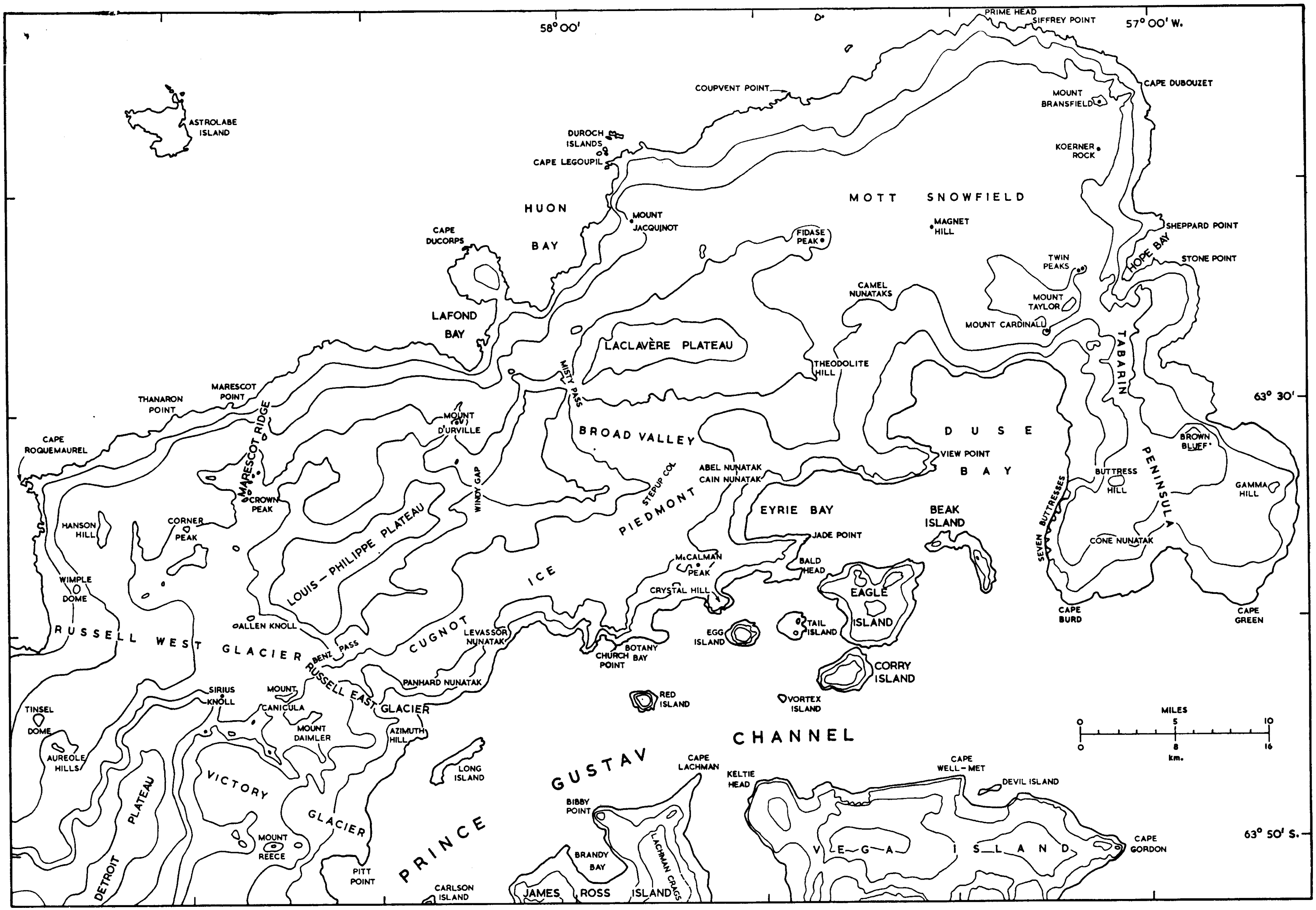


FIGURE 11

Location diagram of north-eastern Trinity Peninsula.

The original interpretations of the data from these areas involved the use of a three-dimensional graticule method supplemented with estimates using the empirical depth rules of Peters (1949). Due to the limited data coverage, it was decided that for the present interpretation only two-dimensional bodies could be considered, the anomaly being fitted by a computer technique based on the programme described by Heirtzler and others (1962).

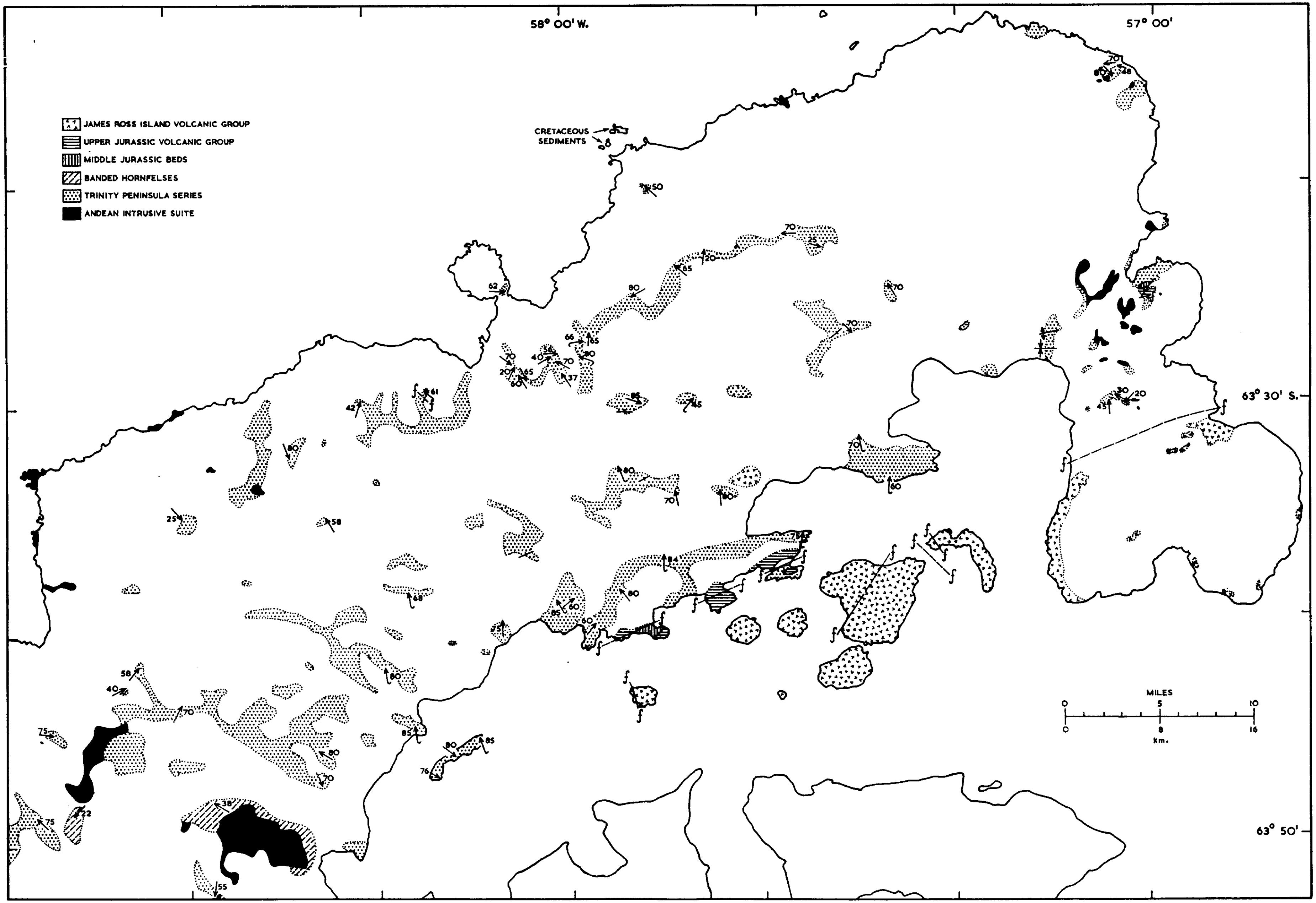
Published geological information for this area includes that of Adie (1957, 1964*a*) Bibby (1966), Elliot (1965) and Aitkenhead (1965), while a more detailed geological description of Trinity Peninsula is given here to supplement that on p. 13–15.

Fig. 12 is a geological sketch map of north-eastern Trinity Peninsula, from which it can be seen that the most abundant of the stratigraphical units (Table V) is the Trinity Peninsula Series. These, the oldest rocks in this area, are late Palaeozoic (? Carboniferous) in age and are a succession of metamorphosed, unfossiliferous geosynclinal sediments. Maximum recorded thicknesses vary from about 3,500 m. in the north-west to over 12,000 m. on the eastern side of the peninsula. Early Mesozoic earth movements have folded the sediments into a major anticlinorium, the main axis of which lies to the south-east of this area. The accompanying regional metamorphism increases in grade from the north-east to south-west. The sediments which crop out extensively on the Detroit, Laclavère and Louis Philippe Plateaux are thermally metamorphosed where they are in contact with rocks of the Andean Intrusive Suite.

The banded hornfelses are a very localized group, which is pre-Upper Jurassic in age and so far found

TABLE V
STRATIGRAPHY OF TRINITY PENINSULA

Age		*Stratigraphical unit*	*Thickness* (m.)
Tertiary	Pliocene	*Pecten* Conglomerate	
	(?) Middle Miocene	James Ross Island Volcanic Group	1,530
	(?) Lower Miocene	Seymour Island Series	150
	Early Tertiary to late Cretaceous	Andean Intrusive Suite	
Mesozoic	Cretaceous (Lower–Middle Campanian)	Snow Hill Island Series	3,580
	Cretaceous	Sediments of Cape Legoupil	3,960
	Upper Jurassic	Andesite-rhyolite volcanic group	3,000
	?	Banded hornfelses	153
	Middle Jurassic	Mount Flora plant beds	306
Palaeozoic	(?) Carboniferous	Trinity Peninsula Series	12,000



58° 00' W.
57° 00'
63° 30' S.
63° 50'
JAMES ROSS ISLAND VOLCANIC GROUP
UPPER JURASSIC VOLCANIC GROUP
MIDDLE JURASSIC BEDS
BANDED HORNFELSES
TRINITY PENINSULA SERIES
ANDEAN INTRUSIVE SUITE
CRETACEOUS SEDIMENTS
MILES
0
5
10
0
8
16
km.

FIGURE 12

only in the vicinity of the Aureole Hills. They are mineralogically different from the adjacent Trinity Peninsula Series sediments and can probably be correlated with the post-Trinity Peninsula Series sediments which are present at the head of Victory Glacier.

Upper Jurassic volcanic rocks, which post-date the regional metamorphism of the Trinity Peninsula Series sediments, consist of andesite-rhyolite tuffs, lavas and tuff-agglomerates. They are exposed in a well-defined zone along the western shore of Prince Gustav Channel; this prompted Bibby (1966) to suggest that structural control had influenced their distribution. Further exposures of Upper Jurassic pyroclastic rocks occur in the vicinity of Hope Bay, Poynter Hill and Tower Peak.

By far the most important geological unit in Trinity Peninsula, as far as the geophysical survey was concerned, is the Andean Intrusive Suite. It occurs throughout this area as a system of batholiths intruded along the central part of the contemporaneously uplifted belt of rocks belonging to the Trinity Peninsula Series and the Upper Jurassic Volcanic Group. The composition of the plutonic bodies is variable, ranging from intermediate to acid, with the more acid representatives occurring mainly in the west. A diorite is exposed around Summit Ridge on Tabarin Peninsula, while a granodiorite and a quartz-gabbro crop out at Mount Bransfield and Coupvent Point, respectively. No representatives have been found in the Laclavère Plateau and Broad Valley areas but farther to the south exposures occur on either side of the Detroit Plateau. North of Russell West Glacier, a variety of compositions has been recorded: a quartz-gabbro at Thanaron Point and a granite at Wimpole Dome. To the south-west there are exposures of granodiorite at Poynter Col, while close to the Aureole Hills along the plateau edge there are tonalites, adamellites and granodiorites. At a similar latitude on the eastern side of the Detroit Plateau, there is a granitic body at Mount Reece.

Apart from an isolated exposure in Broad Valley, the rocks belonging to the James Ross Island Volcanic Group occur east of a line which crosses Tabarin Peninsula and runs down the western side of Prince Gustav Channel. As well as being exposed on Tabarin Peninsula, there are extensive outcrops on James Ross Island and on the islands at the northern end of Prince Gustav Channel. The James Ross Island Volcanic Group is represented by a sequence of olivine-basalt lava flows, tuffs and agglomerates which unconformably overlie the Cretaceous sediments.

a. *Mott Snowfield.* Because of the distribution of the total-field data, the section selected over Mott Snowfield (Fig. 13) runs parallel to and 5 km. to the north of profile A (Allen, 1966*a*) and also extends a further 30 km. to the south-west. Station spacing south-west of Camel Nunatak was approximately at 3 km. intervals with the traverse by-passing Theodolite Hill. A profile from Prime Head south-west for 55 km. to Windy Gap, traversing Laclavère Plateau and Misty Pass, presents a very similar situation to the one interpreted by Allen over Mott Snowfield. This was to be expected as both profiles are parallel and within 10 km. of one another, and both traverse the same magnetic anomalies. Although no interpretation has been carried out by the author on the second profile, it is clear that the bodies producing the anomalies are a north to north-easterly extension of those beneath Mott Snowfield, and they have a similar sub-surface distribution. The section over Misty Pass coincides with profile F of Allen (1966*a*), who completed a two-dimensional interpretation over this section. He attributed the cause of the negative magnetic gradient to possible structural features within the batholith roof, more specifically in the nature of a fault with a 1 km. downthrow to the north-east and a dip of 60°; the depth to the centre of the fault structure was calculated at 4·5 km. Geological evidence supports faulting through Misty Pass (Elliot, 1965). However, in any fault diagnosis in Broad Valley from geophysical evidence, one must consider that the negative anomaly may be partially due to the southern limb of the more extensive anomaly attributed to the intrusive body to the north-west.

Exposures of the Andean Intrusive Suite granodiorite at Mount Bransfield, diorite at Blade Ridge (lat. 63°25′ S., long. 57°05′ W.) and quartz-gabbro at Coupvent Point gave a mean resultant magnetization of $(4{,}060 \pm 860) \times 10^{-6}$ c.g.s. units (Allen, 1966*a*) with a direction along the Earth's present field. Assuming the intensity of the Earth's magnetic field in this area to be of the order of 41,000 gamma, the equivalent susceptibility is approximately 0·01 c.g.s. units. Susceptibilities of the Andean Intrusive Suite rocks can show much variation between sampling localities over north-eastern Trinity Peninsula. In the Mott Snowfield area they show relatively high values, which for interpretation purposes must be considered uniform over the anomaly investigated.

After repeated model fitting, Fig. 14 was considered to give the best fit to the observed vertical- and total-

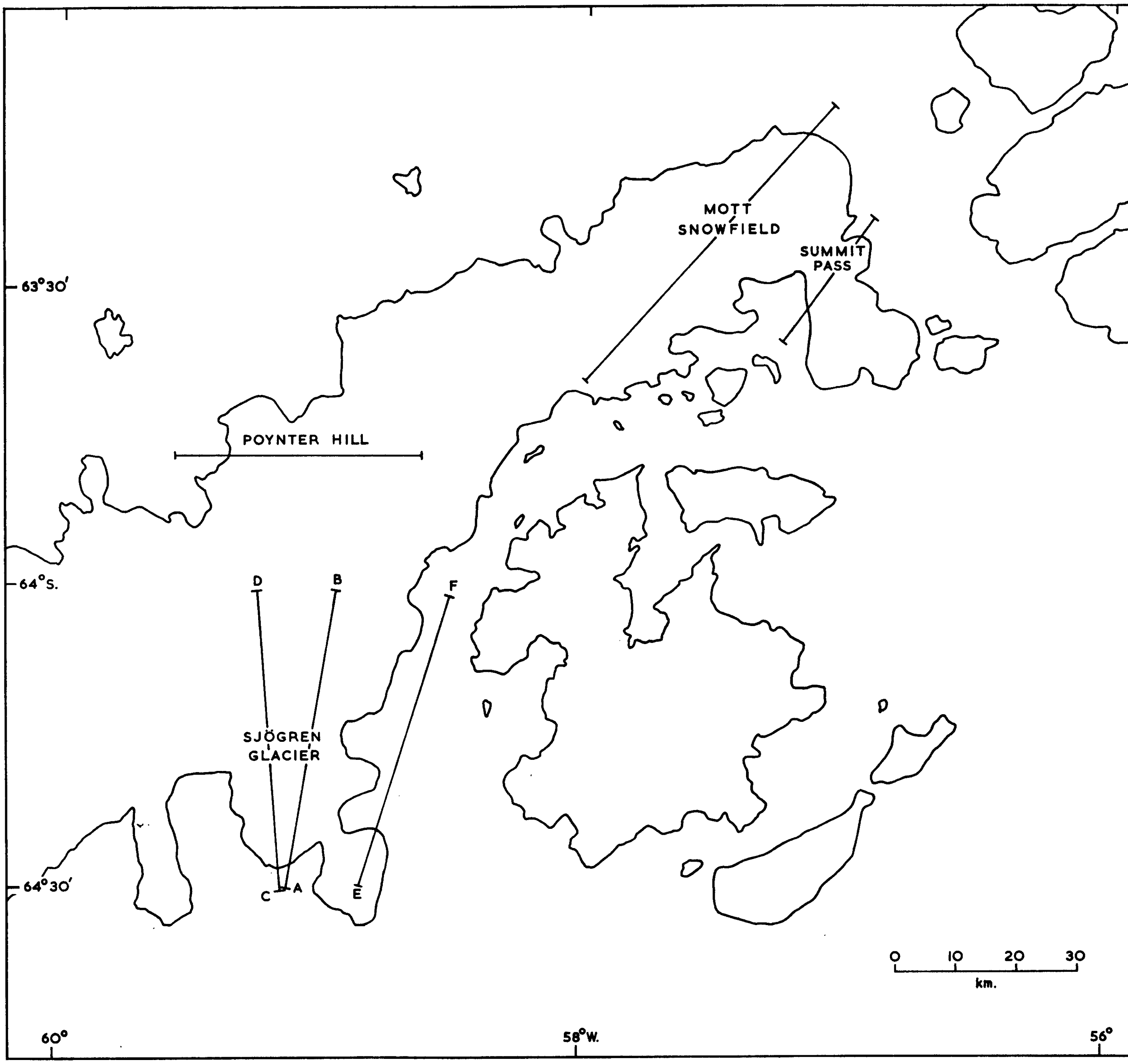


FIGURE 13
Location diagram of the interpreted magnetic sections over Trinity Peninsula.

field magnetic data. The upper surface of the intrusive body compares favourably in position with that postulated by Allen (1966*a*) over the same area, though Allen's section required moving about 3 km. to the south-west for the near-surface extremities of the body to coincide with the anomalies in the vicinity of Magnet Hill. This body approaches the surface in two places, the shallowest of which is at the northern end of the section about 3 km. from the coast. Here it is estimated to be at a depth of about 65 m. below surface level, whereas 1 km. farther south the same part of the body is calculated to approach within 20 m. of the surface. In the vicinity of Magnet Hill, the depth to the upper surface of the anomaly is estimated to be less than 350 m.; Allen (1966*a*) calculated the upper surface to lie within the depth range of 100–400 m.

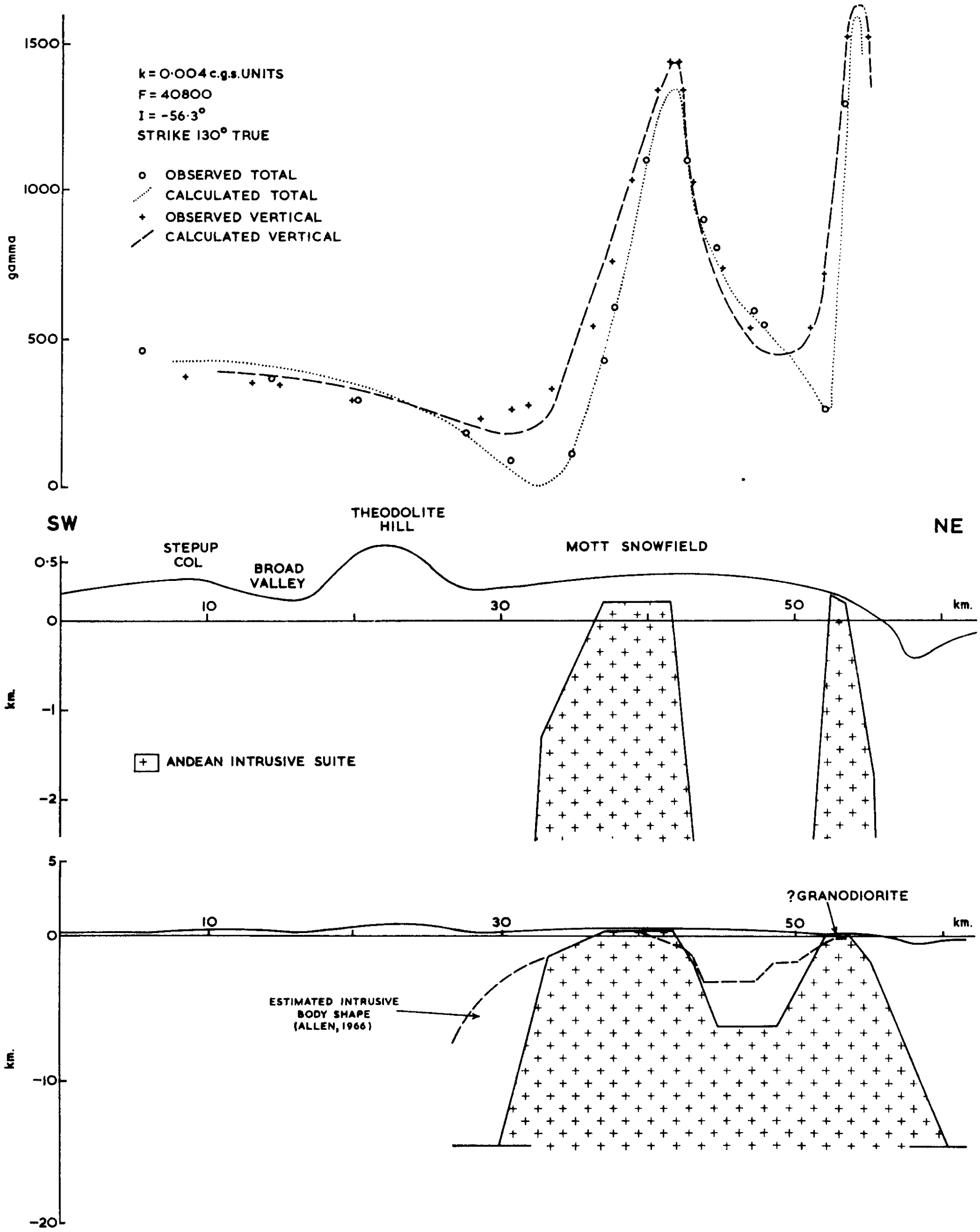


FIGURE 14
Interpretation across the Mott Snowfield total-field magnetic anomaly showing the comparison with the calculated body shape estimated by Allen (1966*a*) from the vertical-field anomalies.

but he also gave a more specific depth of 250 m. The base of the intrusive body is taken to exceed 12 km. on the basis of the likely magnetization and anomaly magnitude.

b. *Summit Pass.* Profiles A and B of Ashley (1962) traverse the anomaly considered in Fig. 13. This section has again been governed by the distribution of the total-field stations. Outcrops of diorite at Summit Ridge have a measured intensity of remanent magnetization ranging from 901×10^{-6} to $3{,}410 \times 10^{-6}$ c.g.s. units (Ashley, 1962), which together with the induced magnetization give an equivalent susceptibility of the order of 0·01 c.g.s. units. In his original interpretation, Ashley (1962) calculated the mean values for all the sampled Tabarin Peninsula diorites. This gave a resultant intensity of magnetization of $(2{,}457 \pm 521) \times 10^{-6}$ c.g.s. units with an equivalent susceptibility of $(5{,}680 \pm 1{,}200) \times 10^{-6}$ c.g.s. units.

The profile in Fig. 15 illustrates not dissimilar total and vertical magnetic fields, showing two main anomalous areas, the one centred over Last Hill and the other over Summit Pass. As with the Mott Snowfield anomalies, a two-dimensional computer method was used with various combinations of body shapes and susceptibilities. The solution considered as giving the most satisfactory fit is shown superimposed on the body as calculated by Ashley (1962). Although it exhibits some differences in form and in the susceptibility selected, both of the estimates indicate an intrusive mass of a laccolithic nature approaching the

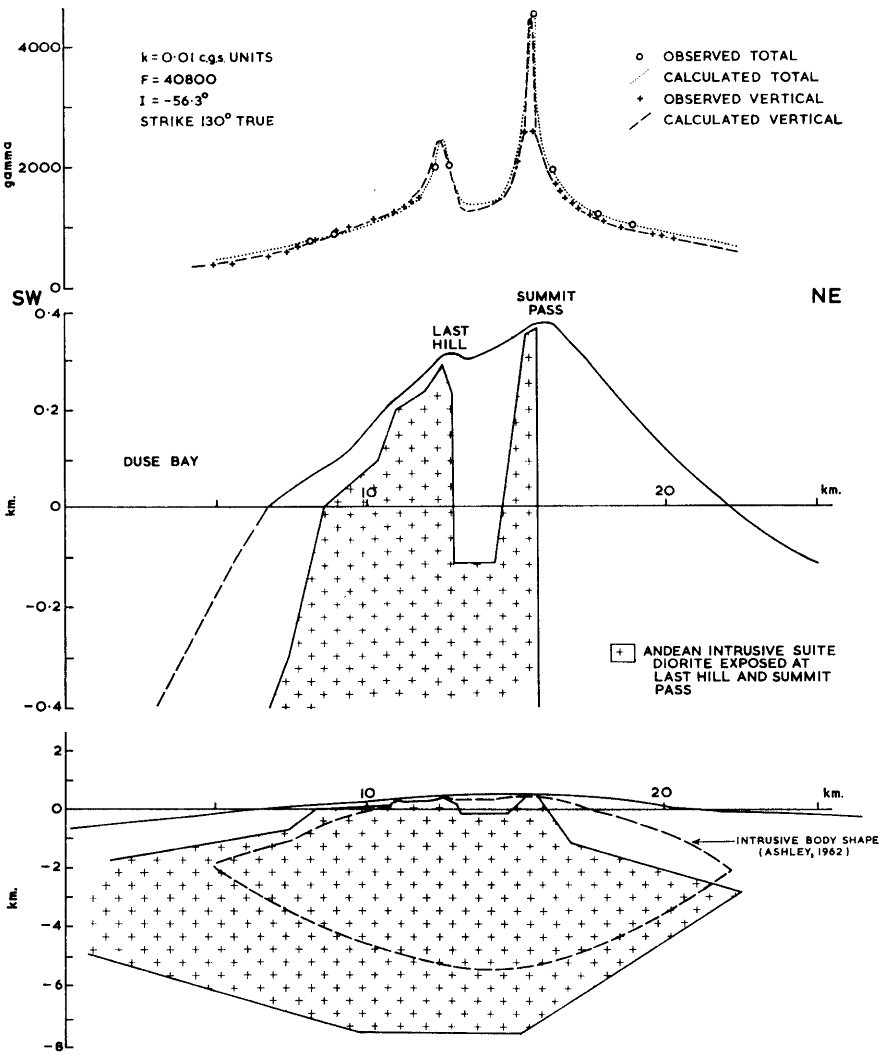


FIGURE 15

Interpretation across the Summit Pass total-field magnetic anomaly showing the comparison with the body shape estimated by Ashley (1962) from the vertical-field anomalies.

surface at Last Hill and Summit Pass. The present solution indicates that in the vicinity of the Last Hill diorite exposure the intrusive body must lie within 20–30 m. of the surface over a distance of about 2 km. At Summit Pass, where the body is not exposed, it must lie within 10 m. of the surface for a distance of about 0·5 km.

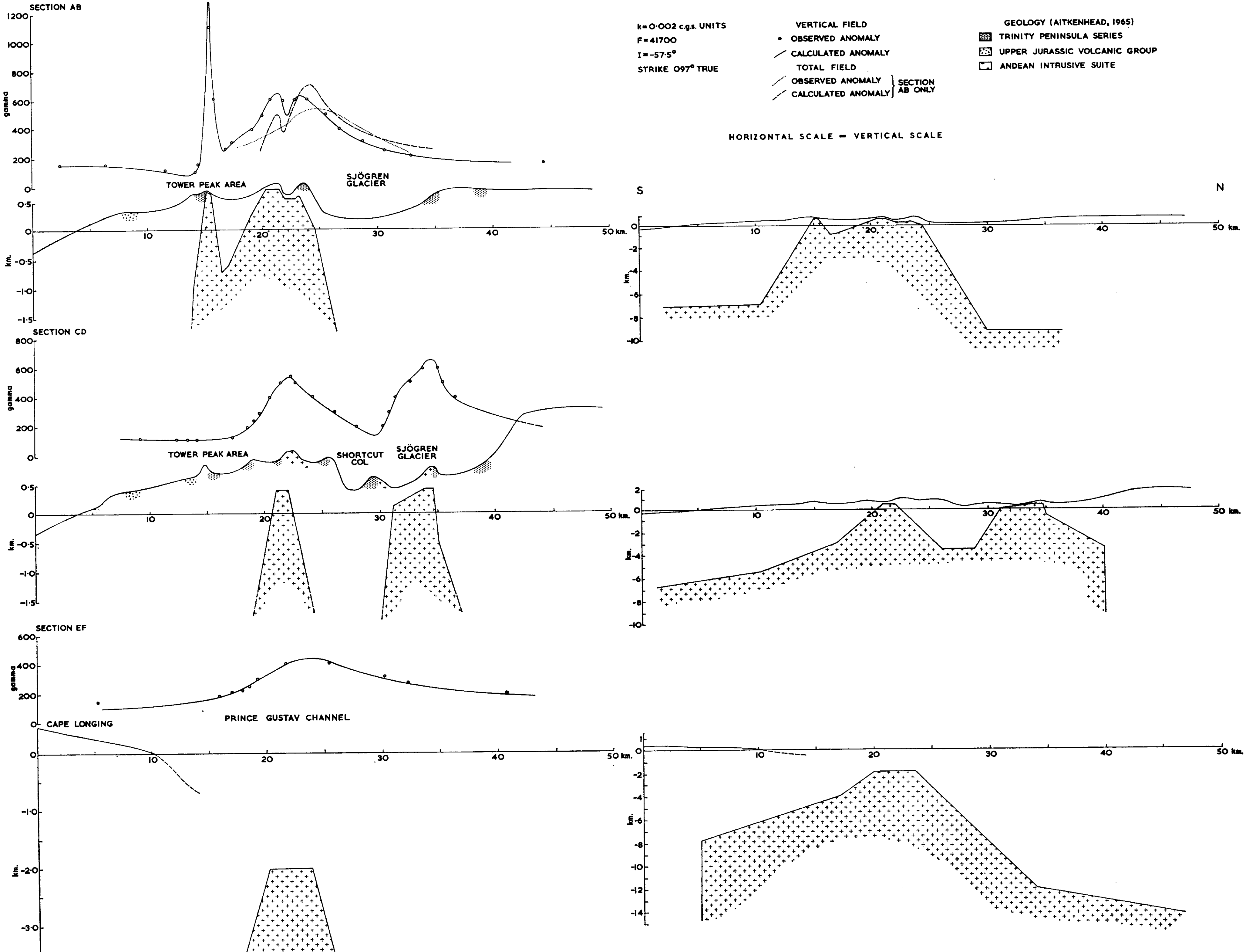
c. *Sjögren Glacier*. The vertical-field survey (Fig. 16) between Pitt Point and Cape Longing is a southerly extension of the programme initiated by Ashley (1962) and Allen (1966*a*). The survey included traverses over Mount Reece, 7 km. west of Pitt Point, and over the Sjögren Glacier area, 20 km. north of Longing Col. Close to the granite exposures of Mount Reece and the associated metamorphic aureole, the vertical field fails to show any significant anomalous trends. Samples of the granite have been analysed by Allen (1966*a*) and they show extremely low magnetizations compared with many other Andean intrusive rocks. Remanent magnetizations of $0–24 \times 10^{-6}$ c.g.s. units were measured with corresponding volume susceptibilities of $6–26 \times 10^{-6}$ c.g.s. units. These are of the same order as intensities of magnetization for the sediments of the Trinity Peninsula Series, and therefore they are unlikely to produce any measurable field anomalies. The exposures of Andean intrusive rocks in the Sjögren Glacier area consist of granite, variable granite and quartz-diorite. No measurements of magnetic intensities have been carried out on the Andean intrusive rocks exposed in this area but with susceptibilities of $18–25 \times 10^{-6}$ c.g.s. units (Table III) for the nearby Trinity Peninsula Series it is unlikely that the exposed acid intrusive rocks could account for the observed anomaly.

The Sjögren Glacier anomaly, as revealed by the vertical-field survey, trends east–west for at least 35 km. and at its broadest it is about 20 km. The eastern edge of the anomaly terminates in Prince Gustav Channel, while the western limit has not yet been determined, but it may possibly extend to the glacier head where Andean intrusive rocks are again exposed. Because the latter are also mainly of granite (Aitkenhead, 1965), they may not exhibit any appreciable anomaly due to low intensities of magnetization.

Interpretation of the observed anomaly was along three transverse profiles (Fig. 13). Trial shapes, based initially on the geological knowledge of the area, were computed using varying susceptibility values. Those bodies giving the closest fit to the observed vertical profiles are shown in Fig. 17. Section A–B also illustrates a comparison between the calculated and observed total magnetic field anomalies, the latter being interpolated from a limited number of total-field stations occupied in the vicinity. The equivalent susceptibility used in the final calculation and which was assumed to be uniform over the body was 0.002 c.g.s. units; this is within a range estimated for acid intrusive bodies of the Andean Intrusive Suite.

Superimposed on the main anomaly are two relatively high anomalous areas, one 3 km. north of Tower Peak (lat. 64° 22′ S., long. 59° 07′ W.) (Fig. 16, inset B) which reaches 900 gamma, and the other of 1,100 gamma (Fig. 16, inset A) situated 5 km. north-west of Jedd trigonometrical station. These have been interpreted as areas where more basic members of the Andean Intrusive Suite approach the surface. Section A–B passes over the anomaly near Tower Peak and, although the Trinity Peninsula Series alone is exposed close to the anomaly "high", the intrusive body has been estimated to lie within a few metres of the surface. An alternative solution could be that the body producing this anomaly is associated with the Upper Jurassic volcanic rocks which are known to occur on, and to the south and west of, Tower Peak. On the southern side of Sjögren Glacier and 8 km. to the north of Tower Peak, the intrusive body approaches to within 20–100 m. of the surface, though it is still outside the metamorphic aureole of the granite outcrop 6 km. to the west. No magnetic readings were taken directly over the granite exposures but several were obtained within a 2 km. radius of it. Depth estimates from these (Fig. 17, section C–D) place the body within 500–600 m. of the surface, necessitating a dip of around 25° from the exposures. At the western end of the glacier, and within the metamorphic aureole postulated by Aitkenhead, the intrusive body lies at a calculated depth of 300–400 m. below the glacier surface. It is exposed, together with the Trinity Peninsula Series, at a short distance to the west as an isolated nunatak in the glacier. The anomaly at the eastern end gradually fades beneath Prince Gustav Channel and this is indicative, provided the susceptibility has remained constant, of a body lying at a gradually increasing depth. Section E–F gives an estimated depth of 2,000 m. to the upper surface, the overlying strata consisting of an unknown thickness of ice, water and non-magnetic rock. A two-dimensional gravity investigation over the relevant part of Prince Gustav Channel (p. 70) indicates the depth of the sea floor is of the order of 800 m.

d. *Poynter Hill.* Vertical- and total-field magnetic surveys have been completed over this area (Fig. 18) and they emphasize the similarity between such components at high magnetic latitudes. The main anomaly can be traced south from Notter Point for 25 km., reaching a maximum of 1,180 gamma at a point 7 km. west of Poynter Hill. Inland, towards the Aureole Hills, the anomaly decreases in magnitude as was found in north-eastern Trinity Peninsula where the coastal "highs" give way towards the plateau to areas showing only small negative magnetic gradients. Elliot (1965) has described the geology of this area, with Upper Jurassic volcanic rocks, the Trinity Peninsula Series and the Andean Intrusive Suite forming the majority of the exposures. The volcanic group is exposed 6 km. south-south-west of Poynter Hill, where it is represented by several beds of lithic crystal tuffs, 0·9–1·2 m. thick dipping steeply to the east. Other exposures occur a further 7 km. to the south-west. Total and vertical magnetic field traverses cross the area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gamma
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F=41700
I=-57·5°
STRIKE 097° TRUE
VERTICAL FIELD
OBSERVED ANOMALY
CALCULATED ANOMALY
TOTAL FIELD
OBSERVED ANOMALY
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SECTION AB ONLY
GEOLOGY (AITKENHEAD, 1965)
TRINITY PENINSULA SERIES
UPPER JURASSIC VOLCANIC GROUP
ANDEAN INTRUSIVE SUITE
HORIZONTAL SCALE = VERTICAL SCALE
TOWER PEAK AREA
SJÖGREN GLACIER
SHORTCUT COL
CAPE LONGING
PRINCE GUSTAV CHANNEL
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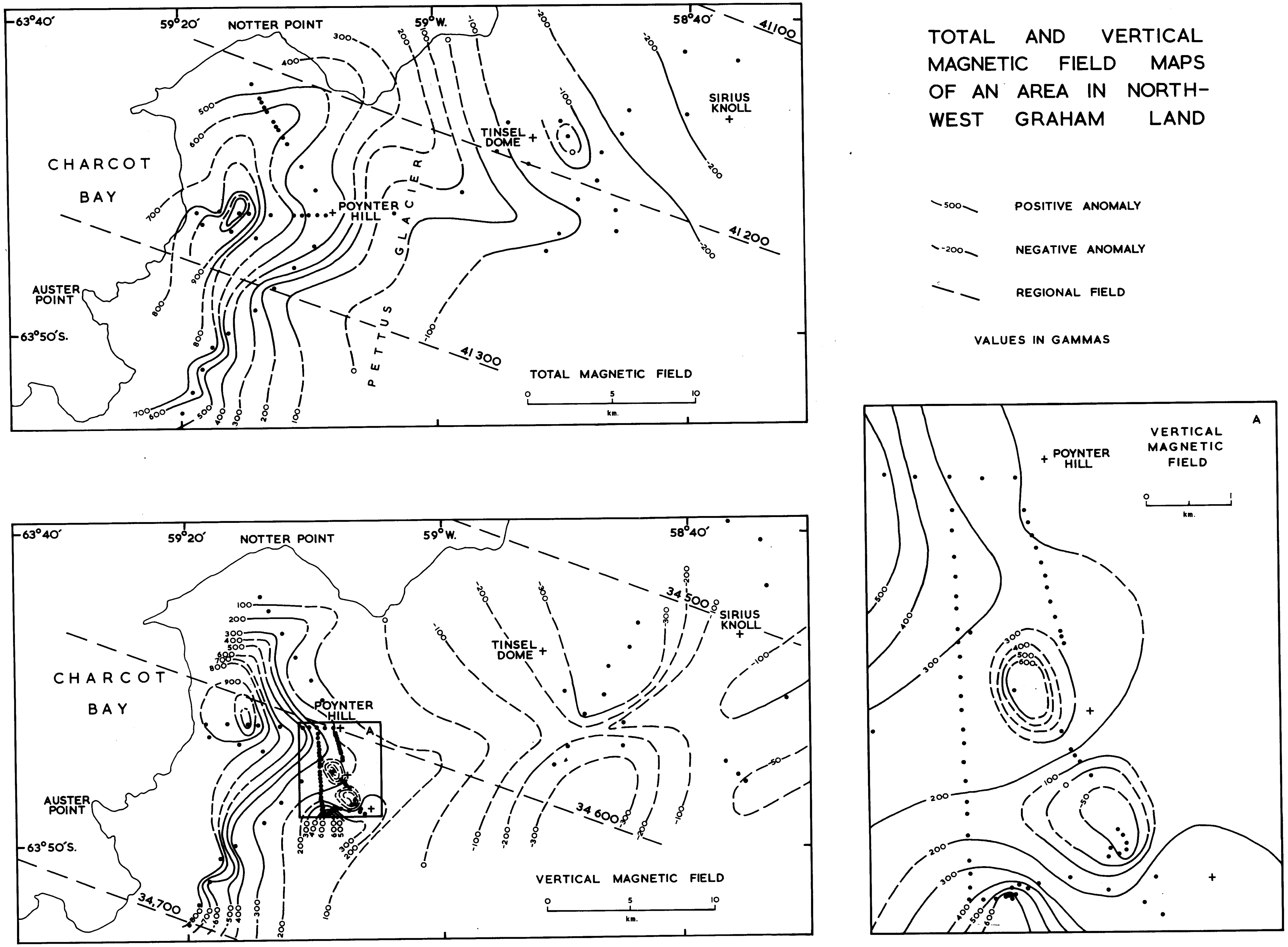

FIGURE 18

Vertical and total-field magnetic anomalies over Poynter Hill.

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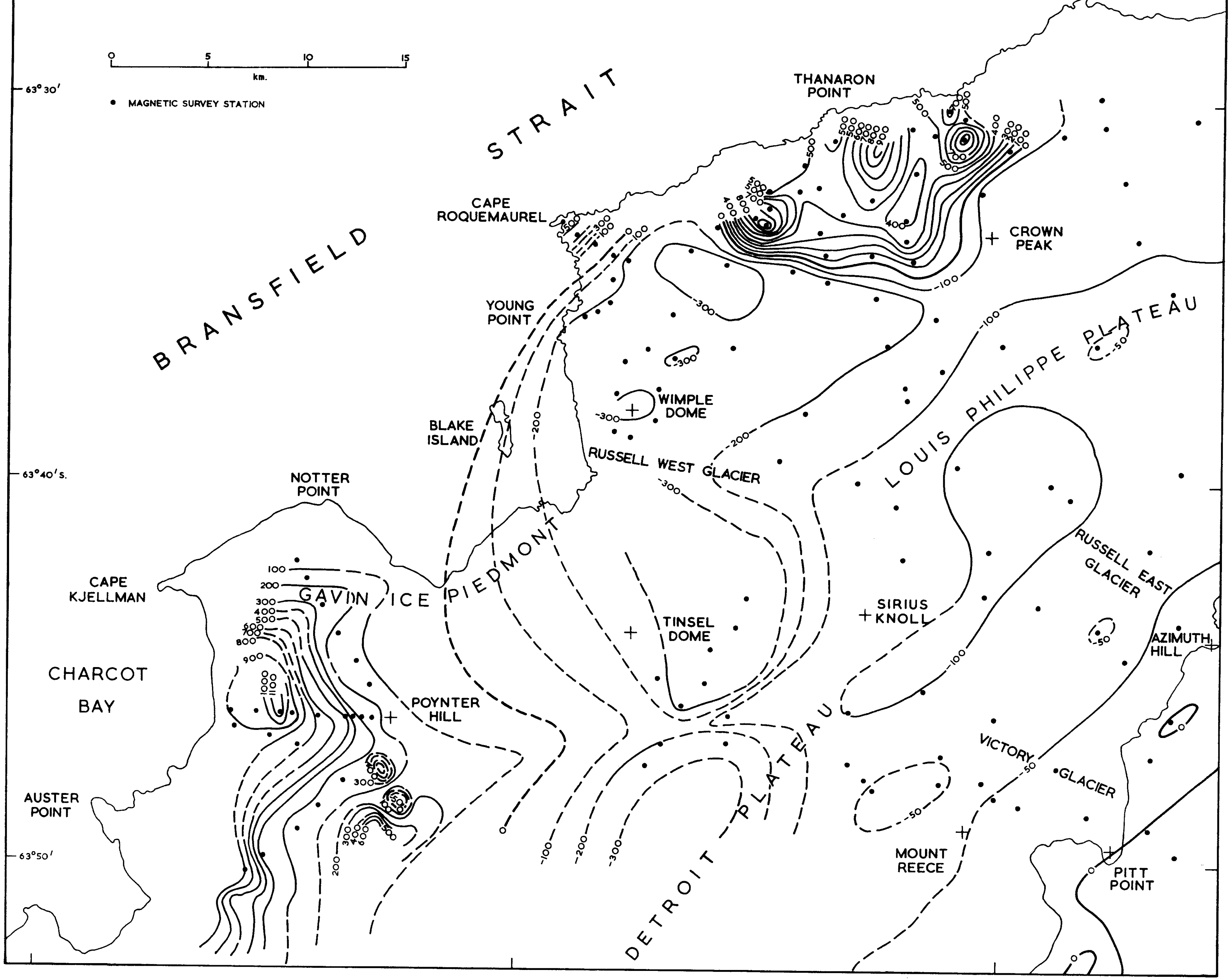

FIGURE 19

Sketch map to illustrate the relationship between the Poynter Hill anomaly and part of the previous vertical-field survey over Trinity Peninsula (Allen, 1966*a*).

of known Upper Jurassic volcanic rocks, reaching a maximum recorded value of + 830 gamma. Tuffs belonging to the Upper Jurassic Volcanic Group in other areas have been measured by Allen (1966*a*), and they give a remanent intensity of magnetization of $1{,}893 \times 10^{-6}$ c.g.s. units and a corresponding equivalent susceptibility of approximately 0·007 c.g.s. units. Rocks exhibiting such a magnetic susceptibility could account for the observed anomaly on the south-west traverse from Poynter Hill. Insufficient data coverage of the anomaly over the volcanic rocks has thus far precluded further interpretation. Fig. 19 shows the Poynter Hill vertical-field anomaly in relation to the vertical-field magnetic survey completed to the north and east (Allen, 1966*a*). Between Thanaron Point, where there is an exposure of quartz-gabbro, and Crown Peak, Allen estimated that the intrusive body approaches to within 0·5 km. of the surface.

Apart from the exposures to the south-west, no further outcrops have been recorded by overland parties west of Poynter Hill. However, in January 1962 (personal communication from N. Aitkenhead) parties from R.R.S. *Shackleton* made several landings on the coast between Thanaron Point and Cape Kater and made the following observations. At Blake Island, metamorphosed greywackes and shales were found exhibiting a northerly dip of 40–60°. On the coast 3 km. north of Auster Point there is a massive, highly metamorphosed, apparently structureless conglomerate, with numerous acid and intermediate dykes. Massive medium-grained acid intrusives were observed at Cape Kater (lat. 63° 46′ S., long. 59° 54′ W.), while at Cape Kjellman an offshore sighting of exposures suggested the presence of greywackes.

On the basis of present knowledge and until such time as additional data are available, it has been assumed that the Poynter Hill anomaly is the result of rocks belonging to the Andean Intrusive Suite. With an estimated susceptibility of 0·006 c.g.s. units and a direction of magnetization along that of the Earth's present field direction, various body forms were considered, that most closely approximating to the selected profile (Fig. 13) shown in Fig. 20. The anomaly fit in the Poynter Hill area could be improved with additional field data but the present observed profile does not warrant a more detailed analysis. The magnetic high has been interpreted as due to a body lying within 30 m. of the surface, whereas further towards the coast the upper surface is estimated to lie at greater depths of 200–300 m. No magnetic data are available for this area so its extension in this direction cannot be assessed.

Allen commented that, because of the marked change in the magnetic intensity of the rocks between Wimpole Dome and Cape Roquemaurel, no interpretation was attempted. A similar situation is indicated for the rocks of the Poynter Hill area and those, again of the Andean Intrusive Suite, to the east near the Aureole Hills. Elliot (1965) has published modal analyses of several Andean intrusive rocks from the Poynter Hill–Aureole Hills area. A granodiorite cropping out 2 km. south of Poynter Hill has 1·7 per cent of iron ore, which if present as magnetite could give rise to a susceptibility of about 0·004 c.g.s. units (Mooney and Bleifuss, 1953). However, analysed samples of tonalite and adamellite collected from the main part of the intrusive body east of Poynter Hill reveal that, although iron ore is present, it is not in sufficient quantity to warrant recording in the modal analysis. The Poynter Hill anomaly may be attributed to a body of granodiorite composition as exposed nearby, but this would appear to give way eastwards to more acid differentiates of the Andean Intrusive Suite.

e. *Summary*. The total and vertical magnetic surveys over Trinity Peninsula have been described and the main anomalous areas interpreted. In the vicinities of Mount Bransfield, Summit Pass and Poynter Hill both components have been considered but over the remaining areas the vertical field alone has been measured and has provided the data for profile interpretation. The intensities of magnetization associated with the Trinity Peninsula Series and the Upper Jurassic volcanic rocks are not in general considered large enough to account for the major anomalies, though some of the Upper Jurassic volcanic rocks may have significant magnetic intensities, e.g. south-west of Poynter Hill, and also thermo-remanent magnetization could possibly produce a measurable magnetization. However, the effect of any thermo-remanent magnetization imparted to the Trinity Peninsula Series by the intrusive rocks has been neglected on the basis that thicknesses involved would only be insignificant. Ashley (1962) calculated that thermo-remanent magnetization of the sediments may contribute up to 91 m. from the margin of the intrusive body in the vicinity of Mineral Hill.

The geophysical evidence has in some cases supported the geological findings. For example, in the north-east of Trinity Peninsula, where the approximate southern limit of the metamorphic aureole (Aitkenhead, 1965) lies just to the south of a line between Fidase Peak and Camel Nunataks. Several kilometres to the north of this line the magnetic data suggest that the body lies within 350 m. of the surface and dips steeply

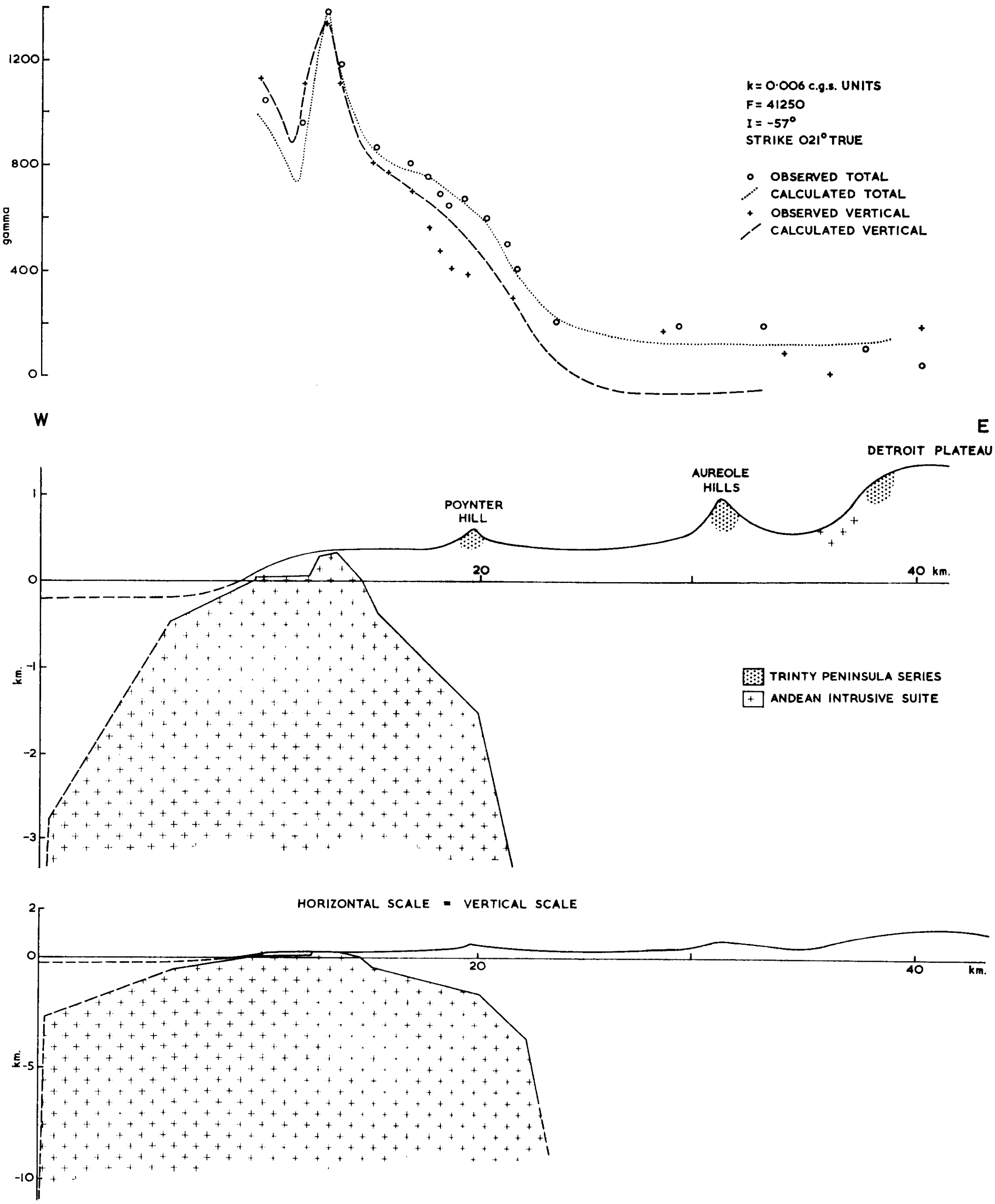

FIGURE 20

Interpretation across the Poynter Hill vertical-field anomaly, showing the calculated sub-surface distribution of the anomaly source rock.

to the south as the limit of the aureole is approached, thus leaving the sediments unaffected in this direction. Over other plutonic bodies the general sub-surface distribution has been estimated and this agrees closely with the limits of the known thermal aureoles.

The Andean Intrusive Suite in this area exhibits a wide range of compositions varying through acid to intermediate representatives. Because of this, some areas covered by the magnetic survey where no anomaly was recorded may nevertheless be underlain by plutonic rocks of a more acidic composition. Throughout

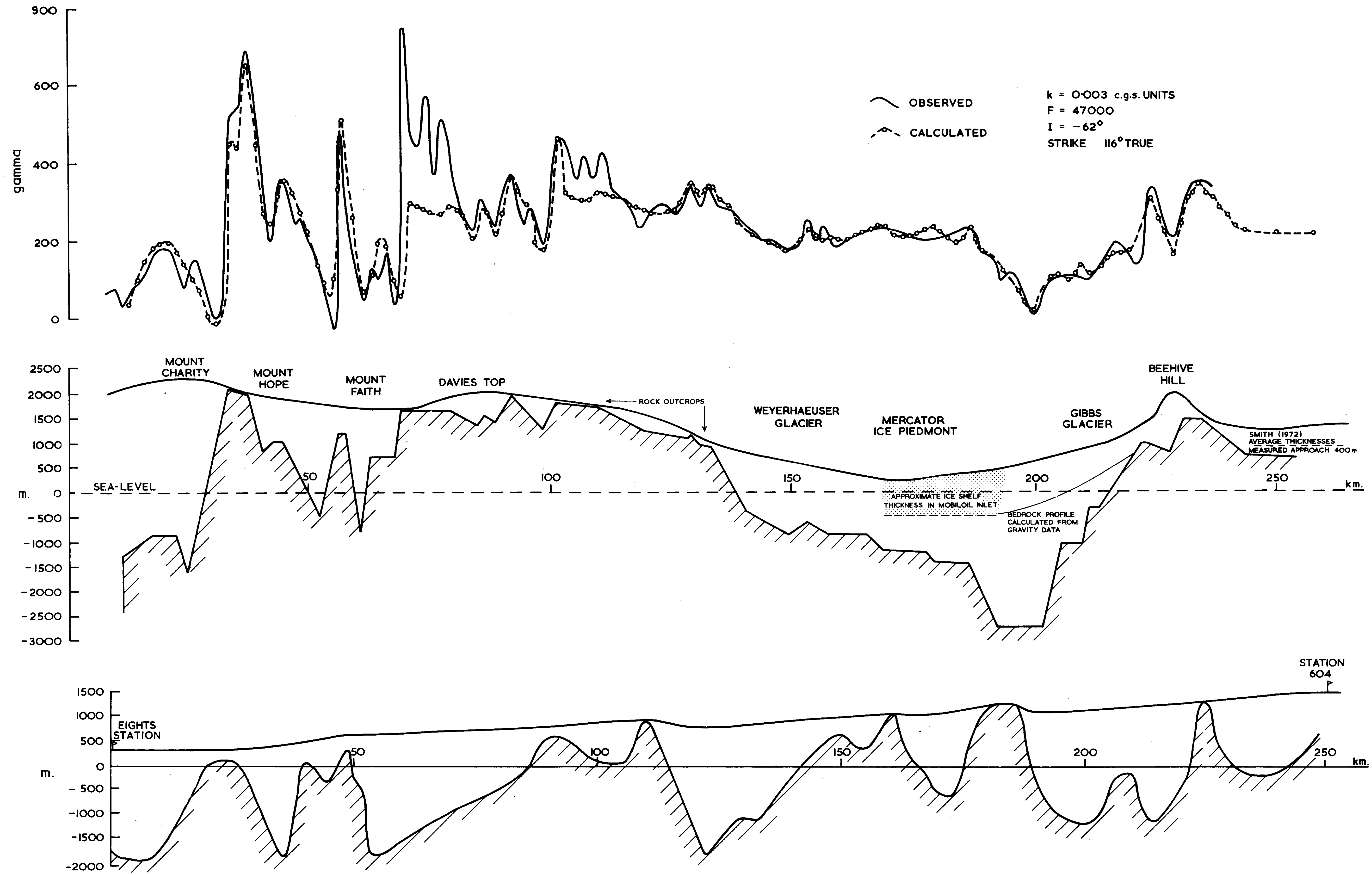

FIGURE 21

Interpretation of the bedrock topography beneath the eastern land magnetic traverse compared with a section (Behrendt, 1964) calculated from seismic reflection and gravity measurements at the southern end of the Antarctic Peninsula.

the interpretations, no allowance was made for variable susceptibilities within any given body, and the magnetizations were all considered as being induced in the direction of the Earth's present magnetic field.

In north-eastern Trinity Peninsula, the interpretation is consistent with the opinion that the area is underlain by an extensive batholith, its presence being observed at least as far south as Fidase Peaks and Camel Nunataks. South of these localities little magnetic activity is observed until a further 50–60 km. down the length of the peninsula. It would thus appear that this north-eastern batholith is isolated from those to the south unless a non-magnetic representative is present, or else it is connected at depths greater than about 10 km. Studies on the Trinity Peninsula Series indicate appreciable thicknesses of sediments in this area: 3,660 m. in the Mount d'Urville area (Elliot, 1965), 1,500 m. to the south of Broad Valley, at least 4,500 m. at Mount Wild (lat. 64° 12′ S., long. 58° 50′ W.), and from Long Island to the edge of the Louis Philippe Plateau there is an estimated minimum of 12,000 m. (Aitkenhead, 1965).

On the southern side of Victory Glacier the Mount Reece granite crops out; it is non-magnetic as is that to the east of Tufft Nunatak (lat. 63° 55′ S., long. 58° 42′ W.). At the same latitude but on the western side of the Detroit Plateau, the non-magnetic granite is exposed in the Aureole Hills. It is suggested that these outcrops form part of the same batholith which has a radius of about 8 km. Both of these areas have associated banded hornfelses, and the metamorphic aureole proposed by Aitkenhead (1965) for the eastern side, if extrapolated, would encompass the granite near the Aureole Hills. Lying 16 km. to the north of the Aureole Hills, the granite outcrop at Wimpole Dome also gives a negligible magnetic field. Within the area of Young Point a granite with a volume susceptibility of $4{,}784 \times 10^{-6}$ c.g.s. units has been measured, and the quartz-gabbro at Thanaron Point gave a resultant intensity of magnetization of $3{,}130 \times 10^{-6}$ c.g.s. units. Such is the non-uniform nature of the plutonic bodies in this area that the sub-surface extent of the Andean Intrusive Suite cannot be estimated by magnetic means. The Thanaron Point and Poynter Hill anomalies have been interpreted using similar susceptibilities, yet each has a more acid differentiate to the east. The lack of magnetic data between these areas prevents further comparison, though a magnetic trough reaching a low of −300 gamma may join the Wimpole Dome and Aureole Hills areas and indicate some geological continuity at depth. Overlying Trinity Peninsula Series sediments may be the cause of some apparently non-magnetic areas. To the south of Poynter Hill and the Aureole Hills, on the eastern side of Pettus Glacier, Elliot (1965) has mapped at least 3,660 m. of these sediments. It should be noted that a total-field magnetic traverse was completed off Thanaron Point in 1961 by the marine group from the Sub-Department of Geophysics, University of Birmingham. This traverse suggested that the magnetic anomaly traceable between Cape Roquemaurel, Crown Peak and Thanaron Point also extends 6 km. to the north-west into Bransfield Strait and is a probable extension of the quartz-gabbro intrusive body noted by Allen (1966*a*). Extensive plutonic activity is indicated in the Sjögren Glacier area with severe thermal metamorphism of the Trinity Peninsula Series sediments. The westerly extent of this intrusion is unknown because it disappears beneath the Detroit Plateau which has not yet been covered by the magnetic survey.

Widespread thermal metamorphism (Aitkenhead, 1965) between Marescot Ridge and Russell West Glacier indicates an extensive underlying plutonic body. An outcrop in the Aureole Hills (Elliot, 1965) shows complete assimilation of the country rock. Low-grade regional metamorphism increases from north-east to south-west and further examples of thermal metamorphism around plutonic bodies have been described by Elliot (1965) and Aitkenhead (1965). Both the geological and geophysical evidence supports a theory that Trinity Peninsula is underlain by a system or systems of plutonic bodies belonging to the Andean Intrusive Suite.

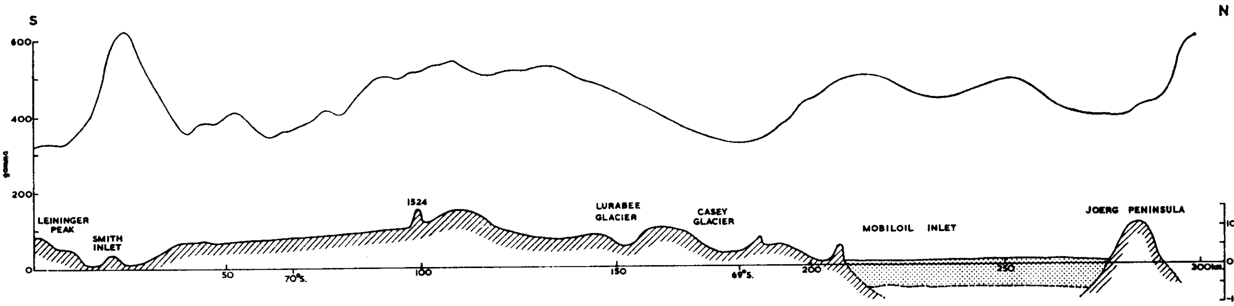

FIGURE 22

A section of the Project Magnet aeromagnetic traverse (flight track 721S) south of lat. 68° S.

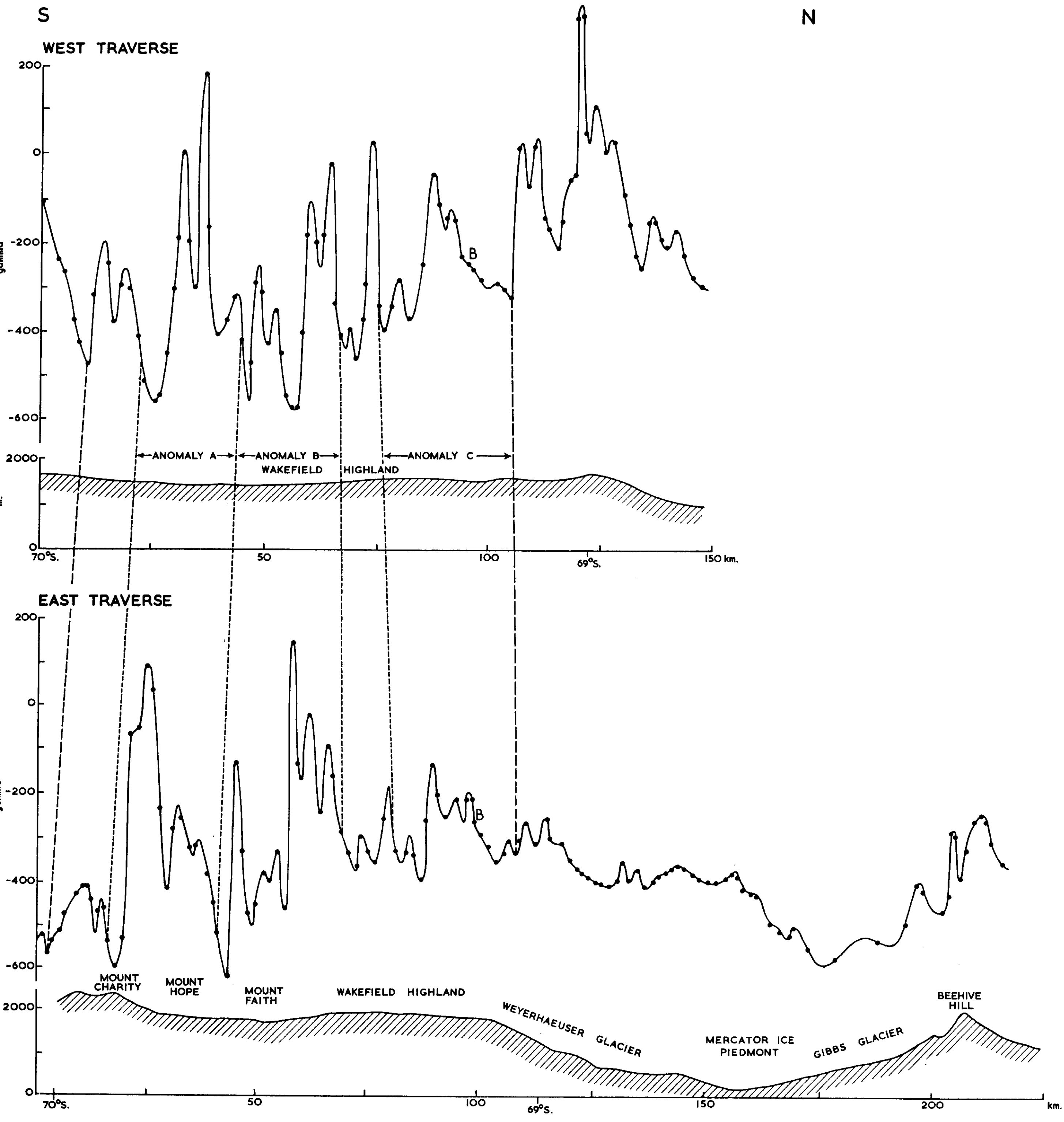

FIGURE 23

Longitudinal profiles of the west and east southern plateau land magnetic traverses demonstrating possible cross-correlation. Locality B is a common point.

2. The southern plateau magnetic traverses

The southern plateau total-field magnetic measurements were observed by J. Ross during the 1966–67 summer field season. Using the observatory data from the Argentine Islands, the measurements were corrected for diurnal variation and reduced to September 1963 to allow for the secular change. Regional field values, obtained by extrapolation of the regional gradient north of lat. 68° S., were removed and the residual profiles (Fig. 23) were drawn to include the results obtained from the earlier traverses of Kennett (1966*a*, *b*) and the author (Fig. 59). In addition to the land survey, part of the Project Magnet aeromagnetic programme (p. 48) traversed the eastern plateau area (Fig. 22).

With neither susceptibilities nor ice-thickness values available, it was decided to interpret the magnetic anomalies on the assumption that their source lay in the variation of the upper surface of the subglacial bedrock. This would then give a comparison with a bedrock profile calculated by Behrendt (1964) 550 km. to the south at the base of the Antarctic Peninsula. Due to the lack of information on the magnetic intensities of the country rocks, a trial-and-error approach was used. The two-dimensional computed rock profile giving the closest agreement between observed and calculated magnetic fields along the eastern traverse is shown in Fig. 21 and illustrates the calculated bedrock topography along 250 km. of traverse. Also in Fig. 21 is a section of the profile calculated by Behrendt (1964) which shows similarities in ice thicknesses, bedrock gradients and susceptibility values. This could imply the presence of rock types of a similar magnetic nature and which probably can be related to the igneous rocks of the Andean Intrusive Suite. Along those parts of the traverse known to be close to rock outcrops there is a marked increase in the amplitudes of anomalies, which supports the conclusions reached from the interpretation. Examples illustrating this include the Mounts Faith, Hope and Charity section, the Davies Top area and the exposures along the eastern side of Weyerhaeuser Glacier, beneath Norwood Scarp.

As expected, some of the anomalies could not be fitted satisfactorily using the average estimated susceptibility of 0·003 c.g.s. units and these must therefore be attributed to rocks of a different magnetic and hence geological character. (Since the interpretation, 123 susceptibility measurements (Table III) have been carried out on igneous and volcanic rocks collected from 19 different localities along the west coast of Palmer Land between lat. 70° 10′ and 70° 42′ S. The resulting mean volume susceptibility of $(2{,}758 \pm 1{,}688) \times 10^{-6}$ c.g.s. units closely approximates the 0·003 c.g.s. units used in the interpretation.) Elsewhere in Graham Land, the Andean Intrusive Suite is represented by a wide range of igneous rocks. Larger anomalies have also been found in the land traverses over the Hub Nunatak, Godfrey Upland and Mount Timosthenes areas, but until further geological material is available it is not proposed to attempt to elucidate the source bodies producing such anomalies.

Radio echo-sounding profiles have recently been published (Smith, 1972) for part of the area covered by the southern plateau magnetic traverses. The flight profiles of consequence are south of lat. 69° 30′ S. and between long. 64° and 66° W. The echo-sounding profiles are obtuse to the lines of total-field magnetic readings but they still serve to show the very irregular nature of the upper surface of the bedrock. In the vicinity of the magnetic profiles, ice thicknesses range from 125 to 1,149 m. Only isolated measurements are available along the eastern land traverse itself but again ice-thickness extremes of 371 and 1,141 m. have been recorded. Over one 4 km. section near Mount Charity, the bedrock topography undergoes a change of 720 m. Due to uncertainties in navigation and also because of the irregular subglacial relief, it is not possible to correlate the flight profile thicknesses with those interpreted from the magnetic traverses. As mentioned previously, both of these depict a rugged bedrock topography and, although the maximum thicknesses are different, this could either be due to offset areas or to the presence of a non-magnetic layer between the ice and magnetic "basement", which would remain undetected during field measurements.

South of lat. 69° S., after converging at point B (Fig. 23), the two land traverses run almost parallel for 35 km. and within 5 km. of each other before they diverge to reach a maximum separation of 40 km. at lat. 70° S. Examination of the magnetic profiles along each of the traverses suggests a possible correlation. This correlation, exemplified by anomalies A, B and C (Fig. 23), showed an east-north-east to west-south-west trend in the northern area and this gradually changes to a north-west to south-east direction. The possible eastward extension of the broader magnetic trends to the Project Magnet aeromagnetic profile was next investigated. One of the difficulties with this section of the aeromagnetic traverse was the flight altitude of 2,750 m. This caused a considerable decrease in the resolution of individual anomalies and thus tended to reflect only magnetic variations of a deeper nature, or of larger structural disconformities at shallower depths. To overcome this, an upward continuation method was applied to the ground-survey data in which the higher-frequency anomalies were filtered using a digital filter with a frequency response which corresponded to the amplitude spectrum of the Project Magnet traverse. By so doing, the near-surface anomalies could be largely eliminated, leaving only those due to deeper sources.

The aeromagnetic profile was thus subjected to a Fourier transform, the resulting amplitude spectrum showing a dominant sinusoidal wave peaking at 3 Hz/320 km. (A 320 km. unit was selected for convenience since no traverse lengths investigated exceeded this value.) At 15 Hz/320 km. the amplitude approached the zero level, indicating that no anomalies of wave-length greater than 21 km. occurred over this section. Digital filters whose frequency characteristics approximated to those of the amplitude spectrum were then constructed to a formula given by Robinson (1967, p. 193) and then applied to the land data.

Because of probable navigation errors in the flight line and possible changes in magnetic trend direction between the land and air traverses, only a 256 km. length of the Project Magnet profile was selected for investigation. Only the eastern land traverse was used but it was extended north to almost lat. 68° S. to include the magnetic "low" corresponding to the topographical feature of Gibbs Glacier–Neny Trough. It was considered that this feature may be reflected in the Project Magnet traverse across Mobiloil Inlet.

After filtering, the two profiles were plotted alongside one another as illustrated in Fig. 24. The southern

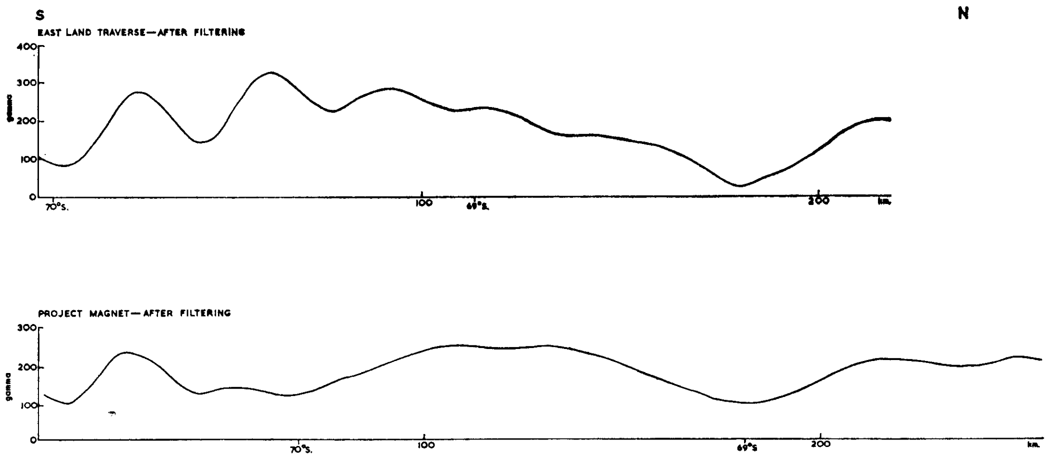

FIGURE 24

Comparison, after filtering, between the eastern land traverse and the corresponding section of the Project Magnet aeromagnetic traverse (flight track 721S).

part of each has individual anomalies superimposed on a relatively undisturbed regional background. Possible correlation of the anomaly over Smith Inlet with those of the land traverse could suggest a continuation of the north-west to south-east trend through Mounts Faith, Hope and Charity, but this must be viewed with some reservation until further data are available. However, it is of interest as the area is distinct from a central zone where the profiles are relatively magnetically quiet apart from a broad, low-amplitude magnetic "high". Over these sections, the land traverse passes over the ice-covered Wakefield Highland, whereas the aeromagnetic traverse crosses a series of glaciers and ridges trending perpendicular to the flight line. The northern area could again be interpreted as exhibiting a north-west to south-east trend, which is exemplified on the land traverse by the structure of Neny Trough. A more detailed analysis of this is discussed on p. 88. The aeromagnetic profile indicates an area of "low" magnetic intensity coincident with Casey Inlet. This may, of course, be due to the geology of the inlet area, but should it be re-located slightly to the north then there would be a very similar magnetic profile to that of Gibbs Glacier. This would then favour a continuation of the structure towards Poseidon Pass. The magnetic profile over Mobiloil Inlet is magnetically quiet, and typical of that over the greater part of the Larsen Ice Shelf.

The true significance of these correlations cannot yet be realized and it would be unwise to speculate much further in the light of the evidence at hand. It does appear, however, that there is some geophysical justification for postulating that the Neny Trough structure could well continue to the south-east towards Poseidon Pass. In addition, some form of geological or structural boundary appears to run north-west to south-east slightly to the north of Mount Faith, and this separates a relatively quiet magnetic area to the north from a more disturbed one to the south.

3. *Project Magnet aeromagnetic traverses*

The Project Magnet aeromagnetic traverses (Bregman and Frakes, 1970) were initiated by the U.S. Naval Oceanographic Office. Tracks 720 and 721 (Fig. 25) were flown over the Antarctic Peninsula and

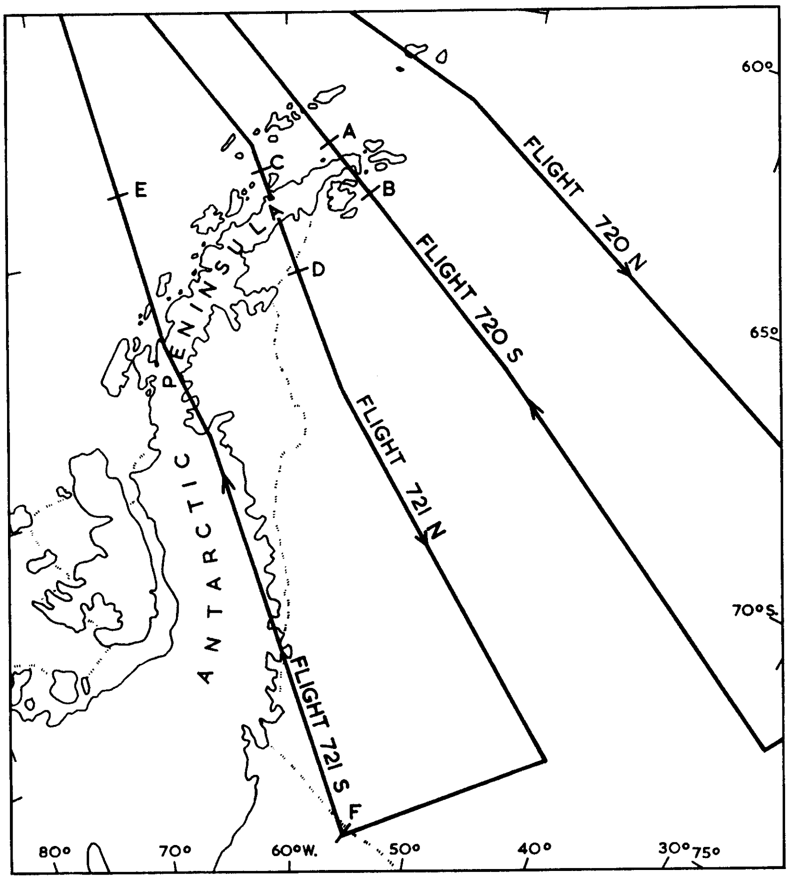

FIGURE 25

Map showing the distribution of the Project Magnet aeromagnetic traverses over Graham Land.

Weddell Sea in February 1965 but, because of their 200–300 km. separation, only major anomalous trends were revealed.

The flight tracks have each been subdivided into north and south segments, and over the peninsula they run between the following localities:

i. *Flight track 720S*. Beak Island (lat. 63° 37′ S., long. 57° 20′ W.) to Cape Legoupil (lat. 63° 20′ S., long. 57° 53′ W.).

ii. *Flight track 721N*. Seal Nunataks (lat. 65° 03′ S., long. 60° 18′ W.) to Cape Herschel (lat. 64° 04′ S., long. 61° 03′ W.).

iii. *Flight track 721S*. New Bedford Inlet (lat. 73° 22′ S., long. 61° 15′ W.) to Lavoisier Island (lat. 66° 12′ S., long. 66° 44′ W.).

These sections, totalling 1,000 km., were corrected for the regional gradient and are included in the profiles AB, CD and EF in Fig. 26. No corrections have been applied for diurnal variation but reference to the magnetograms from the magnetic observatory at the Argentine Islands for the period of the flights indicates that any variations in the magnetic field were never greater than 5 gamma. The traverses have

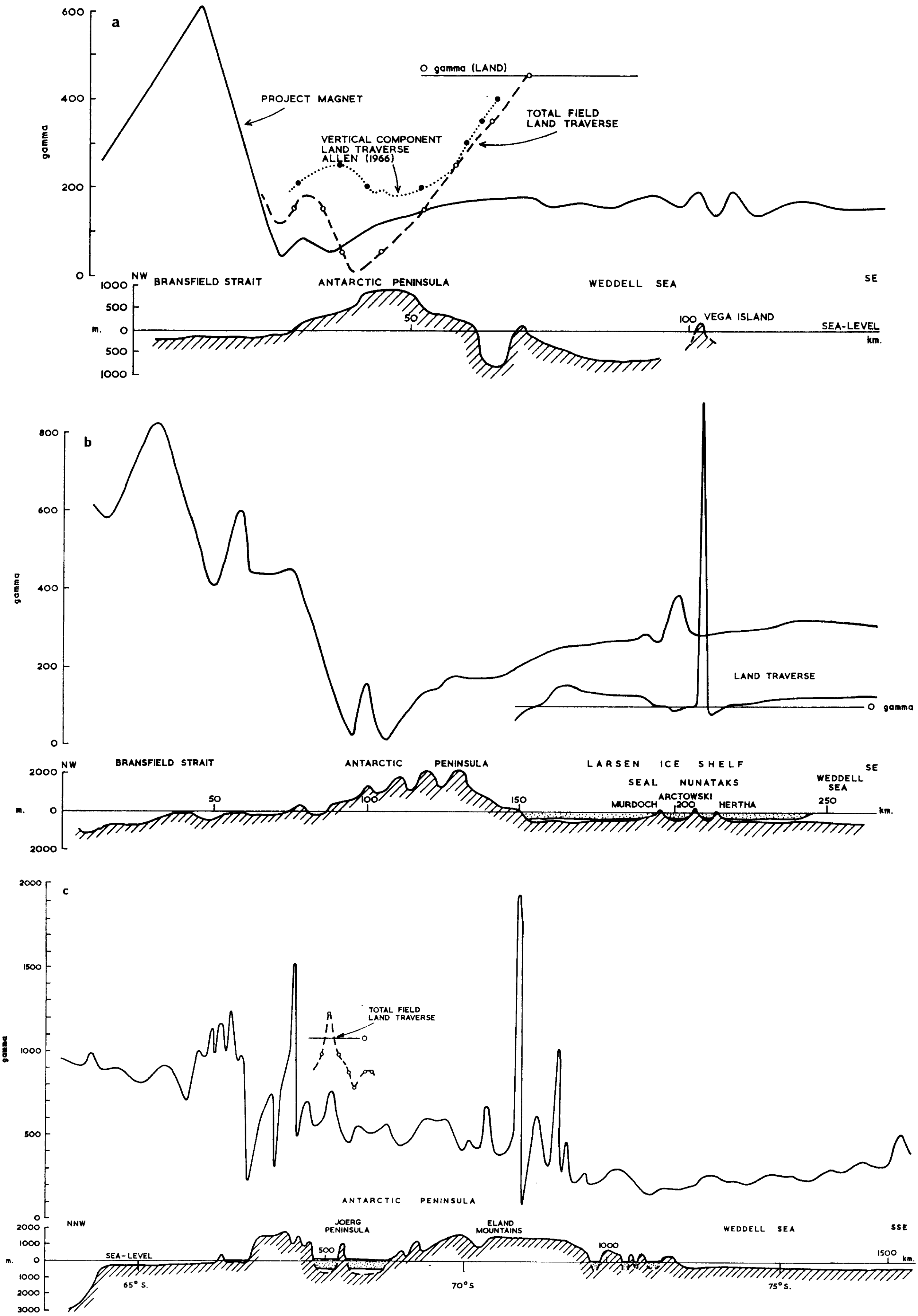

FIGURE 26

The aeromagnetic profiles over the Antarctic Peninsula (see Fig. 25).

a. Section AB.
b. Section CD.
c. Section EF.

been adjusted where possible to fit known land physiographical and magnetic features but navigational errors are debatable. The two more southerly traverses, 721N and 721S, are estimated by the author to be within a few kilometres of their true positions; this was based on the strong magnetic contrast between the Weddell Sea, Antarctic Peninsula and Bransfield Strait. The east–west traverse from Beak Island to Cape Legoupil (flight track 720S) has been re-plotted by several kilometres from the original grid positions in order to correlate not only with the different magnetic provinces but also to coincide with the local magnetic anomaly pattern found from the marine and land traverses within and adjacent to the flight section. Flight altitudes varied between 2·56 and 2·85 km. over an irregular topography which ranges from 1,000 m. below sea-level to over 2,000 m. above sea-level. Selection of individual anomalies suitable for interpretation was limited because of the flight elevation which caused attenuation and merging of adjacent anomalies. The data analysis was necessarily on a preliminary basis, investigation concentrating on short to medium wave-length anomalies (<40 km.) and with amplitudes in excess of 150 gamma. In all, seven isolated anomalies were selected and conditionally interpreted by the half-slope method of Peters (1949), giving maximum depths to the top of the source body. Table VI gives the results obtained which show close agreement where corresponding land traverses have been used to estimate depths to bodies, or where the near-surface geology is known. Exemplifying this is the aeromagnetic anomaly over the Seal Nunataks which has its counterpart in the total- and vertical-field traverses completed over the area, particularly those over Bull and Evensen Nunataks. Almost without exception, all of the Seal Nunataks have similar anomalies to that recorded by the Project Magnet traverse, which appears from the flight plan to be situated over Arçtowski Nunatak. Because of navigational inaccuracies, it may be incorrectly located and be representative of another of the nearby nunataks. Interpretation of the land profiles over these nunataks is given on p. 79, but depth estimates obtained from both land and aeromagnetic profiles place the anomaly source at or within a few metres of the surface.

A magnetic anomaly apparently observed over eastern Joerg Peninsula (Fig. 26c) has so far not been recorded on the land traverse over the same area (Fig. 59). However, it may be that the sparsely distributed

TABLE VI
MAXIMUM DEPTH ESTIMATES FOR THE PROJECT MAGNET ANOMALIES OVER THE ANTARCTIC PENINSULA

Locality	*Calculated maximum depth of body relative to sea-level* (km.)	*Relief over area* (km.)	*Depth below bedrock (no allowance for ice thickness on land)* (km.)	*Rocks within area capable of producing observed magnetic intensities*	
				Stratigraphical group	*Rock type*
Seal Nunataks	−0·2	−0·5 to +0·5	Body exposed	James Ross Island Volcanic Group	Olivine-basalts, vesicular olivine-basalts
West of Cape Herschel	−0·35	−0·25	0·1	—	—
Bragg Island to Lavoisier Island	−2·5 to −1·7	0 to −0·4	2·1 to 1·3	Andean Intrusive Suite Upper Jurassic volcanic rocks	Gabbro Tuffs, porphyritic andesites, silicified basalts
Joerg Peninsula to Cape Freeman	−2·2	−0·8 to +1·0	1·4 to 3·2	Permian Jurassic	Gneissose rocks Granite-adamellite, (?) gabbro
Matheson Glacier	−1·3	+0·5	1·8	Andean Intrusive Suite	Diorite, gabbro
Odom Inlet, Peak 2490	−2·8	+0·5	3·3	Andean Intrusive Suite	Diorite, gabbro

data of the land survey have only registered a section of the negative limb of a larger magnetic anomaly. The other alternative is again the possibility of an error in navigation sufficient to misplace the profile by several kilometres. Midway along the northern side of Joerg Peninsula (Stubbs, 1968) and 35 km. to the north in Seligman Inlet (Marsh, 1968), exposures of gabbro are to be found. Marsh also predicted their presence along the north wall of Trail Inlet. Modal analysis by Stubbs of a hypersthene-bearing gabbro revealed 9·1 per cent of iron ore, which if present in large bodies could account for the anomaly. More detailed land magnetic traverses could clarify the situation.

South of Joerg Peninsula there are only two anomalies which warranted investigation: those at Matheson Glacier (lat. 70° 47′ S., long. 62° 00′ W.) and Odom Inlet (lat. 71° 30′ S., long. 61° 20′ W.). In each instance, the anomalies may be slightly influenced by adjacent bodies causing slight irregularities in the profile form. Maximum depths have nevertheless been again estimated (Table VI) by the method of Peters (1949). This gave depths below sea-level of 1·3 and 2·8 km., respectively. Assuming a line source (Smellie, 1956), then depths below sea-level of 0·75 and 1·3 km. were obtained. To determine more precisely the sub-surface distribution of the bodies, a two-dimensional computer interpretation was applied. Due to the lack of any field specimens for magnetic intensity measurements, susceptibilities were first approximated by using the method described by Vacquier and others (1951). After repeated attempts at matching the calculated and observed profiles, including modifications to the susceptibility, the bodies showing the closest fits are illustrated in Figs. 27 and 28. Because of possible inaccuracies in navigation, all of the interpretation was carried out relative to sea-level with little consideration of the surface topography. With the Matheson Glacier anomaly the upper surface of the body was calculated to lie 250 m. above sea-level, topographic relief in this area averaging about 500 m. For the Odom Inlet anomaly, a depth of 250 m. below sea-level was calculated to the upper surface.

Adie (1955, 1957) has described the coastal geology south of Three Slice Nunatak as consisting of sediments of the Trinity Peninsula Series and quartz-diorites of the Andean Intrusive Suite. Modal analyses revealed an almost negligible amount of magnetite (0·1 per cent) in the latter and therefore they are unlikely

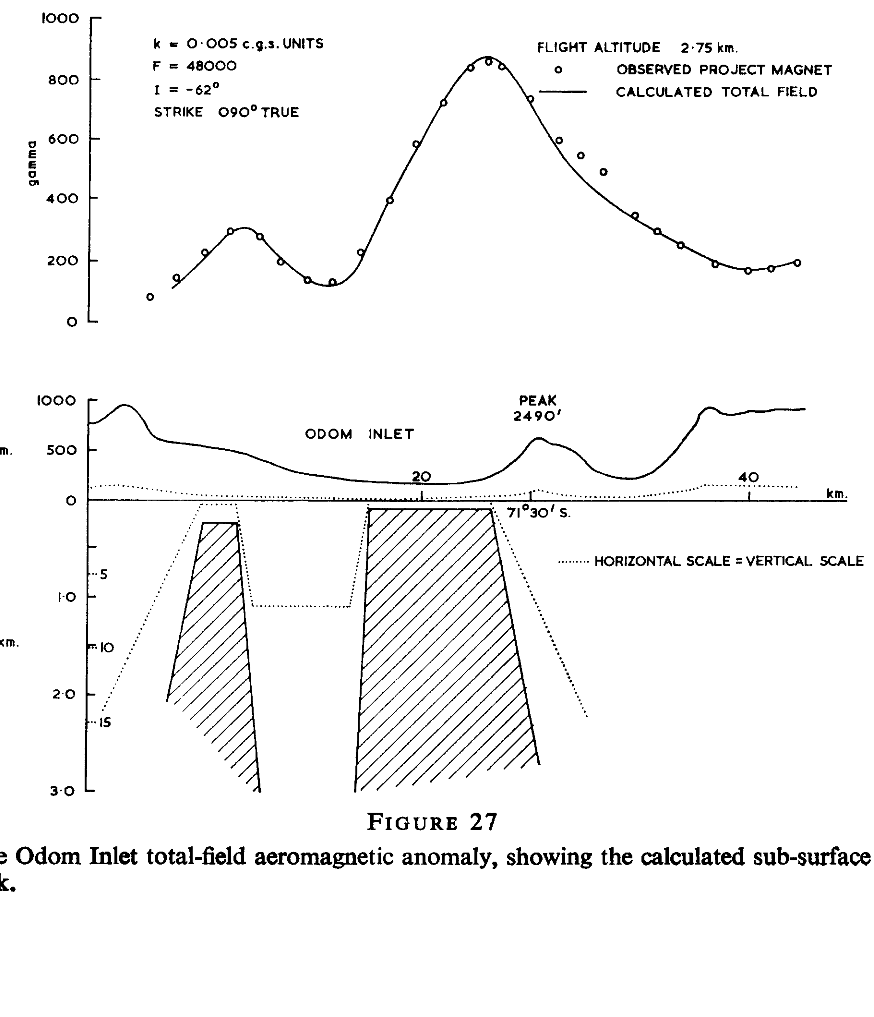

FIGURE 27

Interpretation of the Odom Inlet total-field aeromagnetic anomaly, showing the calculated sub-surface distribution of the anomaly source rock.

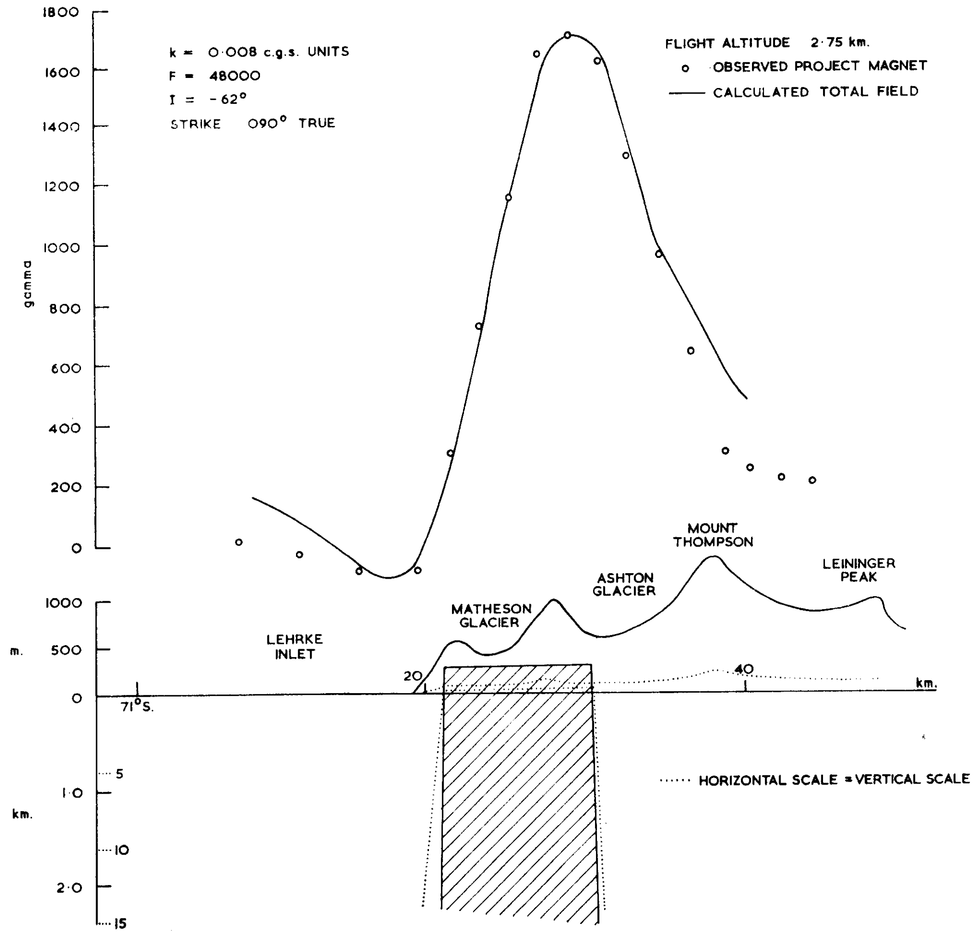

FIGURE 28

Interpretation of the Matheson Glacier total-field aeromagnetic anomaly, showing the calculated sub-surface distribution of the anomaly source rock.

to be the cause of the observed anomalies. However, at Cape Bryant (lat. 71° 12′ S., long. 60° 55′ W.), less than 30 km. from the flight line and within 50 km. of either anomaly, the presence of uralitized hornblende-gabbros has been recorded, modal analysis indicating up to 6·4 per cent magnetite content. These gabbros, also found at Cape Christmas (lat. 72° 20′ S., long. 60° 41′ W.) 120 km. to the south and associated with quartz-diorites of the Andean Intrusive Suite, could account for the observed anomalies.

At the present time there are no other geophysical data available which could provide information on the magnetic "basement". The nearest research area has been the United States Eights station, about 450 km. south of Odom Inlet. These results have been discussed earlier (p. 5) and, assuming that a section similar to Fig. 2 is applicable to the north towards the Odom Inlet and Matheson Glacier areas, then it would appear that the source of the anomalies lies within the 5·29 km. sec.$^{-1}$ layer or at the top of the 6 km. sec.$^{-1}$ layer. Behrendt (1963), who remarked that the 3·5 km. layer is the thickest layer of velocity 5·29 km. sec.$^{-1}$ measured anywhere in western Antarctica, compared this with the 4 km. of Cretaceous sediments found in the northern part of the Antarctic Peninsula (Halpern, 1964). The author suggests that the 3·5 km. layer may be equivalent to the Trinity Peninsula Series, which is known to occur in great thicknesses in northern Graham Land and is also possibly represented on Bowman Peninsula (Adie, 1957) about 150 km. east of the refraction profile. In both instances it is observed in close association with the igneous rocks of the Andean Intrusive Suite which are exposed in varying degrees through the sediments.

It is likely that either the presence of thick sedimentary deposits or of the more acidic members of the Andean Intrusive Suite explains why the aeromagnetic profiles over northern Graham Land are relatively free from short wave-length magnetic anomalies.

4. *Magnetic anomalies off the west coast of the Antarctic Peninsula*

Bregman and Frakes (1970) commented on a series of large magnetic anomalies occurring off the west coast of the Antarctic Peninsula and which they associated with the islands comprising the Scotia Ridge. These anomalies are considered as either reflecting a continuous belt of magnetic disturbance or as originating from isolated centres. Marine magnetic traverses (Griffiths and others, 1964; Harrington and others, 1971; Watters, 1971) off the South Orkney and South Shetland Islands have revealed similar anomalous areas which may be equated to those above especially when due consideration is given to possible flight navigational errors.

It is not intended here to make further interpretation on this group of anomalies but to comment on another set immediately to the west of the mainland between lat. 63° and 67° S. Although of smaller amplitude, they can nevertheless be identified both from the Project Magnet profiles and from the near-shore marine magnetic traverses. Fig. 29 illustrates these anomalies. Maximum diurnal variation recorded at the Argentine Islands during the period of marine magnetic surveying was approximately 70 gamma with average variations at 30 gamma.

The anomalies appear to be superimposed upon a background of higher magnetic intensity than found over the mainland. From Fig. 29, the general anomaly pattern shows that immediately adjacent to the coast there is a positive anomaly (A) having a width of about 20 km., an amplitude of at least 800 gamma and a strike trending parallel to the mainland. A magnetic trough (B) separates the positive anomaly (A) from the positive anomaly (C) to the west. Anomaly C, about 40 km. in width, appears to be formed by at least

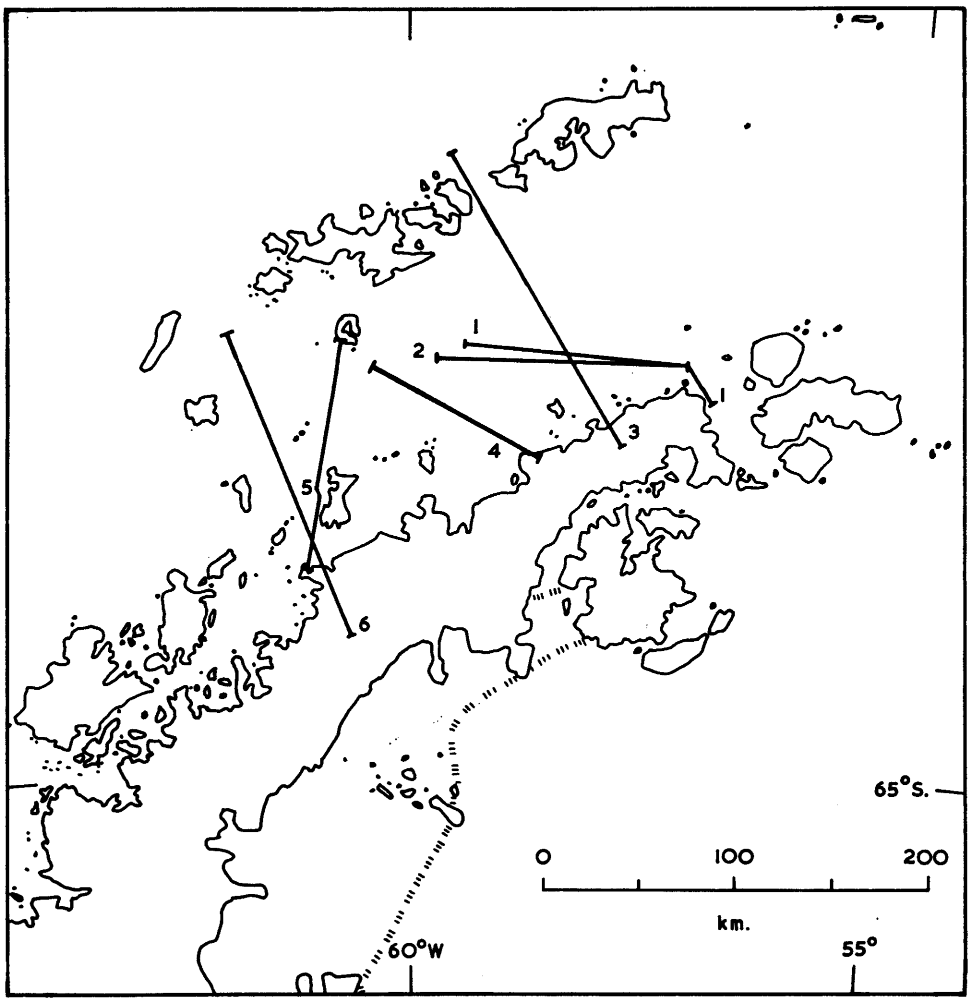

FIGURE 29

Diagram illustrating the distribution of magnetic anomalies off the west and north-west coasts of Graham Land. The cross-sections 1–6 are given on the opposite page.

two or three units, the most distinctive of which is the westernmost, seen in profiles 1 and 2 in Fig. 29. Although the positioning of the profiles, especially that of the Project Magnet traverses, is somewhat questionable in places, the broad anomaly pattern is still apparent. In Fig. 30 the anomalies off the north-western end of the Antarctic Peninsula have been correlated into magnetic zones. Between Tower and

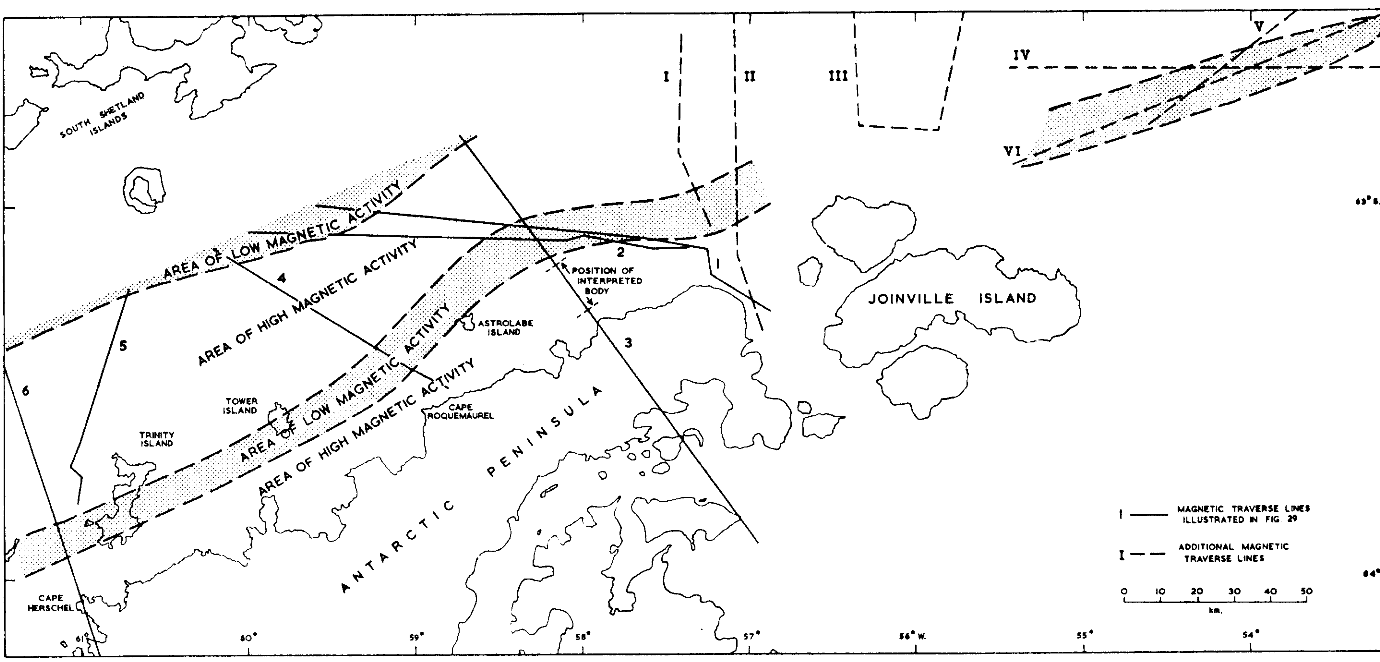

FIGURE 30

Sketch map showing the trend of magnetic anomalies north and north-west of Trinity Peninsula.

Astrolabe Islands the inshore anomaly B swings northward in a manner not dissimilar from that established by the gravity contours over the same area (Davey, 1971).

Anomaly A (Fig. 29, profile 3) of the aeromagnetic flight track 720S was interpreted using two methods. First, application of the method described by Grant and Martin (1966) suggested a body at a depth of about 2·7 km. below sea-level, a width of 15 km. and a susceptibility of 0·005 c.g.s. units. Using this as a basis for a more rigorous two-dimensional interpretation, various magnetic profiles were calculated for specific body forms. It was decided to place a certain amount of restriction on the upper surface of the body in order that it would conform to the seismic refraction investigation over the area (Ashcroft, 1972). After several body fittings, Fig. 31 indicates the model selected as being representative of the calculated profiles.

Ashcroft proposed that major faulting has occurred off north-west Trinity Peninsula with faults down-throwing 1 or 2 km. to the north-west and striking parallel to the coast. The introduction of a 6·3 km. sec.$^{-1}$ layer (Davey, 1971) beneath the shelf area suggests that rocks of a basic composition are present which, according to Nafe and Drake (1959), would have an equivalent density of 2·76 g. cm.$^{-3}$ and could therefore produce the measured anomalies. Geological investigations recently carried out by the Survey on the offshore islands from Cape Siffrey to Trinity Island do show that in addition to sediments and volcanic representatives there are considerable exposures of basic igneous rocks. These include basalts and dolerites variously exposed on Gourdin Island, Lafarge Rocks, Montravel Rocks, Astrolabe Island, Hombron Rocks and Tower Island. Halpern (1964) believed that a gabbroic body may be centred around Bulnes Island 5 km. from Cape Legoupil. From the geology of the area around Cape Legoupil, Elliot (1965) has proposed a major fault between the mainland and the offshore islands and faulting has also been discussed by Halpern (1964). Recent gravity traverses (Davey, 1971) off the north-west coast of Graham Land have enabled the compilation of a more comprehensive Bouguer anomaly map in this area than the one previously published by Davey and Renner (1969). Regional Bouguer anomalies vary between 50 and 100 mgal, and the relative distribution of the main anomaly zones, together with the traverses on which they are based, are shown in Fig. 32. Davey (1971) correlated the marine gravity and seismic data, and concluded that

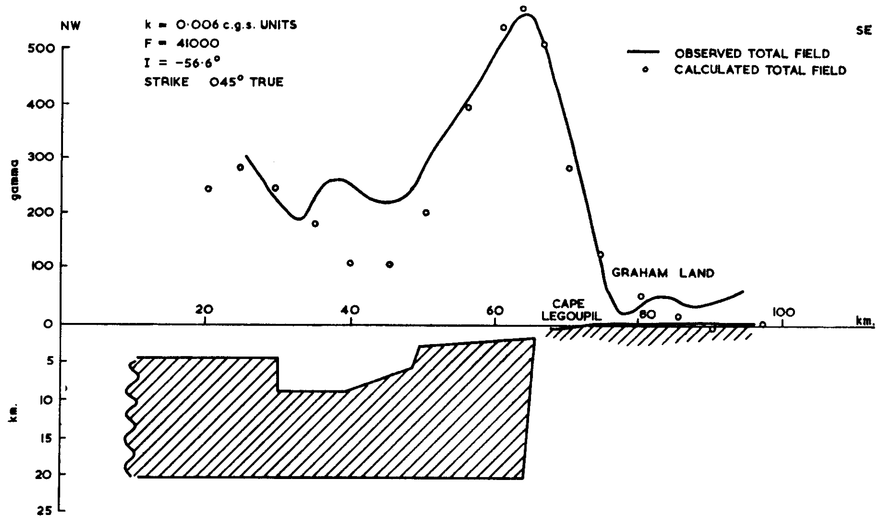

FIGURE 31
Interpretation of the aeromagnetic anomaly west of Cape Legoupil.

similar structures exist off Gourdin Island and Cape Legoupil. Re-positioning of the magnetic traverses within an expected navigational error could similarly improve the correlation with the seismic and gravity results. In this way, the gravity "high" adjacent to the coast between Astrolabe Island and Cape Siffrey would correlate with the area of positive magnetic intensity A. In the interpretation of anomaly A (Fig. 31) it was found that, provided the faulted boundaries are accepted as correct, the calculated anomaly while closely matching in profile that observed was originally located about 13 km. to the south-east. Such a displacement of the observed aeromagnetic profile can easily be attributed to navigational errors. If this were so, the smaller subsidiary anomaly to the east of anomaly A would correlate more closely with that observed from the land traverses. The cause of the positive magnetic and gravity anomalies could be due to a rise in the magnetic basement, while more negative areas may have an associated increase in sediment thicknesses. Davey (1971) calculated that 15 km. north-west of Cape Siffrey about 1,500 m. of sediments of density $2\cdot5$ g. cm.$^{-3}$ overlie the $6\cdot3$ km. sec.$^{-1}$ layer.

Farther south along the west coast of the Antarctic Peninsula the offshore geophysical traverses are few. An aeromagnetic traverse passes over Cape Herschel and another through Cape Rey (Fig. 33); preliminary

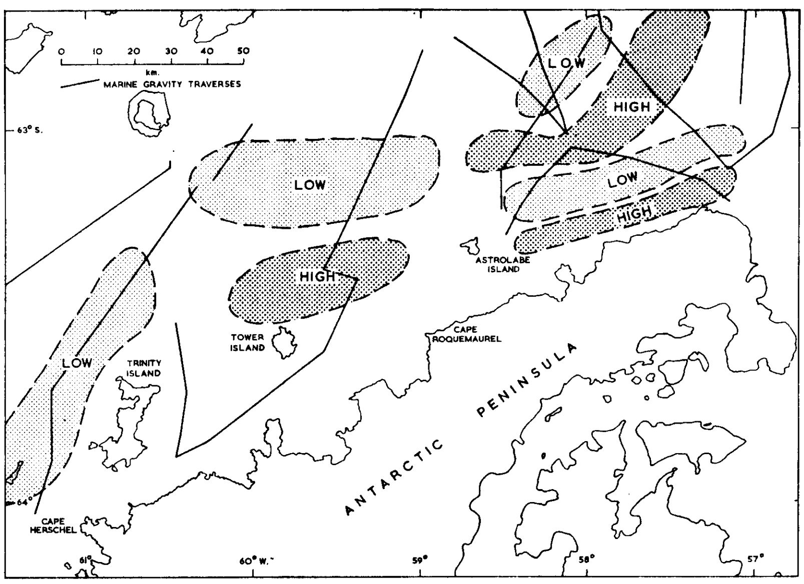

FIGURE 32
Sketch map showing the distribution of Bouguer anomaly zones off north-west Trinity Peninsula.

interpretation (p. 53) has already been carried out on these. The residual anomalies appear to be of greater intensity than those to the north, though errors in the accepted regional field may account for some of this. The total width of the broad magnetic feature is still around 80 km. and in each case the individual anomaly pattern identified to the north can be recognized. Depths to the top of the anomaly-producing bodies for that section of the Cape Rey profile between Bragg Island and Lavoisier

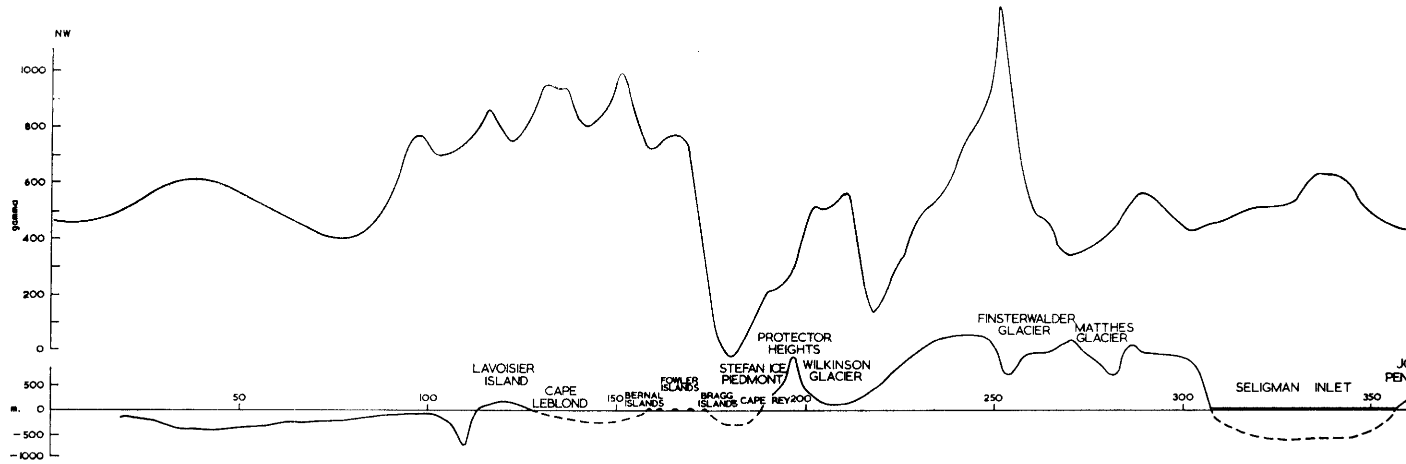

FIGURE 33

A section of the Project Magnet aeromagnetic traverse (flight track 721S) north of lat. 68° S.

Island are calculated to be within 350 m. of sea-level. Gabbro and basic extrusive rocks have been recognized in the latter area but at Intercurrence Island (16 km. north-west of Cape Herschel) dolerites are exposed. The shallow depth of 350 m. below sea-level calculated for the anomalies off Cape Herschel may be explained by such representatives. Off Stonington Island total-field magnetic traverses completed on the sea ice show a westerly increase of 1,400 gamma within 5 km. of the mainland.

The geophysical investigations along the west coast of the Antarctic Peninsula have supported, particularly in the northern areas, the presence of a continuous anomalous band, the characteristics of which are recognizable along all of the observed magnetic profiles. It would appear that there is a sharp rise in magnetic basement immediately to the west of the mainland and that it remains shallow over a belt with a width of at least 80 km. Superimposed on the main anomalies are the smaller yet distinct anomaly trends caused by variations in the upper surface of the magnetic body. In the northern part of this area the faulting which is recognized immediately to the west of Cape Legoupil is also substantiated by the seismic investigations and to a lesser extent by the gravity measurements. The faulting is in the form of an up-faulted block about 16–20 km. in width trending parallel to the coast. The magnetic and gravity anomaly maps indicate that between Tower Island and Astrolabe Island a distinct northerly swing is seen in the contours, a characteristic which is also reflected in the bathymetry of the area. Although there is no direct evidence for structural control to account for the contour swing, it is worth recording the presence of three bathymetric troughs to the north-east of Astrolabe Island. These north-west to south-east features, reaching depths down to 870 m., cut deeply into the shelf area from Bransfield Strait. A somewhat similar structure is also apparent to the south-west of Astrolabe Island. Halpern (1964) has postulated a left-lateral displacement along offshore north-west to south-east faults, one of which through Cape Legoupil could feasibly continue through one of the bathymetric troughs to the north-west.

On the mainland, parallel though offset faulting has been suggested through Misty Pass from both geological (Elliot, 1965) and geophysical evidence (Allen, 1966*a*). A continuation of such a fault line to the south-east follows the major topographical feature of Broad Valley and then, via Eyrie Bay, attains Prince Gustav Channel. In Broad Valley there is the only exposure of tuffaceous agglomerates belonging to the James Ross Island Volcanic Group west of a line through Tabarin Peninsula and Prince Gustav Channel. If movement has occurred along this fault, it must be sinistral in nature. Several ship-borne magnetic traverses which run across the area north and east of Cape Siffrey–Joinville Island have proved useful in delineating a possible 160–200 km. continuation of the magnetic pattern described lying to the south-west. It is not proposed here to give a detailed analysis of these profiles, since they are being independently assessed by the marine geophysics group at the University of Birmingham. However, it is necessary to comment on the anomaly distribution north-east of Joinville Island. Watters (1971) has contoured the magnetic values in this area and has compared the anomalies with those caused by the Andean Intrusive Suite on the mainland. The marine anomalies reach a recorded minimum of about −900 gamma and they transgress a bathymetry of between 200 and 2,000 m. They also exhibit an elongated north-east to south-west trend and therefore show a different structural attitude to those found on land. Furthermore, from an examination of their relevant sections, it is apparent that the magnetic pattern is similar to that on the profiles adjacent to the west coast of Graham Land. Of particular significance is the area of low magnetic activity, referred to earlier as low B, which can be tentatively traced along the west coast of Graham Land, through profiles I and II, to link with the isogams contoured from profiles IV, V and VI (Fig. 30). Insufficient data are available concerning the true extent of this "low", especially to the immediate north of d'Urville Island, which would appear to interrupt the otherwise continuous belt of negative magnetic activity. However, it is possible that in this area the anomaly trend may swing gently to the north thus by-passing the island, or that faulting may have caused some displacement. As yet, the magnetic structure has not been identified east of long. 53° W. though it would appear to have terminated by this latitude.

5. *Magnetic anomalies over the ice shelves of the Antarctic Peninsula*

The land magnetic surveys over the Antarctic Peninsula have shown the existence of numerous often large-amplitude anomalies which have been assigned to the igneous rocks belonging either to the Andean Intrusive Suite or to those of a Jurassic age. Likewise, the Project Magnet aeromagnetic flights (Bregman and Frakes, 1970) have registered the existence of such anomalies on two traverses over Graham Land and also over Palmer Land.

In marked comparison with the land traverses, those measured over the Larsen Ice Shelf exhibit negligible magnetic activity except for the local anomalies associated with the igneous rocks at Tonkin Island (Fig. 59) and the James Ross Island Volcanic Group at the Seal Nunataks (Fig. 47). Compared with the land values, the ice-shelf anomalies, where present, are of small amplitude (<300 gamma) and large wave-length (>10 km.). The three Project Magnet traverses over the continental shelf of the western and southern Weddell Sea are featureless, being almost devoid of any magnetic pattern. Marine surveys over the continental shelf of the Ross Sea (Adams and Christoffel, 1962) reveal a similar magnetic situation, again suggesting the absence of near-surface igneous bodies likely to cause magnetic disturbances. Thiel and Behrendt (1959) have published vertical-intensity measurements for the Filchner Ice Shelf. The data are corrected for temperature effect and overnight drift but they remain uncorrected for diurnal variation. Removal, by the author, of an approximate regional gradient shows magnetic profiles exhibiting some degree of correlation with the bottom topography, as deduced by Behrendt (1962) from the seismic soundings. Over areas where the ice is shown to be grounded, increased magnetic activity is observed with amplitudes of about 400 gamma. Elsewhere along the profiles, the anomaly amplitudes are of the order of 100–200 gamma though the distribution of the maxima suggests that the diurnal variation may be a major contributory factor. Examination by the author of the *K* and *Q* indices from the geomagnetic observatory at Halley Bay (lat. 75° 30′ S., long. 26° 42′ W.) indicates that variations of this order could be expected over the survey period.

From seismic and gravity data, Behrendt (1962) estimated that the Berkner Bank beneath the Filchner Ice Shelf is composed essentially of low-density morainic material, while from seismic methods in the "Little America" area of the Ross Ice Shelf Crary (1961) measured a maximum thickness of 1,325 m. of glacial sediments. Although such a thickness of low-susceptibility sediments would not significantly reduce basement anomalies, they could reduce the amplitudes of shallower short wave-length anomalies. On the evidence available, sediment thicknesses cannot be estimated for the areas beneath the Larsen Ice Shelf nor can it be concluded that the same magnetic pattern holds throughout the ice shelves of the Antarctic Peninsula. A total-field traverse carried out by J. Ross on the Wordie Ice Shelf revealed anomalies of up to 1,800 gamma, which is considerably greater than might have been expected. From field observations, radio echo-sounding measurements and from the examination of air photographs, the Wordie Ice Shelf is seen to have an irregular surface relief including ice rises and abundant crevassing. The physiographical evidence thus supports an irregular sub-surface bedrock topography, which, if of a sufficient magnetic intensity, could explain the anomalies and indicate the absence of thick deposits of sediments. Radio echo-sounding data (Smith, 1972) over the Wordie Ice Shelf show that ice-thickness variations may be broadly correlated with the magnetic observations.

III. GRAVITY SURVEYS

DURING the marine geophysical programme, Worden gravimeters Nos. 556 and 743 have been used by small landing parties operating from R.R.S. *Shackleton* and R.R.S. *John Biscoe*. Access to the mainland and offshore islands can be extremely difficult due to the lack of sheltered landing sites, ice cliffs and frequent adverse sea and weather conditions, and the pressing commitments of the ships whose time is primarily governed by the relief and re-supply of the Survey's stations. However, about 110 stations have so far been occupied though the majority of these are concentrated along the west coast of Graham Land north of the Argentine Islands station on Galindez Island (lat. 65° 15′ S., long. 64° 16′ W.). Those on the east coast are shown in Table VII. Gravity base stations do exist on the Survey's stations at Adelaide Island

TABLE VII
GRAVITY STATIONS PREVIOUSLY OCCUPIED
(GRIFFITHS AND OTHERS, 1964) IN NORTH-EASTERN TRINITY PENINSULA

Gravity station	*Date of original occupation*	*Lat. °S.*	*Long. °W.*
Hope Bay	January 1960	63° 24′	56° 58·6′
Eagle Island	January 1960	63° 38′	57° 24′
View Point	February 1960	63° 33′	57° 21′
Cape Burd	February 1960	63° 39′	57° 07′
Jonassen Island	February 1960	63° 33′	56° 43′
Cape Longing* (Longing Col)	November 1961	64° 28′	59° 00′
Crystal Hill†	December 1962	63° 34·2′	57° 49′
Beak Island	December 1962	63° 39′	57° 17·3′

* Originally occupied by aircraft from Deception Island, giving an estimated height of 76 m. a.s.l. The author estimated a latitude nearer 64° 25′ S. and a height, from aneroid barometers, of 78 m. above sea-level. The ice-shelf height in this area was recorded as 35 m. a.s.l.
† The author estimates the latitude to be 63° 39′ S.

(lat. 67° 46′ S., long. 68° 55′ W.), Stonington Island (lat. 68° 11·3′ S., long. 67° 00·5′ W.) and at Fossil Bluff (lat. 71° 20′ S., long. 68° 17′ W.).

Prior to 1963, the station at Hope Bay had been concerned only with vertical- and total-field magnetic surveys. These were continued in 1963–64 but the programme was enlarged by the addition of a gravity survey using a Norgaard gravimeter. Gravity stations involving about 60 sites were occupied over Tabarin Peninsula and Duse Bay. The same season as land gravity was commenced from Hope Bay, P. Kennett, working in conjunction with geological field parties, began a gravity and magnetic programme from Stonington Island. The gravity survey, using a Worden gravimeter (No. 556), included local traverses on the west coast (Kennett, 1965*b*), while the main summer programme resulted in the completion of a traverse of the peninsula to the Larsen Ice Shelf, where readings were taken northward to Cape Disappointment (Kennett, 1965*c*, 1966*a*, *c*). In the following season (1964–65), the author continued the geophysical surveys from Stonington Island with traverses to Hope Bay, thus enabling the survey to close with those stations (Table VII) occupied by Griffiths and others (1964). In all, over 600 stations have been occupied on the east coast of the Antarctic Peninsula between lat. 68° 35′ S. (Mobiloil Inlet) and lat. 63° 24′ S. (Hope Bay).

During the 1964–65 summer season, a LaCoste and Romberg geodetic gravimeter was kindly loaned

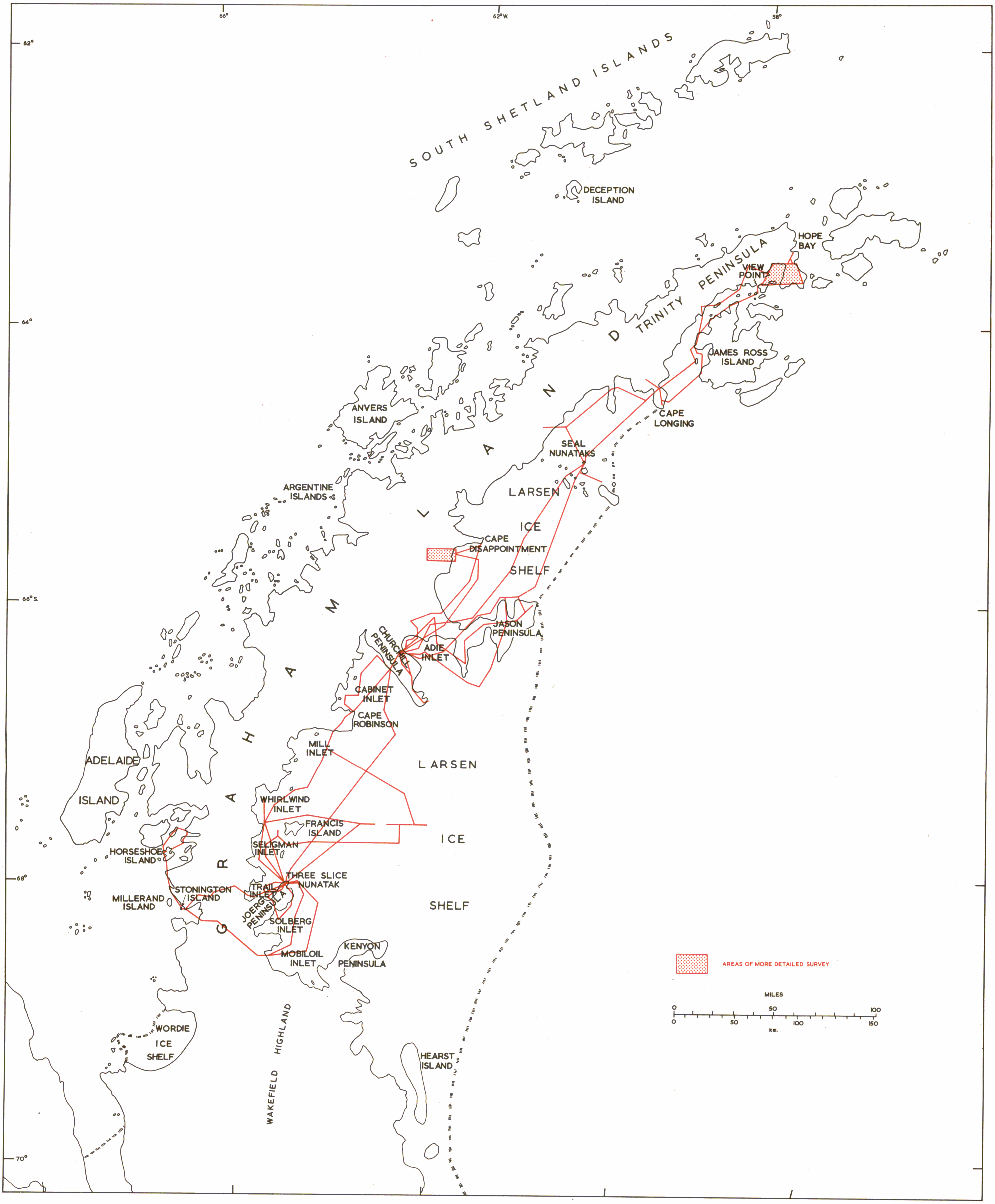

FIGURE 34
Gravity traverses in Graham Land.

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to the British Antarctic Survey by the U.S. Navy Oceanographic Office. The instrument, housed in R.R.S. *Shackleton*, was used to strengthen the gravity link between Antarctica and the international network in South America (Kennett, 1965*a*). In 1965–66 and 1969–70, F. J. Davey (1971) of the Sub-Department of Geophysics, University of Birmingham, used a Graf-Askania Gss 2 sea gravimeter installed in H.M.S. *Protector* to complete several marine traverses from the Falkland Islands via Bransfield Strait to Marguerite Bay.

Isolated land gravity stations have been occupied seasonally on the west coast of the Antarctic Peninsula as part of the marine programme. Apart from this, land geophysical investigations lapsed until 1967. At this time, regional traverses re-commenced with journeys essentially south from Stonington Island to and also radiating from the field station at Fossil Bluff.

This report deals essentially with the work completed from Stonington Island during 1964–65 but it necessarily incorporates the traverses of previous workers in the area (Fig. 34). The land and part of the marine gravity data were published as a Bouguer anomaly map of Graham Land by Davey and Renner (1969). This showed that the Bouguer anomalies became more negative towards the central axis and also at higher latitudes. Subsequent marine work in Bransfield Strait has added to this (Davey, 1971), whilst it is hoped that current land investigations will eventually clarify and extend our knowledge of the structure of the area around and south of lat. 68° S.

A. Field Observations, Errors and Rock Densities

1. *Gravity traverses from Stonington Island*

Apart from the Trinity Peninsula area, it has as yet neither been practical nor possible to traverse from the west coast to the east coast at any point north of Stonington Island. This has, of course, severely restricted investigations into a most important aspect of Antarctic Peninsula geology and geophysics, namely the crustal structure beneath its axis. Overland access to the east coast from Stonington Island is again subject to the same problems in that the plateau area must be traversed. At the latitude of Marguerite Bay, the plateau averages 40 km. in width, attains a height of 1,500 m. and its precipitous edges only permit access by glacier routes which in themselves are often highly crevassed and hazardous. Fortunately, there are two lines of access from Stonington Island (Fig. 35) which have been successfully followed and along which geophysical traverses have been run. The more southerly route, along which a geophysical

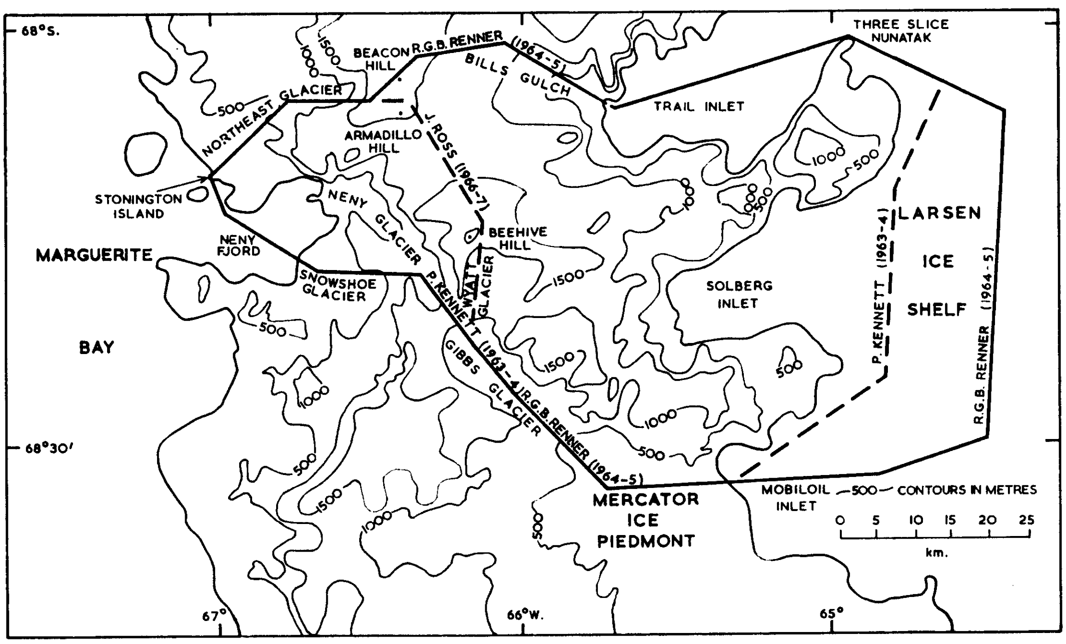

FIGURE 35

Topographical map illustrating geophysical traverse lines from Stonington Island to the Larsen Ice Shelf.

traverse was first made by Kennett in 1963–64, was via Neny Fjord–Snowshoe Glacier–Gibbs Glacier to Mobiloil Inlet. This, the slightly longer route can only be undertaken when good sea-ice conditions exist in Marguerite Bay but it has the advantage that it only ascends to 1,000 m. Neny Trough, through which much of the route follows, is a large topographical feature presenting one of the few if not the only easy route traversing the peninsula. Because of the structural complexities involved, the geophysical readings obtained are probably atypical of the regional anomaly pattern. The more northerly route, Northeast Glacier–The Amphitheatre–Beacon Hill–Bills Gulch to Trail Inlet, is a more direct one but it ascends to 1,700 m. and descends the steep and badly crevassed glacier through Bills Gulch. Geophysical measurements along this traverse do, however, give a more representative picture of the west–east gravity gradients and, because of its uniqueness, it is of great importance. Although elevation errors affect the overall accuracy, the resulting profile still provides qualitative information as to the structure of the peninsula as well as a comparison with the anomalous trend shown by the more southerly route. The accuracy may have been slightly improved if the west–east traverses could have been reversed but unforeseen circumstances curtailed the surveys on both of the return legs.

A compromise between these two routes has recently been followed by reaching the plateau via The Amphitheatre and then travelling south via Armadillo Hill–Beehive Hill–Wyatt Glacier to join the Gibbs Glacier section of the southern route. This overcomes both the sea-ice uncertainties and avoids the badly crevassed Bills Gulch. Both gravity and magnetic traverses have now been made along this route.

2. *Errors involved in Antarctic gravity surveys*

The majority of base-station values in Antarctica are tied in to the world gravity network of the University of Wisconsin with an accuracy better than 1 mgal (Bentley, 1964), but within the Antarctic gravity survey errors are likely to be large due to the adverse climatic, topographical and logistic difficulties often encountered. Errors may be classified into those in measured gravity and those resulting from the reduction of the data; they may be constant, as a result of the inaccuracy of base-station values, systematic through instrumental drift or random as caused by incorrect elevation determinations. Other errors requiring consideration in observed gravity include instrumentation irregularities such as variable travelling drift, non-linearity of the calibration factor between successive determinations and the effects of extreme temperature and pressure variations.

By far the largest single contributory factor governing the accuracy of gravity anomalies in Antarctica are those errors resulting from the inaccurate measurement of surface elevation. In general, two methods have been employed: optical levelling and a method based on barometric pressures. The former is by far the more accurate but it has obvious disadvantages in Antarctica, while the latter though the more widely used is dependent on reliable meteorological data to provide the regional horizontal pressure gradients. Using aneroid barometers but neglecting errors arising from barometric pressure, Behrendt (1964) gave a standard deviation in elevation of $\pm$ 2·38 m. over a 6 km. interval, but a consideration of both measuring and barometric errors increased this figure to $\pm$ 50 m. A value of $\pm$ 50 m. was also estimated by Behrendt and others (1962) for work in the Mount Murphy–Hudson Mountains area, whereas for work over the gentler terrain of the Filchner Ice Shelf $\pm$ 18 m. has been given (Behrendt, 1962). Increased accuracy may be obtained by using a closed loop traverse, Bentley and Ostenso (1961) having estimated a maximum error relative to a base station of $\pm$ 15 m. Bentley (1964), who discussed the problem of elevation errors at length, commented that many previously quoted elevation errors may be underestimates. For the Antarctic continent, he estimated that an average total error relative to a base station of $\pm$ 50 m. would be applicable, with a standard error of the mean of $\pm$ 25 m. Errors over ice-shelf areas and relative errors between adjacent stations can be expected to be less due to greater barometric control. During the work on the Larsen Ice Shelf, station heights were again determined using three aneroid barometers. Kennett (1965*c*) has discussed the errors involved. Uncertainties in ice thicknesses may also contribute substantially to the accumulative error. A change of 100 m. in the estimated depth of the ice–rock interface introduces an error of $\pm$ 7·4 mgal into the Bouguer anomaly.

Grushinsky and Frolov (1967) subdivided the combined errors according to locality. For coastal stations, a value of up to $\pm$ 2 to $\pm$ 3 mgal was given, in the hinterland $\pm$ 5 to $\pm$ 8 mgal and for sea ice $\pm$ 3 to $\pm$ 4 mgal. In discussing errors for most of the Antarctic interior, Bentley (1964) adopted a figure of $\pm$ 20 mgal for both free-air and Bouguer anomalies with $\pm$ 10 mgal being a more representative value for mean anomalies.

By normal geophysical standards, the above errors are unacceptably large but relative gravity values between adjacent stations are much more acceptable. Behrendt (1964), in a geophysical survey at the southern end of the Antarctic Peninsula, estimated that the relative error between adjacent stations, separation 6 km., was $\pm$ 0·7 mgal with a cumulative error between two seismic stations, 40 km. apart, of $\pm$ 2 mgal.

3. *Errors in gravity surveys of Graham Land*

a. *Errors in absolute gravity.* Gravity stations in the Antarctic Peninsula are linked to the international gravity network of South America by ties to Buenos Aires and Punta Arenas. From the original data (Griffiths and others, 1964), the estimated standard error of the link to Stonington Island was calculated at $\pm$ 1·4 mgal. During the 1964–65 Antarctic summer, Kennett (1965*a*) used a LaCoste and Romberg geodetic gravimeter to carry out further links to South America and he concluded that absolute gravity values on the peninsula should be decreased by 3·1 mgal from those adopted by Griffiths and others (1964). In 1966, Davey (Davey and Renner, 1969), using a Worden gravimeter, also revised the link and this corroborated the findings of Kennett. Table VIII shows a comparison of the individual measurements for the absolute value of gravity for Stanley (Falkland Islands), Deception Island and Stonington Island.

TABLE VIII

COMPARISON OF ABSOLUTE GRAVITY VALUES

(Gravity values in cm. sec.$^{-2}$)

	Stanley	*Deception Island*	*Stonington Island*
Griffiths and others (1964)	981·2452	982·2254	982·5125
Kennett (1965*a*)	981·2433	982·2223	982·5094*
Davey (personal communication)	981·2420	982·2224	982·5086

* Interpolated value from the survey by Griffiths and others (1964).

More recent links using Worden gravimeters have been completed but the accuracies of these are debatable. Due to the low and consistent drift of the LaCoste and Romberg gravimeter, it was decided to use 982·2223 cm. sec.$^{-2}$ for the value at Deception Island until a more definitive tie has been established. All of the land field data were related to Deception Island via the value of 982·5094 cm. sec.$^{-2}$ at Stonington Island. Early in 1969 a volcanic eruption at Deception Island destroyed the gravity base station used in the above links. Until the establishment of a new base station, alternative sites will have to be occupied (see Davey and Renner, 1969).

Apart from a few readings on the west coast, the present survey is concerned with the east coast of the Antarctic Peninsula where there are relatively few gravity base stations. This is most noticeable in the southern areas where their absence is due to the lack of suitable and accessible rock outcrops. Base stations originally occupied on the east coast have been listed in Table VII, the nearest to Stonington Island being at Cape Longing over 500 km. distant and which may involve 30 days of sledge travel. Intermediate stations on rock were occupied by Kennett (1966*a*) at Cape Robinson (lat. 66° 52·5′ S., long. 64° 05′ W.), Churchill Peninsula (lat. 66° 20′ S., long. 62° 55′ W.) and Cape Disappointment (lat. 65° 33′ S., long. 61° 45′ W.), but the absolute values at these localities are subject to cumulative drift errors as only a single outward traverse was completed; Kennett estimated a maximum drift error of $\pm$ 5 mgal at Churchill Peninsula. The survey by the author used the base stations of Griffiths and others (1964), but in addition secondary links were run to stations occupied by Kennett at Three Slice Nunatak, Cape Robinson and Churchill Peninsula.

Due to the time involved between occupation and re-occupation of base stations, the drift rate of the gravimeter assumed major importance. Because of this, drift checks were frequently made and the drift was subdivided into static and travelling components. The latter was obtained by subtracting the static (overnight and lie-up) drift from the overall drift between closures.

The winter months of July and August spent at the scientific station were used to study the static drift of the instrument. Local survey runs were also made to study the travelling drift. For the period of the summer journey the following drift rates were calculated:

Overall drift	+0·031 mgal/hr.
Static drift	+0·027 ± 0·005 mgal/hr.
Travelling drift	+0·054 ± 0·007 mgal/hr.

Close examination of the drift rates (Fig. 36) indicates that the greatest variation of rate occurred after severe changes in the ambient temperature. As the instrument was not kept at a constant temperature, the inability of the internal compensation of the gravimeter to correct for such reversals resulted in an increased temperature drift until the instrument again reached stabilization. As well as a general increase in the summer temperatures from September to January, there was also a fairly severe diurnal variation. The only

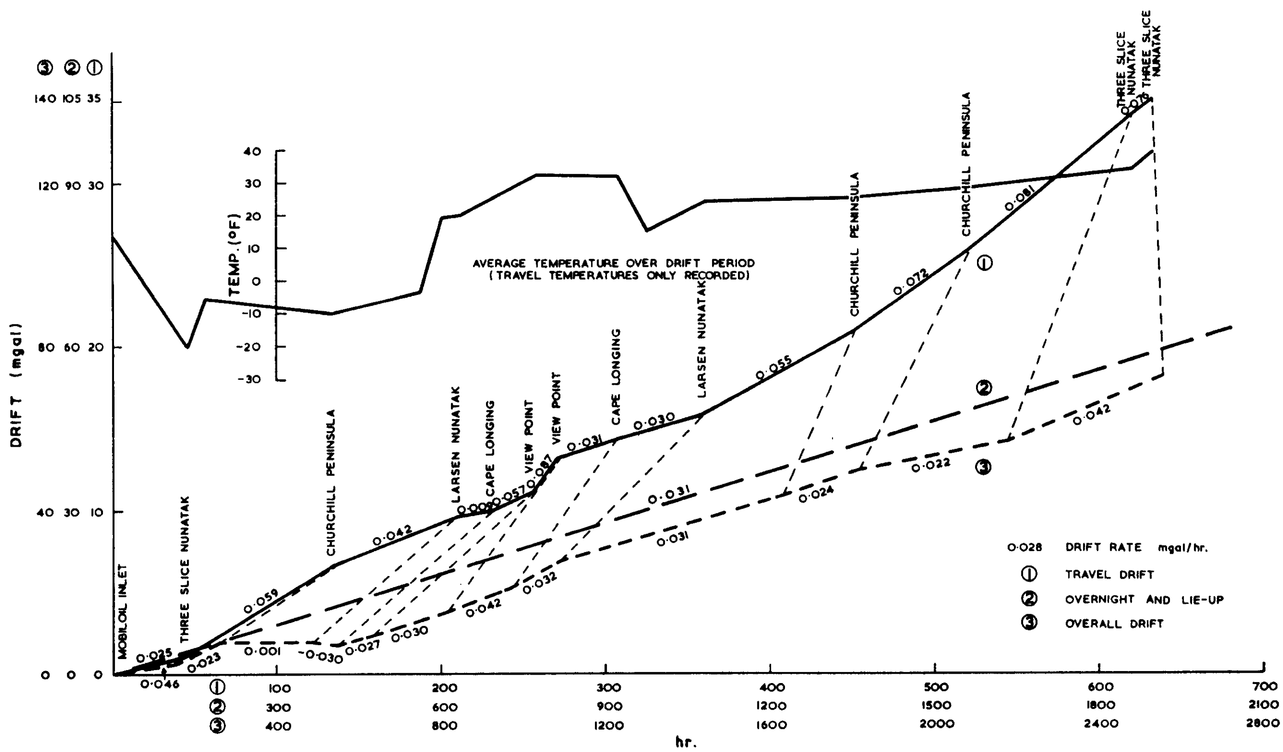

FIGURE 36

Graph of gravimeter drift rates as observed throughout the summer field journey of 1964–65.

temperatures measured on the journey were those taken at field stations and those taken before, during and after overnight stops and enforced lie-up periods. It is perhaps of significance that during the autumn journey (April–June) a travel drift of 0·025 mgal/hr. was found compared with 0·054 mgal/hr. during the summer journey. An explanation of this could lie in the fact that temperatures, although much lower than during the summer journey, were far more stable. Also, the surfaces travelled over were much softer which resulted in slower progress but a more even ride. During the summer journey of 1963–64 Kennett recorded the following drift rates:

Static drift	0·028 ± 0·004 mgal/hr.
Travel drift	0·045 ± 0·013 mgal/hr.

The gravity base station at Churchill Peninsula is about 90–100 hr. travelling time from either the base station to the south at Three Slice Nunatak or to the north at Cape Longing. This link presents the maximum error likely to occur due to drift closures.

In 1965 the Worden gravimeter was returned to England, where it was re-calibrated over two gravity base lines established by the U.K. Geological Survey. A calibration factor of 0·22902 mgal/scale division at 83° F [28° C] was determined compared to one of 0·22944 mgal/scale division measured in 1962 prior to the instrument going to the Antarctic. The latter value was used in the data reduction and if incorrect would cause a maximum error of 0·7 mgal in the value of absolute gravity at Churchill Peninsula.

Consideration of the above errors results in an estimated maximum error in the absolute values of gravity in the Churchill Peninsula area of ± 4·5 mgal. The standard error for the absolute values of gravity on the east coast of Graham Land is estimated to be ± 1·5 mgal.

b. *Errors in data reduction.* The theoretical values of gravity were taken from tables derived from the International Gravity Formula. As the gravity stations have an estimated position accuracy of ± 0·5 km. on land and ± 1·6 km. on the ice shelf, this corresponds to errors of ± 0·4 and ± 1·1 mgal, respectively.

Surface elevations were measured using three Walker aneroid barometers without any allowance for horizontal pressure gradients. Kennett (1966*a*) estimated that for the traverse across the Antarctic Peninsula heights may be subject to inaccuracies of ± 30 m., producing an error of ± 9 mgal in the free-air anomaly. Using a constant density of 2·67 g. $cm.^{-3}$, then for a given station on rock the corresponding Bouguer correction will be in error by ± 3·4 mgal; for a station situated on ice of density 0·9 g. $cm.^{-3}$ the error is ± 1·1 mgal. The combined Bouguer reduction (free-air and Bouguer correction) produces an error of approximately ± 6 mgal for stations on an all-rock column and ± 8 mgal for stations on an all-ice column. Stations forming part of shorter traverses on the peninsula and occupied from the Larsen Ice Shelf can be expected to have a greater accuracy as atmospheric conditions over a limited area and duration are likely to be much more stable. North of Cape Longing, several stations were visited which were on rock at sea-level, and for these the above errors are excessive and do not apply.

During the Antarctic summer season of 1966–67 several ice-thickness profiles were obtained over the central part of the Larsen Ice Shelf using a radio echo-sounding technique. Many of the profiles were intentionally flown over the geophysical traverses which greatly increased the accuracy of the height estimates (Renner, 1969) because it enabled the derivation of surface elevations to an accuracy of approximately ± 3 m. Allowing for local undulations in the ice shelf and incorrect contour extrapolation, a maximum heighting error of ± 30 m. is estimated. Applying the free-air correction to areas of relatively flat ice shelf, an error of ± 3 mgal is introduced compared with an error of ± 9 mgal in areas of extreme irregularity. No Bouguer correction was applied because in a floating ice shelf the mass deficiency below sea-level is compensated by the mass excess above.

Where possible, terrain corrections were applied using the relevant topography on the 1:200,000 maps published by the Directorate of Overseas Surveys together with the Admiralty Charts for stations affected by the bathymetry. The absence of detail on some of the land maps does in places limit the accuracy of the terrain corrections, especially on the routes to the plateau where there are severe topographical features. Apart from the few stations near such irregularities, the terrain correction can be expected to be accurate to within ± 1 mgal. Stations on the ice shelf in general did not require terrain corrections as many were situated up to 100 km. from the nearest significant topographical features.

Considering the errors involved from the measured gravity and from the reduction in the readings, the following combined standard errors have been calculated:

i. *For stations at sea-level*
The error in Bouguer anomaly is ± 1·5 mgal.

ii. *For land stations*
The error in Bouguer anomaly for an all-rock station is ± 6·0 mgal and for an all-ice station is ± 8·0 mgal.

iii. *For ice-shelf stations*
The error in free-air anomaly in relatively flat areas is ± 3·0 mgal and in areas of extreme undulation is ± 9·0 mgal.

The errors in Bouguer anomalies have not been calculated as the bathymetry beneath the Larsen Ice Shelf is unknown.

Fig. 37 illustrates the network of gravity base stations so far occupied between the west and east coasts of Graham Land.

4. *Rock densities*

Density measurements have been carried out on over 200 samples representing 165 different localities on the east coast of Graham Land between lat. 64° and 68° 30 ′S. The majority of the samples were collected

TABLE IX

DENSITY VALUES FOR ROCKS FROM THE EAST COAST OF GRAHAM LAND

Rock group	*Geological age*	*Rock types*	*Locality*	*Number of specimens*	*Number of localities*	*Mean dry density* (g. cm.$^{-3}$)	*Standard deviation*	*Standard error*
	Mid-Jurassic	Shales–phyllites (iron-enriched)[1]	Starbuck and Stubb Glaciers	2	2	2·79[6]	—	—
		Metamorphosed sandy mudstones, muddy sandstones, sandy shales, quartzites[3,4]	Moider Peak, south-east of Eden Glacier	25	19	2·76	±0·09	±0·02
Sedimentary (varying degrees of metamorphism)	Pre-Jurassic (post-metamorphic complex)	Above without quartzites	Moider Peak, south-east of Eden Glacier	20 33	14 27	2·79 2·75	±0·09 ±0·1	±0·02 ±0·02
		Quartzites only	Moider Peak, south-east of Eden Glacier	5	5	2·63	±0·06	±0·03
		Regionally metamorphosed sandstones and siltstones	Western Joerg Peninsula	8	8	2·74	±0·06	±0·02
	(?) Trinity Peninsula Series	Cataclastically and regionally metamorphosed phyllites	Three Slice Nunatak	(1)	(1)	(2·73)	—	—
	Trinity Peninsula Series	Greywackes	Downham Peak area	5	1	2·74[5]	±0·02	±0·01
Volcanic	Upper Jurassic to Triassic	Acid and intermediate pyroclastic rocks, rhyolitic-dacitic tuffs, tuff-breccias, lavas, agglomerates[3,4]	Coastal nunataks of Oscar II Coast, Churchill Peninsula, Spur Point, Cape Robinson, Friederichsen Glacier	24	21	2·63	±0·06	±0·01
		Crystal tuffs	Starbuck and Stubb Glaciers	5	5	2·67[6]	±0·04	—
		Rhyolite	8 km. west of Gemini Nunatak	1	1	2·64[2]	—	—
		Vesicular rhyolite	Cape Alexander	1	1	2·56[2]	—	—
		All measured samples of plutonic and hypabyssal rocks[3,4]	Oscar II, Foyn and Bowman Coasts	62	58	2·66	±0·07	±0·01
		Plutonic rocks: Granites, adamellites	Oscar II, Foyn and Bowman Coasts	26	24	2·62	±0·04	±0·01
		Granodiorites	Oscar II, Foyn and Bowman Coasts	10	9	2·71	±0·08	±0·03
Plutonic	Jurassic–Andean (few representatives)	Tonalites-diorites	Oscar II, Foyn and Bowman Coasts	8	7	2·73	±0·07	±0·02
		Gabbros	Churchill Peninsula—Adie Inlet, Joerg Peninsula, Cape Robinson, Seligman Inlet	7	7	2·82	±0·03	±0·01
		Granite	Starbuck and Stubb Glaciers	3	3	2·61[6]	±0·02	—
		Alkali-granite	Peak 1040, Churchill Peninsula	1	1	2·65[2]	—	—
		All measured samples belonging to metamorphic complex[3,4]	Oscar II, Foyn and Bowman Coasts	64	40	2·74	±0·01	±0·001
		Banded gneiss, composite gneiss, hornblende-biotite-gneiss	Oscar II and Foyn Coasts	27	25	2·70	±0·06	±0·01
	Metamorphic complex	Quartz-plagioclase-gneiss	Oscar II and Foyn Coasts	15	10	2·69	±0·04	±0·01
Metamorphic		Metagabbros	Oscar II and Foyn Coasts	2	2	2·81	±0·06	±0·04
	(Basement Complex)	Amphibolites	Oscar II and Foyn Coasts	7	7	2·94	±0·04	±0·02
		Granodiorite-gneiss, banded biotite-gneiss	(?) Cape Robinson area (boulders only), eastern and western Joerg Peninsula	5	4	2·63	±0·05	±0·03

[1] Fleet, 1966.
[2] Personal communication from P. Kennett.
[3] Marsh, 1968.
[4] Stubbs, 1968.
[5] Personal communication from J. Mansfield.
[6] Kennett, 1966*c*.

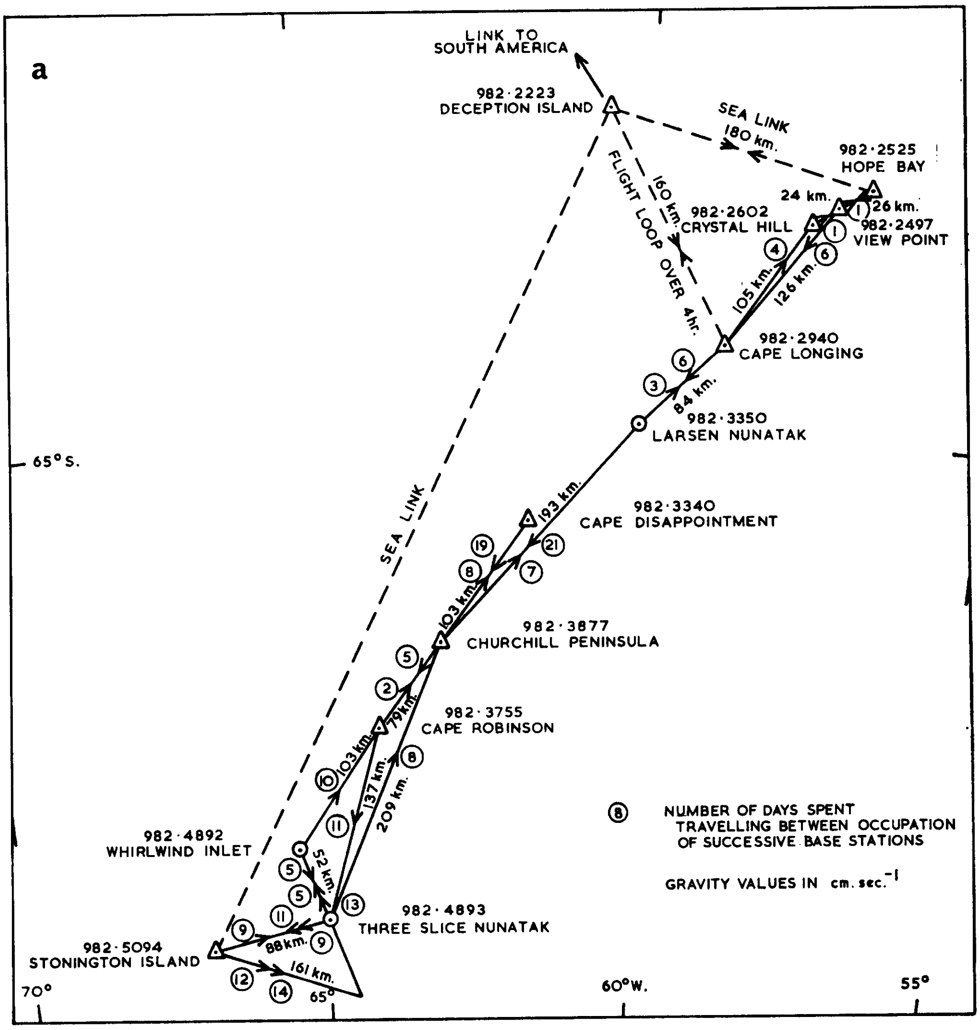

FIGURE 37a

Absolute values of gravity over the base-station network on the east coast of Graham Land.

by geological parties operating along the Oscar II, Foyn and Bowman Coasts, though these have been supplemented with specimens collected during and as part of the geophysical programmes. Identification of the rock specimens formed an integral part of the geological surveys carried out by Fleet (1968), Marsh (1968) and Stubbs (1968).

The samples have been classified in four major groups (Table IX). Where necessary these have been subdivided into stratigraphical divisions which in themselves contain variable rock types. From the table, the four main rock groups have densities as follows:

	g. cm.$^{-3}$
Sedimentary	$2{\cdot}75 \pm 0{\cdot}1$
Volcanic	$2{\cdot}63 \pm 0{\cdot}06$
Plutonic	$2{\cdot}66 \pm 0{\cdot}07$
Metamorphic	$2{\cdot}74 \pm 0{\cdot}01$

An interesting comparison can be made with the average density values given by Allen (1966*a*) for the Mott Snowfield area of Trinity Peninsula. A value of 2·76 g. cm.$^{-3}$ was given for the Andean Intrusive Suite, whereas the density of the Trinity Peninsula Series sediments has been determined as 2·67 g. cm.$^{-3}$. These values are almost the reverse of those given in this paper and, if they are correct, it suggests that further sampling is essential, particularly of these rock types in north-eastern Trinity Peninsula.

Adie (1953) completed density measurements on several samples of vesicular olivine-basalt from the

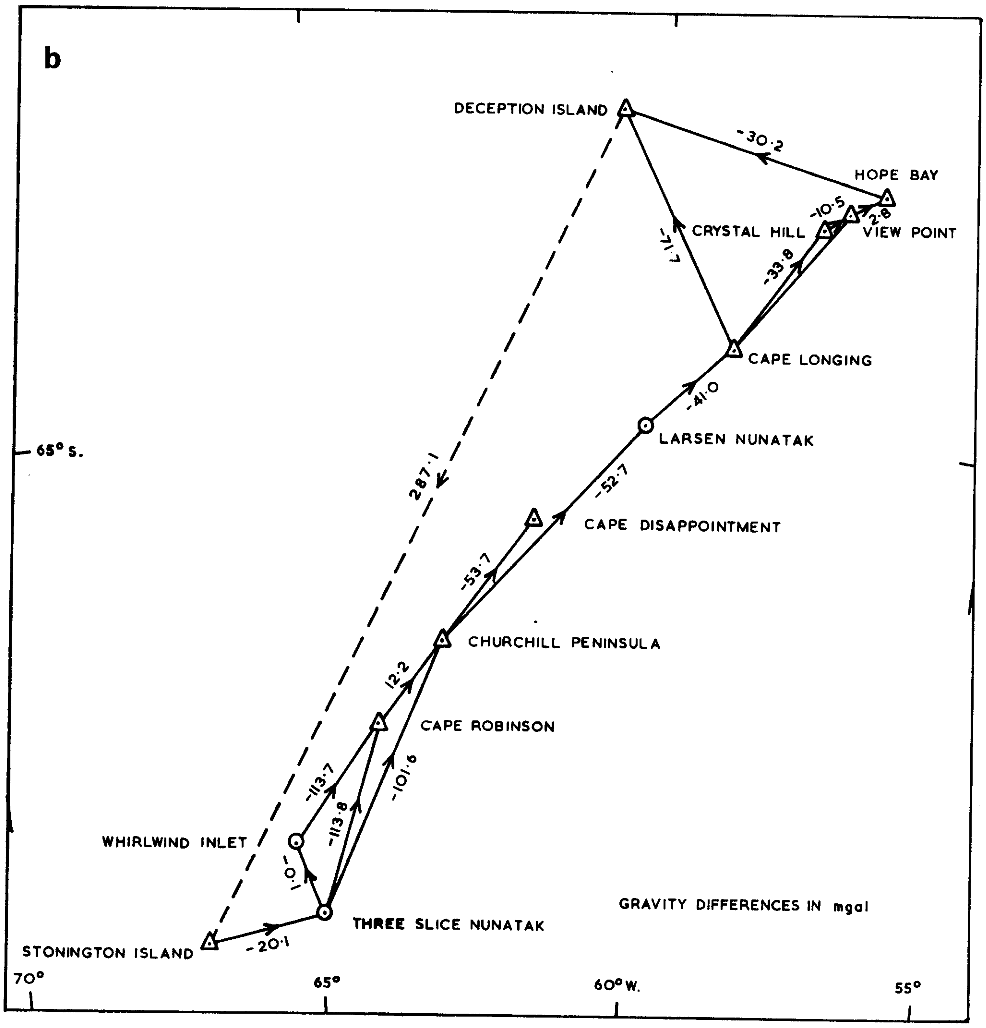

FIGURE 37b
Adjusted differences in gravity values between east coast base stations.

James Ross Island Volcanic Group. Three samples from the Seal Nunataks averaged 2·26 g. cm.$^{-3}$ compared with one of 2·14 g. cm.$^{-3}$ from Dagger Peak, James Ross Island.

B. INTERPRETATION OF GRAVITY ANOMALIES

1. *Tabarin Peninsula*

The geology of the southern part of Tabarin Peninsula, as described by Adie (1953), is formed by an interbedded succession of olivine-basalt lava flows, tuffs and agglomerates belonging to the James Ross Island Volcanic Group. These volcanic rocks crop out south of an east-north-east to west-south-west line which bisects Tabarin Peninsula slightly to the north of Buttress Hill and Brown Bluff. The nature of the boundary is not apparent but a faulted origin has been suggested (Adie, 1957). Ashley (1962) has interpreted the associated magnetic anomalies as due to at least two previously more extensive olivine-basalt lava flows whose presence is now confined to isolated localities situated 30–60 m. beneath the ice cover. A maximum thickness of lava beneath Buttress Hill was estimated at 240 m. but elsewhere on Tabarin Peninsula the calculated thicknesses for the lava flows were 120 m. or less. From magnetic observations in Duse Bay together with geological evidence from Eagle Island, Ashley estimated the James Ross Island Volcanic Group to be at least 1,220 m. thick. Bibby (1966), in a generalized stratigraphical succession, gave a thickness estimate for the James Ross Island Volcanic Group of 305–1,530 m.

Over the southern part of Tabarin Peninsula the Bouguer anomalies (Fig. 38) indicate the presence of a gravity "low", the northern limit roughly coinciding with the postulated east-north-east to west-south-west fault. Although the station distribution over this northern area is thin, it does indicate a distinct change in gravity gradient which, although not conclusive of a faulted boundary, supports the geological hypothesis. The southern limit of the gravity "low" has not yet been delineated though the single gravity reading at Cape Burd does not maintain its presence there. Likewise, the eastern and western extents are unknown due to a lack of land and marine gravity data. The values on Eagle and Beak Islands suggest that the anomaly closes in this direction but, with the complex regional gravity gradients involved, it is difficult to arrive at any conclusions purely from the geophysical evidence.

No samples for density measurements have been collected by the author from the James Ross Island Volcanic Group nor from the fossiliferous Cretaceous sediments. The average density of numerous samples of sandstone, shale, limestone and chalk, as given by Clark (1966, p. 23–24), is about 2·3 g. cm.$^{-3}$. Bibby described the sediments of the Snow Hill Island Series as largely of uncemented sands and gravels with intercalated sandy clays and clays. This succession, which is unconformably overlain by the James Ross Island Volcanic Group, reaches a thickness of about 3,580 m. Insufficient knowledge is available concerning the relative thickness or densities of the lavas, tuffs and agglomerates of the James Ross Island Volcanic Group but from the exposures visited and from the examination of hand specimens it would appear that there is much unconsolidated material and therefore the majority of extrusive rocks present are unlikely to have a high density. This is certainly the case with the samples of vesicular olivine-basalt measured by Adie (1953) which gave values of 2·14–2·28 g. cm.$^{-3}$. Fleet's (1968) observations on equivalent rock types at the Seal Nunataks support a friable unconsolidated nature.

The gravity "low" associated with southern Tabarin Peninsula is calculated to lie within the range of 10–20 mgal depending on the form of regional gradient removed. Assuming the negative anomaly is caused by the juxtaposition of the less dense James Ross Island Volcanic Group with the Trinity Peninsula Series, then a simple two-dimensional Bouguer slab formula may be used to calculate the density contrast. If the maximum thickness of 1,530 m. is assumed for the volcanic rocks, the density of the James Ross Island Volcanic Group which would be required to account for an observed anomaly of 20 mgal would have to be of the order of 2·44 g. cm.$^{-3}$. However, neither the thickness of the volcanic rocks nor the magnitude of the anomaly are known with certainty, and therefore the calculated density must only be a first approximation. An alternative solution, particularly if the thickness and density are an overestimate, is that Cretaceous sediments underlie the volcanic rocks as observed on nearby James Ross Island.

2. *Duse Bay*

Using the data over Duse Bay, the free-air anomalies were contoured and are shown in Fig. 38 together with the Bouguer anomaly map constructed from the surrounding land gravity values. This illustrates a general negative gradient towards the axis of the peninsula and also shows the negative anomaly associated with the James Ross Island Volcanic Group rocks over Tabarin Peninsula, Beak Island and Eagle Island. Extrapolation and removal of the regional Bouguer anomaly gradient leaves a considerable negative anomaly over central Duse Bay. The anomaly was assumed to be entirely the result of bathymetry, with a density contrast of 1·64 g. cm.$^{-3}$.Therefore, interpretation yields minimum depths to bedrock as no allowances have been given for possible unconsolidated sea-floor sediments.

A three-dimensional gravity interpretation was necessary and in the first approximation the selected model shape was based on the known sea depths obtained by R.R.S. *Shackleton*. Unfortunately, Duse Bay is often covered by fast ice which does not allow easy access by ship, and therefore over the northern side of Duse Bay the bathymetry used initially was an extrapolation of the isobaths as contoured by Ashley (1962). The computer programme described earlier and based on the method of Talwani and Ewing (1960) made it evident that depths considerably greater than those recorded in southern Duse Bay were needed to account for the gravity "low" 7 km. east of View Point. A reverse procedure therefore was adopted, making use of the iterative three-dimensional programme.

By superposition of a chosen grid network over the free-air anomaly map, over 100 gravity values were interpolated and used in the iterative computation. Above 12 iterations the differences between re-computed depths was insignificant with only minor improvement in the cumulative error. Over Duse Bay an accuracy of 0·2 mgal was achieved between observed and calculated gravity values.

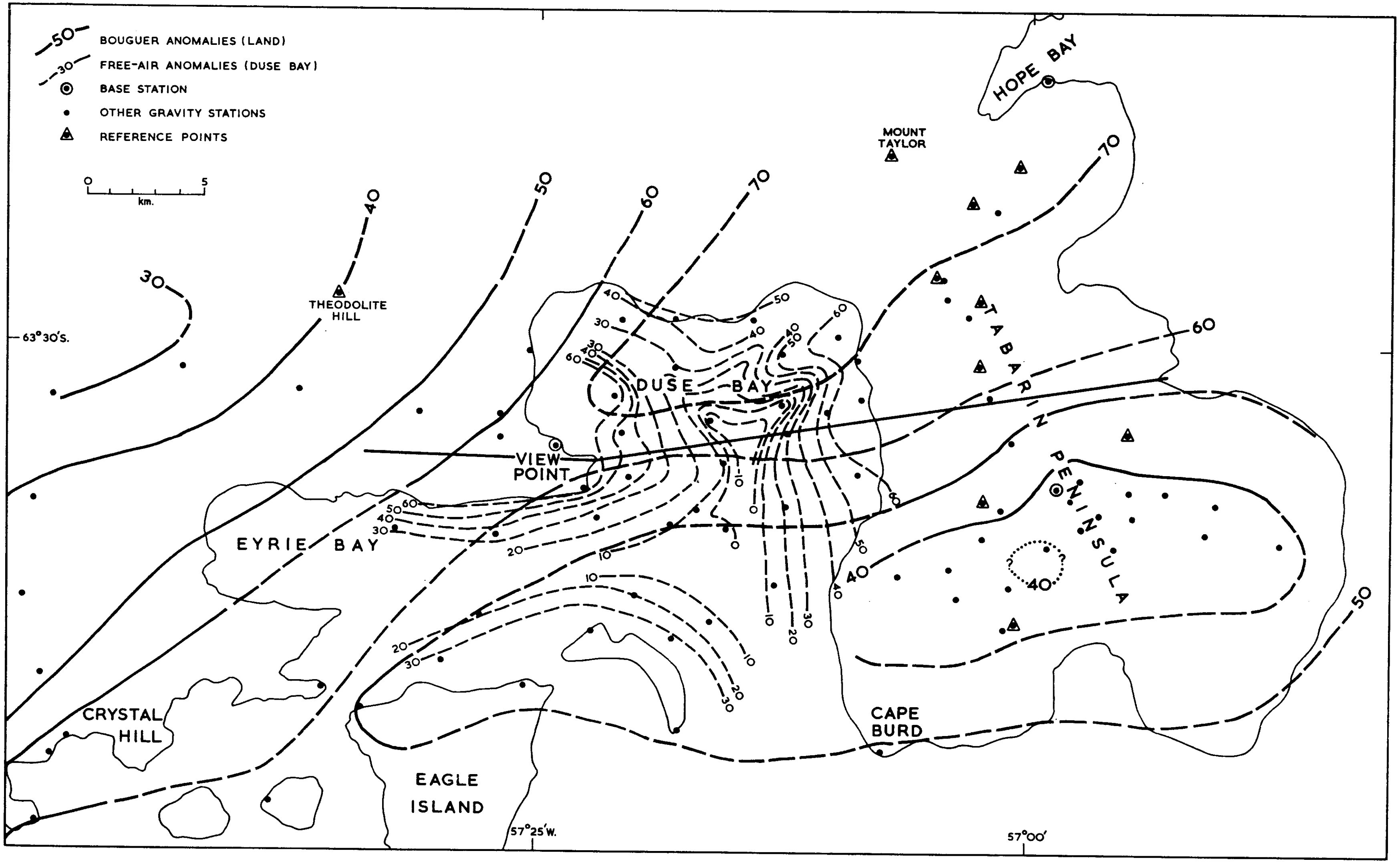

Figure 38

Bouguer and free-air anomaly map over Duse Bay and Tabarin Peninsula.

The three-dimensional form giving the closest fit is shown contoured in Fig. 39. Those points at which sea depths are known from ship soundings are also indicated and they show good agreement with the depths computed within the grid network. Provided the bedrock density remains constant at 2·67 g. cm.$^{-3}$, this coincidence of the calculated and observed sea depths is strongly suggestive that the bathymetry is

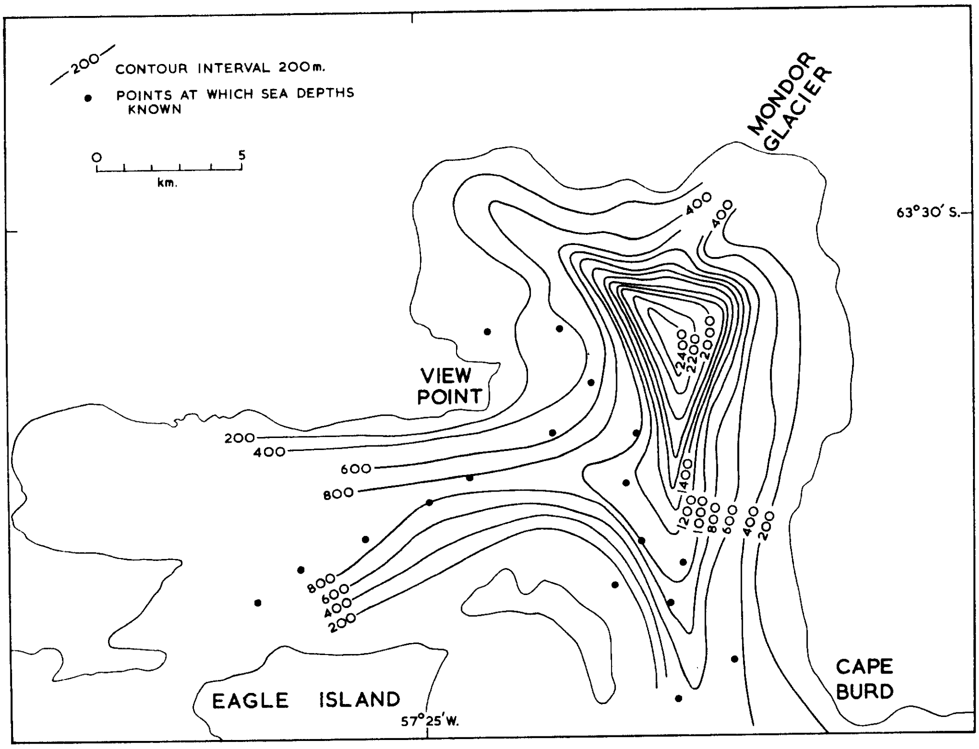

FIGURE 39

Three-dimensional interpretation for the bathymetry in Duse Bay.

primarily responsible for the anomalies (Fig. 40). It may also imply the absence of thick deposits of unconsolidated sediments.

In the southern part of Duse Bay the maximum recorded depth of 970 m. occurs at a point 3 km. north-east of Beak Island. This depth lies within a basin delineated by the 500 fathom [915 m.] isobath. Calculations from the free-air anomalies indicate that this basin extends and deepens to the north, reaching maximum depths of 2,400 m. Depths of 880 m. have been previously sounded in Prince Gustav Channel, and 6 km. south of Beak Island 765 m. has been recorded. On the west coast of the Antarctic Peninsula and within the continental shelf, areas considerably deeper than 500 m. have been measured. West of Marguerite Bay, 1,200 m. has been mapped (on a northerly projection of the line followed by George VI Sound), whereas between the mainland and the offshore islands there are depths of 915 m. Attention should also be drawn to three parallel north-west to south-east trending channels cut into the shelf edge 25 km. off the north-west coast of Trinity Peninsula. These features are about 11 km. in width between the 100 fathom [180 m.] line and attain a maximum observed depth of 870 m. The more easterly one terminates only a few kilometres from Cape Siffrey and runs almost as an extension of Antarctic Sound. The method of formation of these features is unknown though glacial overdeepening with or without associated lines of weakness is a possibility. To approach the maximum depths calculated in Duse Bay, however, one has to look to the deep water of Bransfield Strait on the South Shetland Islands trench where isobaths greater than 1,000 fathoms [1,800 m.] form part of major physiographical features.

On the Filchner Ice Shelf, Behrendt (1962) described seismic reflection methods which indicated that the trough underlying the eastern section of the ice shelf reaches depths of over 1,500 m. It can be traced from the head of the ice shelf out on to the continental shelf of the Weddell Sea. Here a glacial origin has been proposed for the trough with an estimate that an additional 400 m. of ice would cause the necessary grounding and erosion.

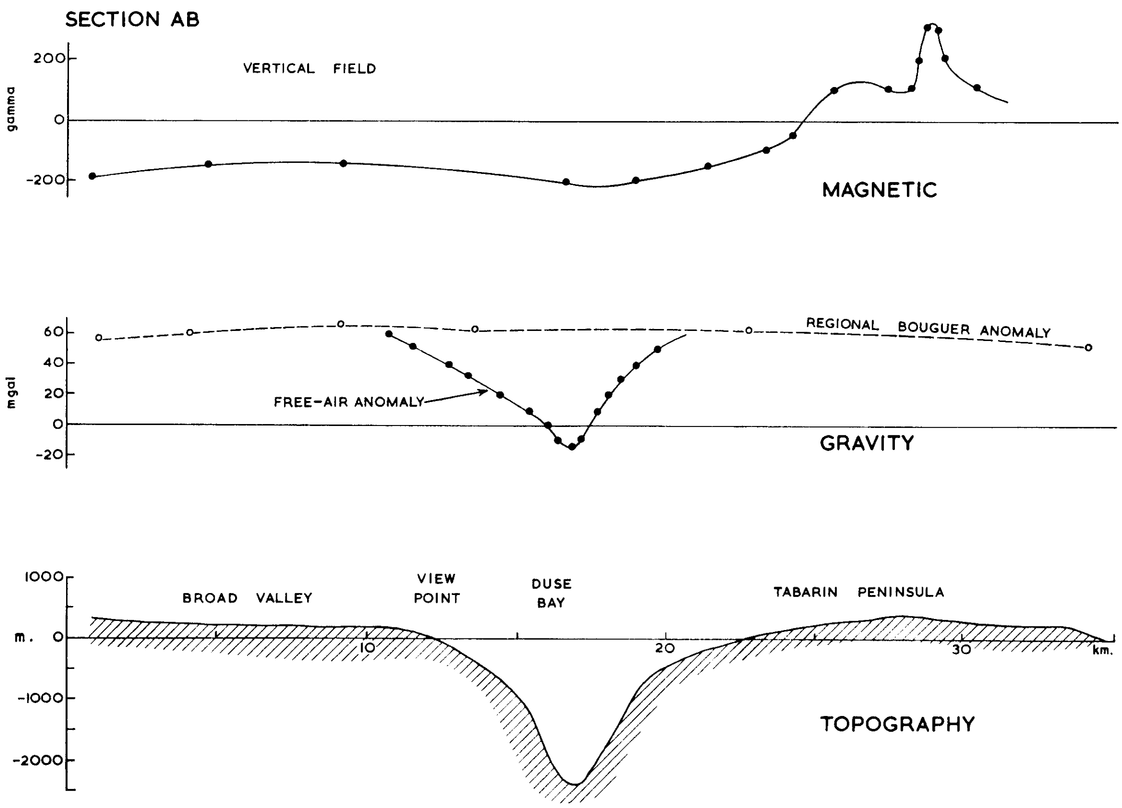

FIGURE 40
Magnetic, free-air gravity anomaly and calculated topographical profiles across Duse Bay.

Glacial erosion as the main agent responsible for the formation of the Duse Bay basin is at first sight questionable due to the depths involved. However, Koerner (1964) stated that since the glacial maximum at least 300 m. of ice have been removed from the area. Glacial striae at View Point indicate that the ice moving up Prince Gustav Channel deflected that flowing over the View Point area and was of a sufficient thickness to move independently of local surface features. Duse Bay itself would have acted as a natural accumulation area for the ice from Mondor Glacier, Mott Snowfield and the Laclavère Plateau. With such increased ice thicknesses there could have been grounding and erosion to depths comparable to those estimated. Bibby (1966) has also held glacial action responsible for the sea depths between Tabarin Peninsula and Vega Island.

3. *Prince Gustav Channel*

Prince Gustav Channel is regularly covered by sea ice, which may break up by variable amounts during the summer seasons. The degree to which it disperses is most noticeable in the northern part of the channel for farther to the south it is less influenced by external factors, as well as showing a gradual mergence with the ice shelf in the vicinity of Persson Island. Although the sea ice did permit travel during the 1964 field season, it was not possible to obtain gravity readings over the channel north of Cape Obelisk due to increasing tidal oscillations. Where possible, land gravity stations were occupied along the shores of Prince Gustav Channel and these indicate that the Bouguer anomaly is slightly more positive on the western side. Midway along Prince Gustav Channel the Trinity Peninsula Series is exposed on Long Island and a gravity reading here follows the pattern observed on the mainland. Those measurements taken on Eagle, Beak and Tail Islands, however, suggest an easterly decrease in gravity value, which from a regional position indicates the reverse of the gravity gradient found elsewhere over Graham Land. The explanation for this reversal is probably due to the less dense volcanic rocks and unconsolidated sediments on James Ross Island. Any major structure affecting the geological relationships between the opposite shores of Prince Gustav Channel is therefore likely to occur east of Long Island.

At the southern end of Prince Gustav Channel, two parallel gravity traverses (Fig. 41) were surveyed across the ice shelf between Cape Longing and James Ross Island. For the original data reduction the sea depth of 500 m. was again used in the Bouguer correction as no other information was available. When the Bouguer anomaly map of the area was completed, it was evident that a gravity "low" was associated with the channel and this was attributed to bathymetry. An iterative two-dimensional computer interpretation was applied, assuming a constant density difference of 1·64 g. cm.$^{-3}$ and a background regional value of

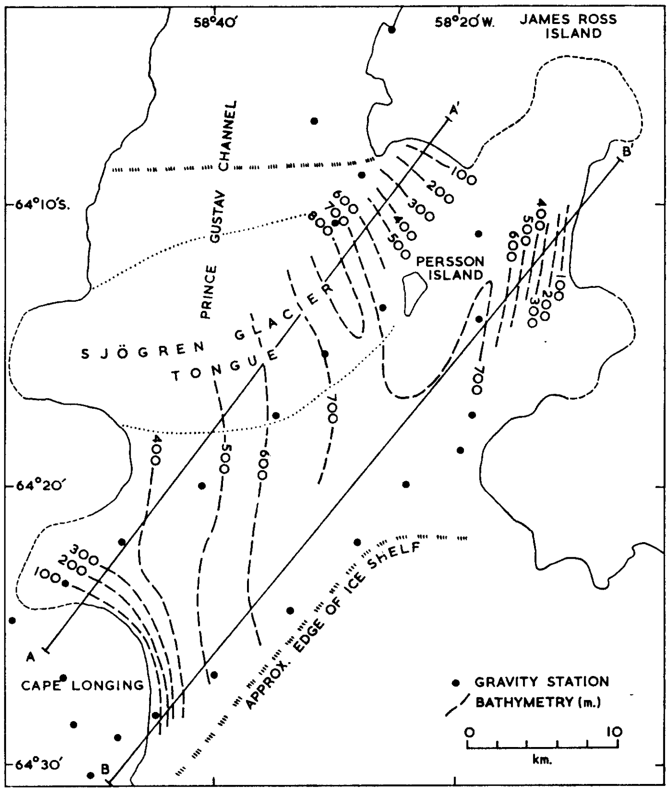

FIGURE 41

Sketch map of the southern end of Prince Gustav Channel, showing lines of interpreted profiles with the resulting bathymetry.

+ 50 mgal. The results of the interpretation are shown in Fig. 42 with maximum depths of 820 and 720 m. being estimated in Prince Gustav Channel and Röhss Bay. During periods when Prince Gustav Channel was relatively ice-free, several sea depths were sounded as far as Alectoria Island. Depths of 670 m. were found off Pitt Point, 880 m. between Red Island and Botany Bay, and 675 m. near Vortex Island. These depths are thus consistent with those interpreted at the southern end of the channel and with those inlets on the Larsen Ice Shelf where glacial overdeepening is suggested.

Koerner (1964) suggested that Prince Gustav Channel acted as an outlet stream for ice from the east coast of the mainland and from the west coast of James Ross Island. Persson Island is estimated to have been covered to a height of at least 130 m. above present sea-level and, as previously stated, some areas indicate that more than 300 m. of ice have been removed since the glacial maximum. Ice thicknesses of this order and the associated glacial excavation could account for the observed and calculated bedrock depths.

4. *Cape Disappointment–Starbuck Glacier area*

Both Fleet (1965) and Kennett (1965*c*) independently assessed the surface elevation of the Larsen Ice Shelf in the Cape Disappointment–Starbuck Glacier area. Three north–south rifts were observed (Fleet, 1965) in this area but only the westernmost was visited. Here, Fleet identified brash ice in the bottom of the rift and he used this to conclude a surface elevation of 25 m. This value was considerably less than the 66 ± 12 m. derived by Kennett from aneroid barometer measurements taken at nearby localities, and this

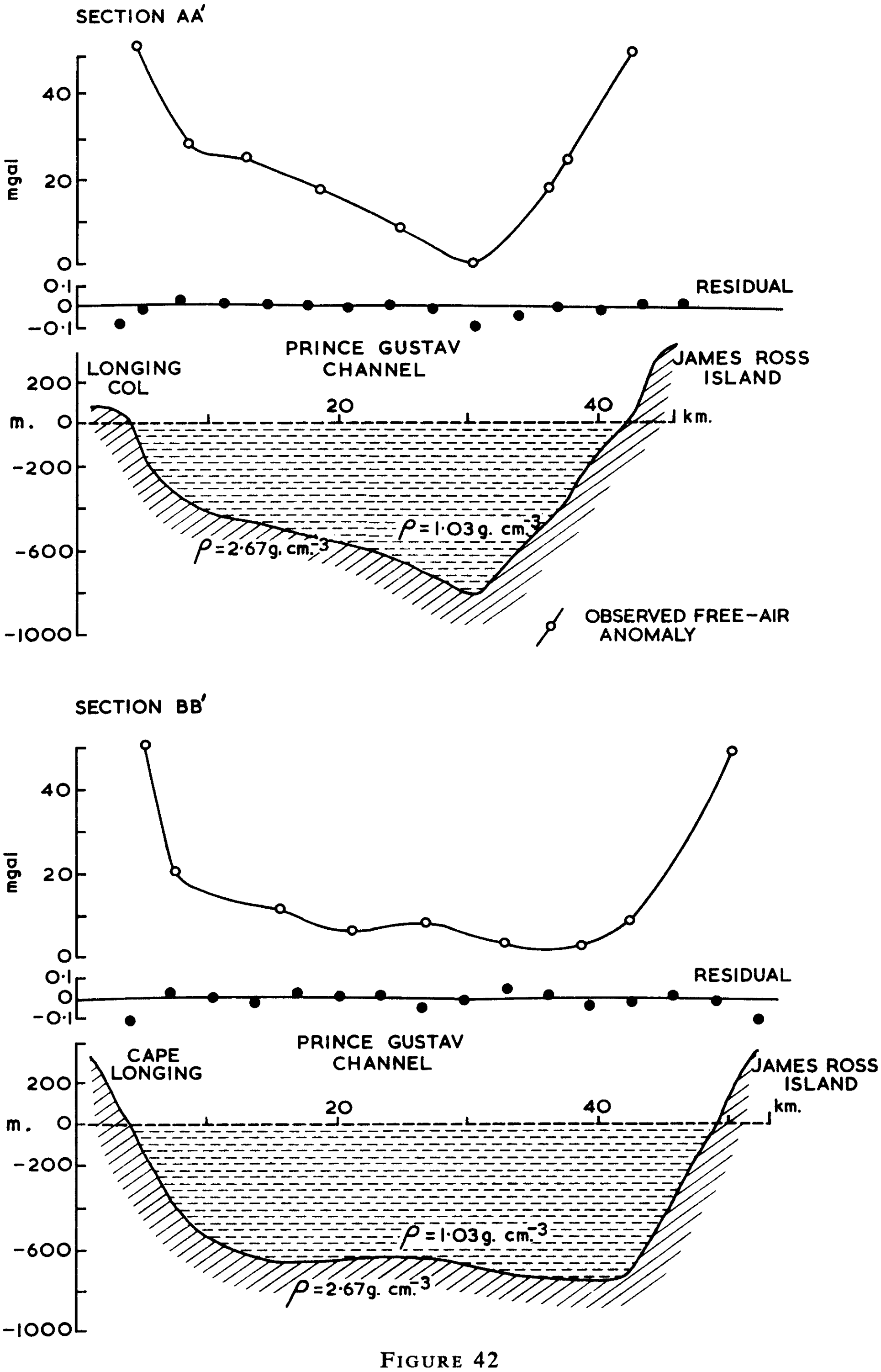

FIGURE 42

Two sections (Fig. 41) across Prince Gustav Channel, showing the calculated bedrock profile.

led Kennett to regard Fleet's value to be of a local nature and associated with the rifting. Heights of 30 m. a.s.l. have been observed at the edge of the Larsen Ice Shelf 75 km. to the east (D.O.S. 1:200,000 sheets W6558 and W6560). If the value of 25 m. was correct, this would involve surface gradients between the mainland and the ice edge opposing that found from other ice-shelf profiles and contrary to the hydrostatic equilibrium principle.

Analysis of the gravity data from this area can possibly provide one solution to the discrepancy though the relatively complex regional field in the vicinity does not permit an accurate quantitative investigation. Fig. 43 shows the free-air anomalies rising towards Cape Disappointment with the steepest gradient occurring in the area between the rift and the mainland. Provided there is no lateral change in rock density,

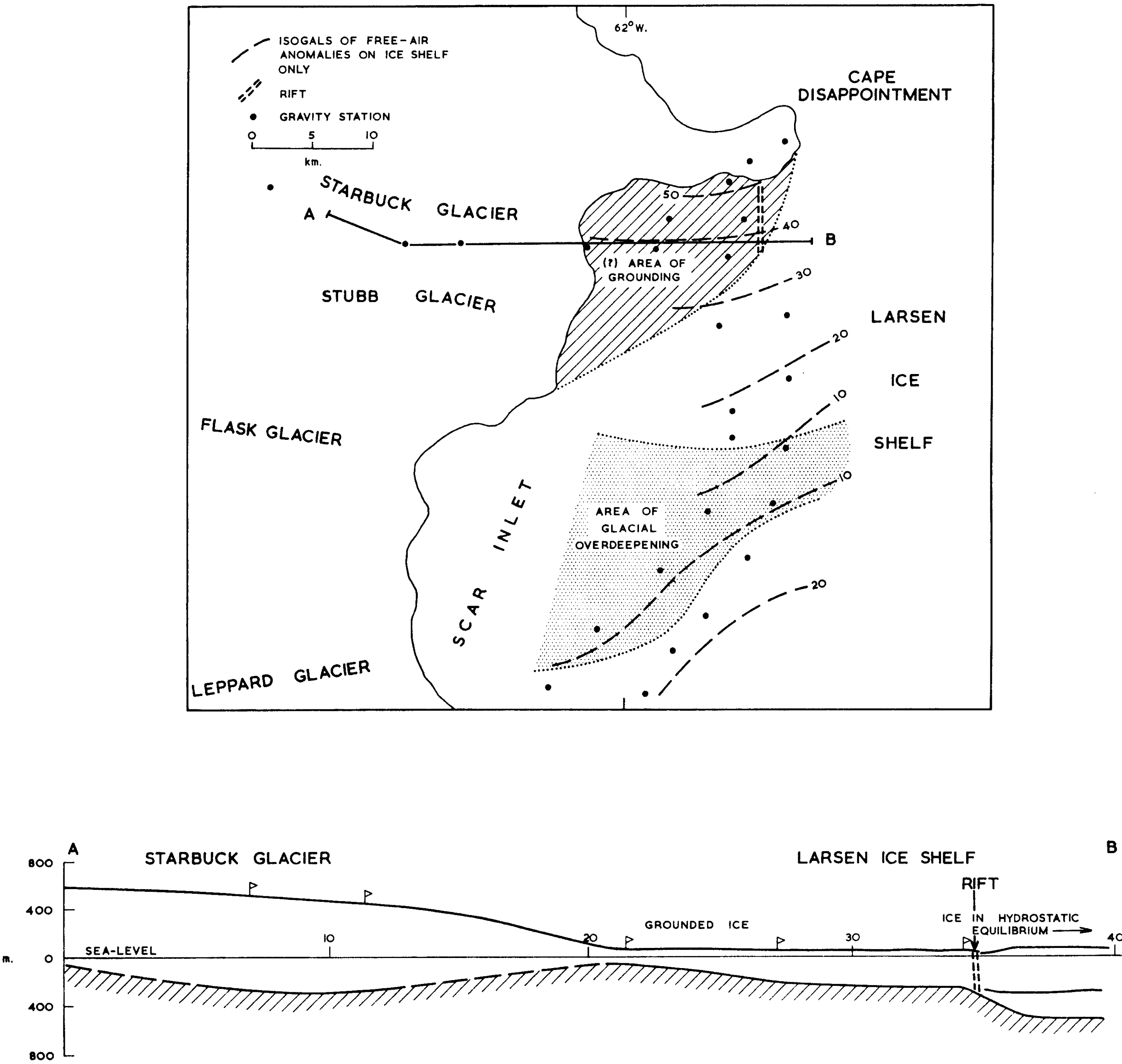

FIGURE 43

Free-air gravity anomalies off Starbuck Glacier. Profile AB is a two-dimensional interpretation of the gravity anomalies.

then the free-air anomalies may be used as a first approximation to calculate variations in bedrock topography. The low free-air anomalies farther towards the south (Fig. 43) have thus been interpreted as due to depressions within the bedrock and caused by glacial overdeepening by Flask and Leppard Glaciers. Gravity interpretation over Starbuck Glacier by Kennett (1966*c*) indicated that the bedrock 10–15 km. west of the "coastline" lies at a depth of about 250 m. below sea-level. Estimates by the author on grounded-ice models suggest that at the junction of the Starbuck Glacier ice with the ice shelf the bedrock lies at a depth of about 80 m., which is somewhat less than that estimated by Kennett through the conjectural extrapolation of the gradient of the glacier bed. 6 km. to the east of the junction, the author has calculated the bedrock depth to be at 200–240 m. and on the western lip of the considered rift at 230 m. below sea-level. A further 4 km. to the south of the last locality the estimated depth shows an increase to 320 m. It would therefore appear that the longitudinal profile beneath Starbuck Glacier shows, as Kennett found in Stubb Glacier, a rise in bedrock elevation towards the east.

From the calculations, a possible explanation of the different estimates in the surface elevation may be that the rift marks the position of the junction between the land ice and the ice shelf (Fig. 43). Thiel and Ostenso (1961), working on the Ross Ice Shelf, observed that the junction was characterized by a "valley", the depth of which appeared proportional to the slope of the incoming mainland ice. Robin (1958) reported a similar feature from the Maudheim area, where consideration was also given to the stresses involved due to the periodic bending through tidal motion. Whether or not the thinning is a direct result of tidal bending or an indirect result of hydrostatic equilibrium processes, the "necking" phenomenon on ice shelves is a recognizable feature. Fleet's observations could have been made in such a depressed area and therefore of limited extent.

The fracturing, however, may not necessarily be a direct consequence of the tidal motion but it may in part be due to the proximity of Cape Disappointment producing a differential rate of ice flow. Fleet has brought attention to another rift immediately to the north of Cape Disappointment caused by the stress between rapidly moving ice and the stagnant ice adjacent to the cape. It was suggested that the two parallel structures to the east of the one studied are probably older, now inactive rifts, which originally formed farther to the west.

Elsewhere along the east coast of Graham Land increased free-air anomalies have been found adjacent to the mainland. Examples of these are at Cape Northrop, Cape Choyce, Cape Worsley and in the Cape Longing area, and all have been interpreted as areas of a rise in the bedrock with or without associated grounding. In some instances, crevassing and surface irregularities accompany the grounding but in others the transgression is devoid of any major surface indication.

5. *Trail Inlet*

The area around Trail Inlet has few rock outcrops suitable for gravity stations and therefore with the exception of those stations considered to be on or near to bedrock, i.e. Whirlwind Inlet and Three Slice Nunatak, the regional gradient removed was based mainly on extrapolation of the data to the north. The residual gravity anomaly map (Fig. 44) discloses that a gravity "low" covers the central area of the inlet. In the original data reduction, estimated depths of 550–650 m. to bedrock were used and these were governed by ice-thickness soundings. East of Trail Inlet, the depths to bedrock used beneath the Larsen Ice Shelf reverted to the estimated 500 m. Although the eastern end of Joerg Peninsula has an associated positive anomaly which may slightly affect the values in the southern central part of the inlet, it has again been assumed that depths to bedrock alone are responsible for the gravity anomalies.

Interpretation was carried out on five profiles (Fig. 45), four of which were at right-angles to the direction of ice flow and the other a longitudinal profile east from Bills Gulch. In any two-dimensional computation one must assume the body extends to infinity in the third dimension. For the longitudinal profile, however, the enclosing walls influence the two-dimensional calculation and the depths given were thus an underestimate of the true depths. Some indication of the order of error can be obtained where the profiles intersect, and as a consequence the longitudinal profile depths were increased proportionally.

Fig. 45 illustrates the calculated depths to bedrock using densities of $2 \cdot 67$ g. cm.$^{-3}$ for rock and $1 \cdot 03$ g. cm.$^{-3}$ for sea-water. The ice-thickness values are those determined from the radio echo-sounding traverses (Renner, 1969). From the sections, Trail Inlet can be seen as an area of over-deepening with depths below sea-level as great as 900 m. At the transgression of the Trail Inlet ice on to that of the Larsen Ice Shelf the bedrock rises to around 500 m. below sea-level with a corresponding "necking" of the ice. This reduction in ice thickness can be explained by the emergence of the inlet ice into the broader front of the Larsen Ice Shelf, and this is also due to the release from the inlet where the ice movement is restricted by Three Slice Nunatak and north-eastern Joerg Peninsula. Increased ice thicknesses with glacial overdeepening is an obvious cause for the greater depths and has support from geomorphological observations within this area. Marsh and Stubbs (1969) suggested that a significant erosional level can be recognized in this area at 457

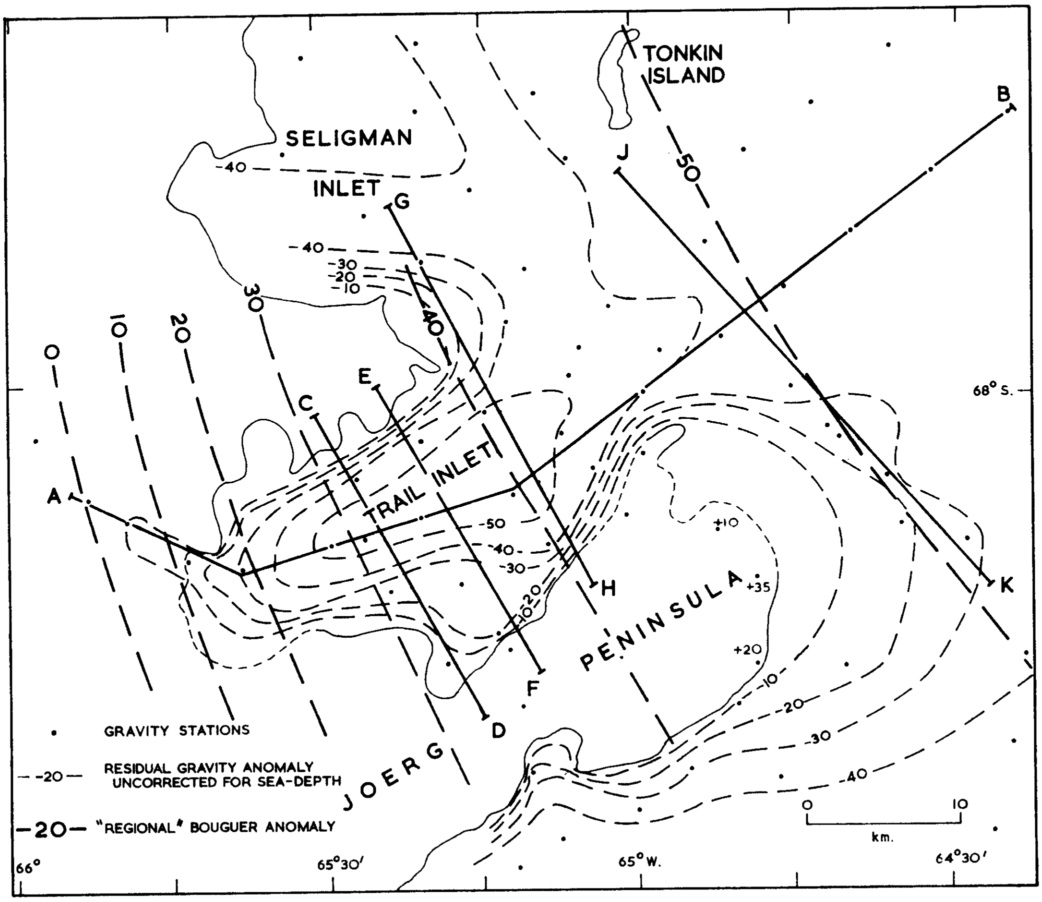

FIGURE 44
Residual gravity anomalies over the Trail Inlet–Joerg Peninsula area.

m., and that Three Slice Nunatak (503 m.) itself is a crag-and-tail feature orientated parallel to the flow of ice from Daspit Glacier, which provides the largest influx of ice into Trail Inlet.

6. *Joerg Peninsula area*

The Bouguer anomaly "high" over the eastern section of Joerg Peninsula warrants further comment as the removal of the regional gradient yields maximum positive residual anomalies approaching 10, 35 and 20 mgal at the three easternmost stations (Fig. 44). The anomalies of 35 and 20 mgal are surprisingly large and therefore factors which could affect their validity were at first queried.

In both instances, the stations were re-occupied during the relaying of equipment. After drift allowance, consistent gravity values were obtained. As no aneroid barometer readings were taken along this section of the traverse, station elevations were obtained from the D.O.S. maps of the area, 1:200,000 sheet W6864 and 1:500,000 sheet 13. The elevations of both stations were of the order of 180 m. but, even assuming the station to be at ice-shelf level, there would still remain a positive anomaly at the locality of the present maximum. Over the eastern tip of Joerg Peninsula, ice thicknesses of 195–225 m. have been determined (Smith, 1972). Provided flight traverses have been correctly located, then the stations being considered would be on an all-ice column to sea-level. Although this differs from the ice–rock distribution used in data reduction, it would tend to make the anomalies even larger.

Further elevation and the ice-thickness data are therefore essential before a more accurate Bouguer anomaly map can be prepared but on present available evidence it would appear that eastern Joerg Peninsula has an associated positive Bouguer anomaly with a maximum value which could approach 35 mgal. Consideration of elevation accuracy, topography and ice thicknesses leads the author to suggest that the true anomaly is of the order of + 25 mgal.

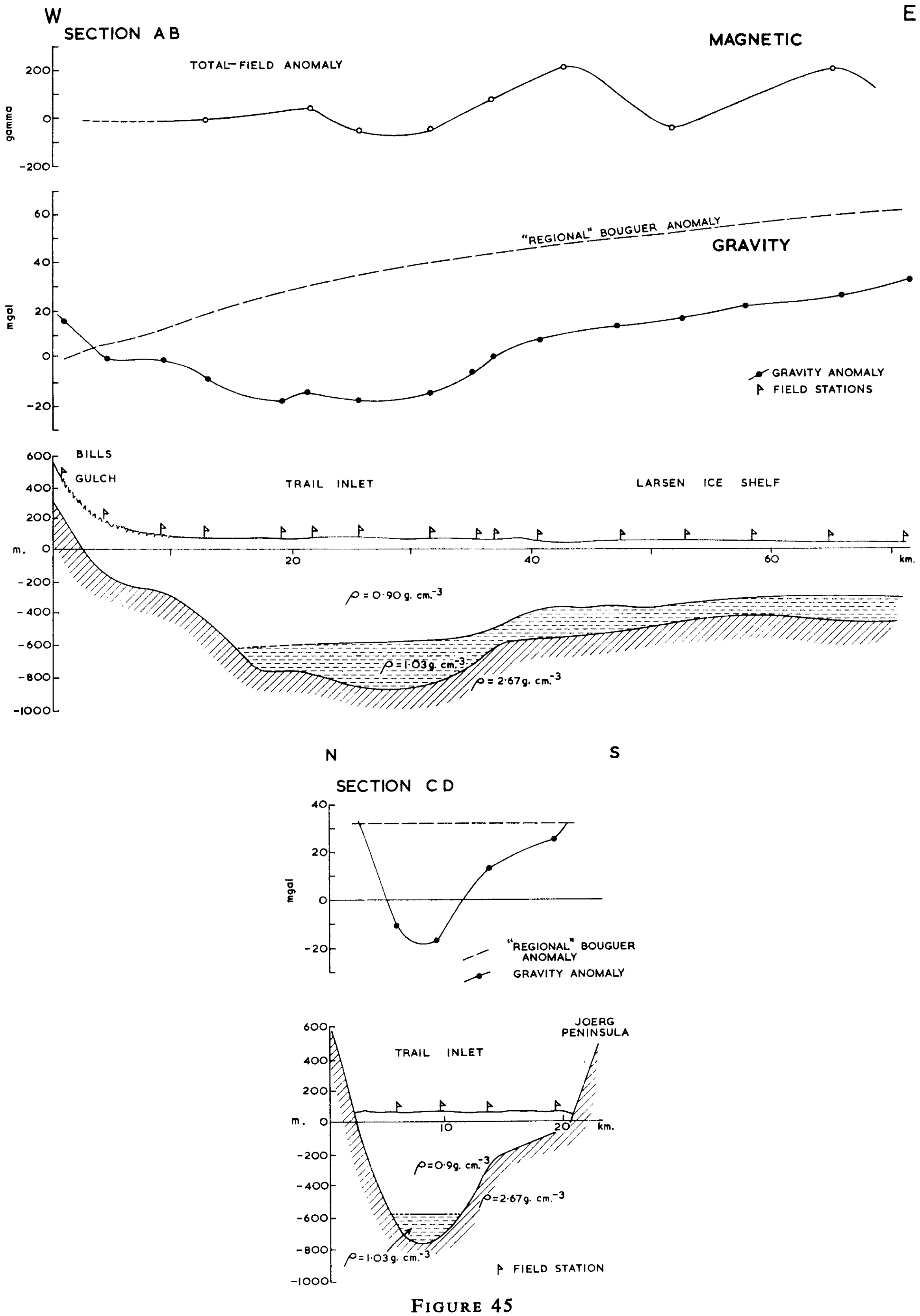

FIGURE 45

Longitudinal and transverse bedrock profiles calculated from the free-air gravity anomalies in Trail Inlet. The lines of section are shown in Fig. 44.

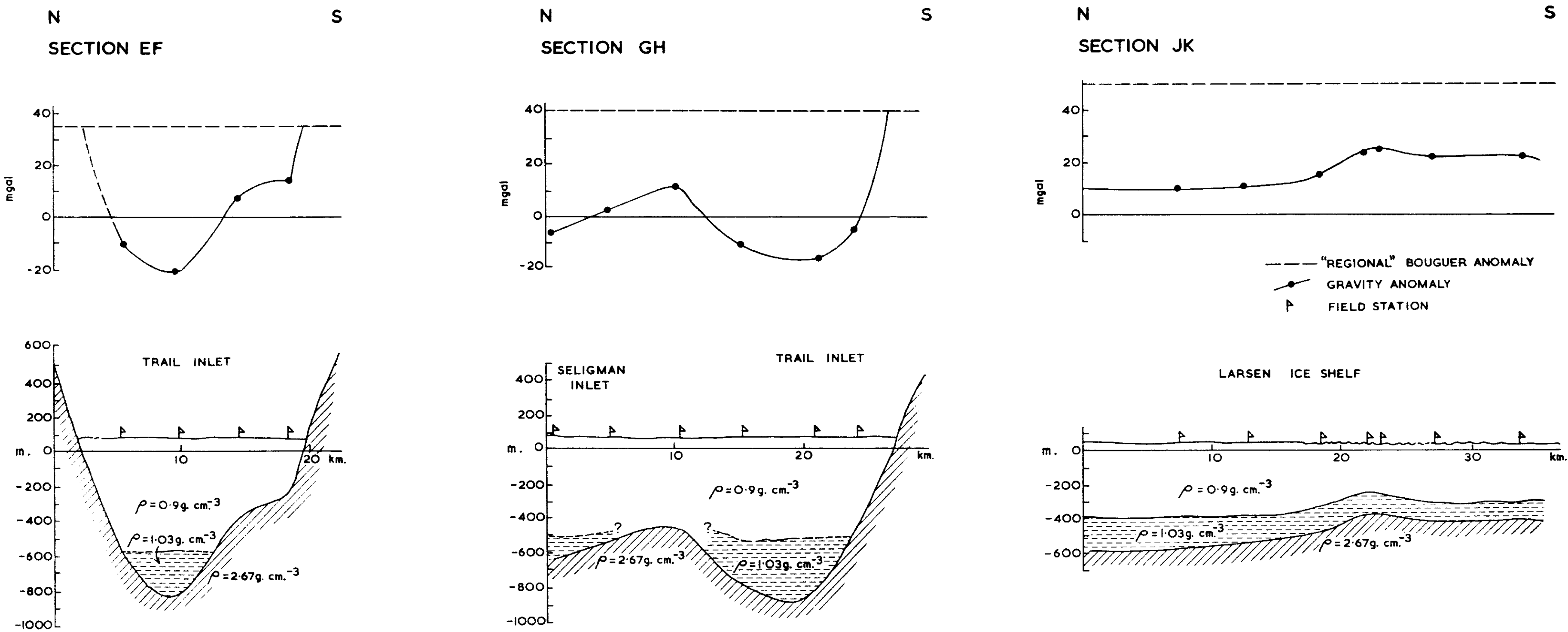

FIGURE 45—*continued*

Off the eastern end of Joerg Peninsula, the Larsen Ice Shelf is represented by a considerable belt of highly disorganized ice. This belt extends 10 km. to the east and, from the presence of a positive gravity anomaly, it has been interpreted as due to a rise in the bedrock. It is possible, however, that the positive anomaly is also partially associated with the body causing the eastern Joerg Peninsula land anomaly. For the present, the interpretation of the land anomaly only is considered. The two gravity stations occupied in the pass between Solberg Inlet and Trail Inlet suggest that the gravity "high" over eastern Joerg Peninsula does not extend west of this feature. At the station midway through the pass, assuming an all-ice column to sea-level, the maximum anomaly could only be $+5$ mgal, unless of course the ice does extend below sea-level. However, field physiographical observations suggest that this is not so. The station immediately to the north of the pass is located as a "land" site at ice-shelf level. Correction for an ice column to sea-level and the removal of the regional gradient leaves a residual anomaly of -13 mgal, which on a simple two-dimensional Bouguer slab interpretation of a grounded-ice model suggests ice to 180 m. below sea-level.

The rocks of the northern side of the Joerg Peninsula area have been described by Stubbs (1968) and they comprise gneisses, acid to basic intrusive rocks, dykes and sediments. Density measurements, completed for rock types within this area, are given in Table IX. The gneisses have an average density of 2·63 g. cm.$^{-3}$, the acid intrusive rocks 2·62 g. cm.$^{-3}$, the sediments 2·74 g. cm.$^{-3}$, and one sample of basic dyke, an amphibolized dolerite, 2·86 g. cm.$^{-3}$. The latter is not considered extensive enough to be the cause of the anomaly. Unfortunately, instrumental difficulties curtailed the total-field magnetic surveys over this area, but it would appear (Fig. 59) that the eastern end of Joerg Peninsula is associated with a magnetic feature which has so far only been recorded as a negative anomaly of 300 gamma. The geophysical evidence to date implies that eastern Joerg Peninsula is underlain by sediments which, at Three Slice Nunatak, have been correlated with the Trinity Peninsula Series (Adie, 1957).

A detailed interpretation is not justified because of the lack of data control but an estimate of a minimum thickness can be obtained from the maximum Bouguer anomaly. Using a density contrast of 0·07 g. cm.$^{-3}$ and an anomaly of 25 mgal, approximately 8,000 m. of sediments are required. Although the Trinity Peninsula Series has been recorded in greater thicknesses in the Prince Gustav Channel area, no such thicknesses have been found over Joerg Peninsula. Stubbs has recorded a minimum of 1,000 m. of sediments in western Joerg Peninsula but south of Three Slice Nunatak, although these deposits are quite extensive, no thickness measurements are available. M. R. A. Thomson (personal communication) has observed sediments forming Rock Pile Peaks (lat. 68° 25′ S., long. 65° 09′ W.) to the east of the 1,000 m. summit. Neither age correlation nor thickness determination were possible at the time but it did appear that there exists a situation similar to that at Joerg Peninsula in that there is structural control between the sediments and the outcrops to the west. Adie (1957) has shown this boundary as a faulted one in what appears to be a continuous faulted zone from Hearst Island north to Mobiloil Inlet.

IV. INTERPRETATION OF AREAS ASSOCIATED WITH BOTH GRAVITY AND MAGNETIC ANOMALIES

A. Seal Nunataks

The Seal Nunataks are a group of 16 volcanic islands projecting through the Larsen Ice Shelf. These nunataks, consisting of agglomerates and weathered vesicular olivine-basalt lavas (Fleet, 1968), are often cut by olivine-basalt dykes which trend north-west to south-east or west-south-west to east-north-east. Along their crests they parallel the trend of the respective host nunataks, thus suggesting an origin through fissure eruptions. Recently completed K-Ar determinations by Rex (1971) have given an age of less than 1 m. yr., which is younger than that obtained from the field relationships (Adie, 1964*a*) by which they have been correlated with the mid-Tertiary James Ross Island Volcanic Group. Within this group of volcanic nunataks, Upper Cretaceous sediments have been found at Cape Marsh on Robertson Island and also at the more distant Pedersen Nunatak. Total- and vertical-field magnetic surveys have been undertaken over this area, and there are also several gravity traverses.

1. *Interpretation of the gravity survey*

A 70 km. gravity traverse, with a station spacing of 5–6 km., was run eastward from Drygalski Glacier to within 15 km. of the ice-shelf edge at Christensen Nunatak. This traverse intersected at Larsen Nunatak that of the earlier north–south traverse passing through the Seal Nunataks. The Bouguer anomaly map (Davey and Renner, 1969) showed a gravity "high" coincident with the Seal Nunataks which may be explained by the use of incorrect bathymetry in the original Bouguer reduction. Assuming a maximum density contrast of 1·64 g. cm.$^{-3}$ between bedrock and sea-water, then a difference of 1 mgal represents 14·6 m. variation in bedrock topography. This density contrast was applied in the two-dimensional iterative gravity programme on the condition that the anomaly is due solely to bathymetric variation.

Fig. 46 gives the profile resolved after eight iterations and to an accuracy of 0·1 mgal. It distinctly shows

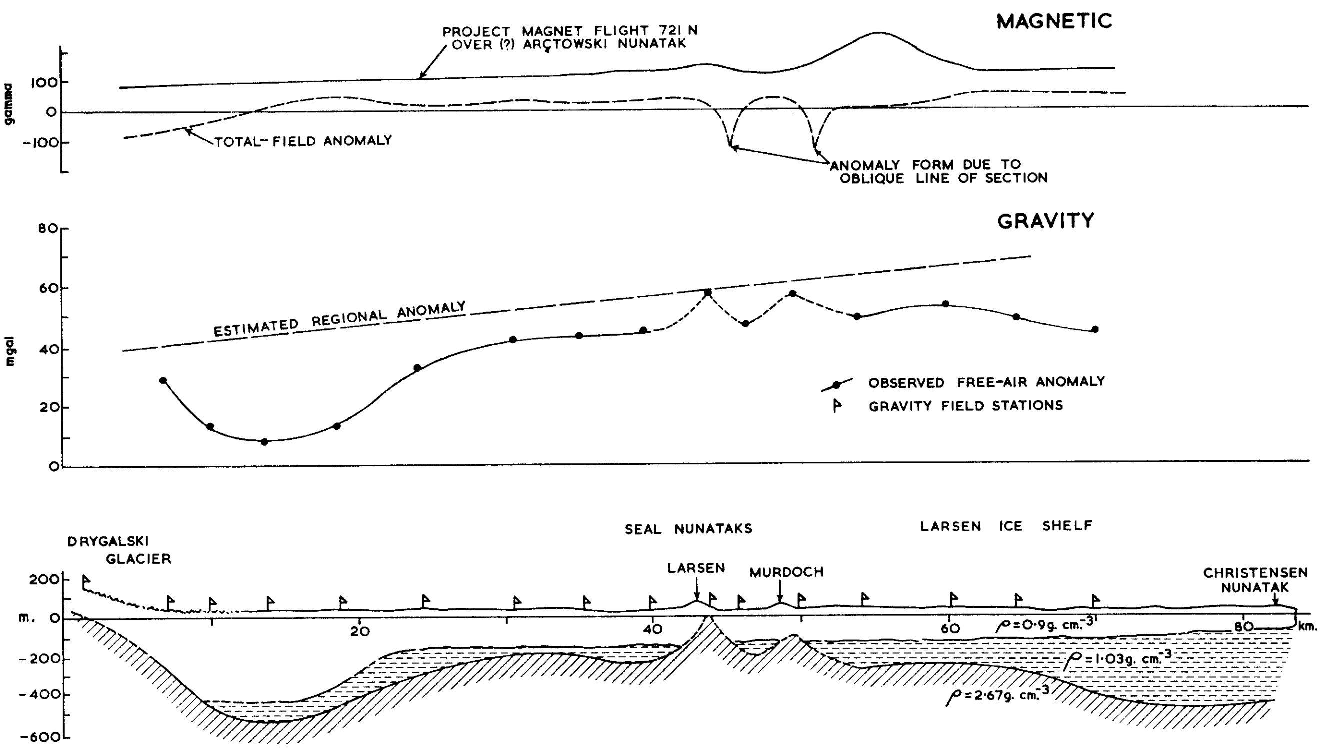

FIGURE 46

West–east gravity and magnetic profiles across the Larsen Ice Shelf from Drygalski Glacier to the Seal Nunataks. Bedrock elevations are calculated from the residual gravity anomalies.

that there is a general rise in bedrock beneath the nunataks which is reflected in the grounding of the ice shelf within a few kilometres of individual nunataks. The slight offset (Fig. 46) of the summits of Larsen and Murdoch Nunataks from the sub-surface topographical high is due to the siting of the gravity stations. The stations were close to the rock outcrops and, as the line of the gravity profiles does not traverse the peaks themselves, an apparent displacement appears. It was of interest to note the presence of a crevasse 9 km. south of the nunataks; this was observed at a similar latitude on both of the traverse lines, here 10 km. apart. Although it was less than 0·5 m. in width, it ran approximately east–west and at right-angles to the line of travel; it may therefore indicate a demarcation of differential flow between the slower ice whose movement is impaired by the Seal Nunataks and that forming the remainder of the ice shelf. On the traverse north of the Seal Nunataks no equivalent feature was encountered though several crevasses were observed within a few kilometres of "Base Teniente Matienzo".

At the eastern end of the profile, sea depths of 500 m. have been postulated by the author though in the Christensen Nunatak–Robertson Island area local depths can be extremely variable. Within a short distance of the ice-shelf edge at this point, soundings of 275, 46 and 92 m. were taken (p. 12). The ice depths

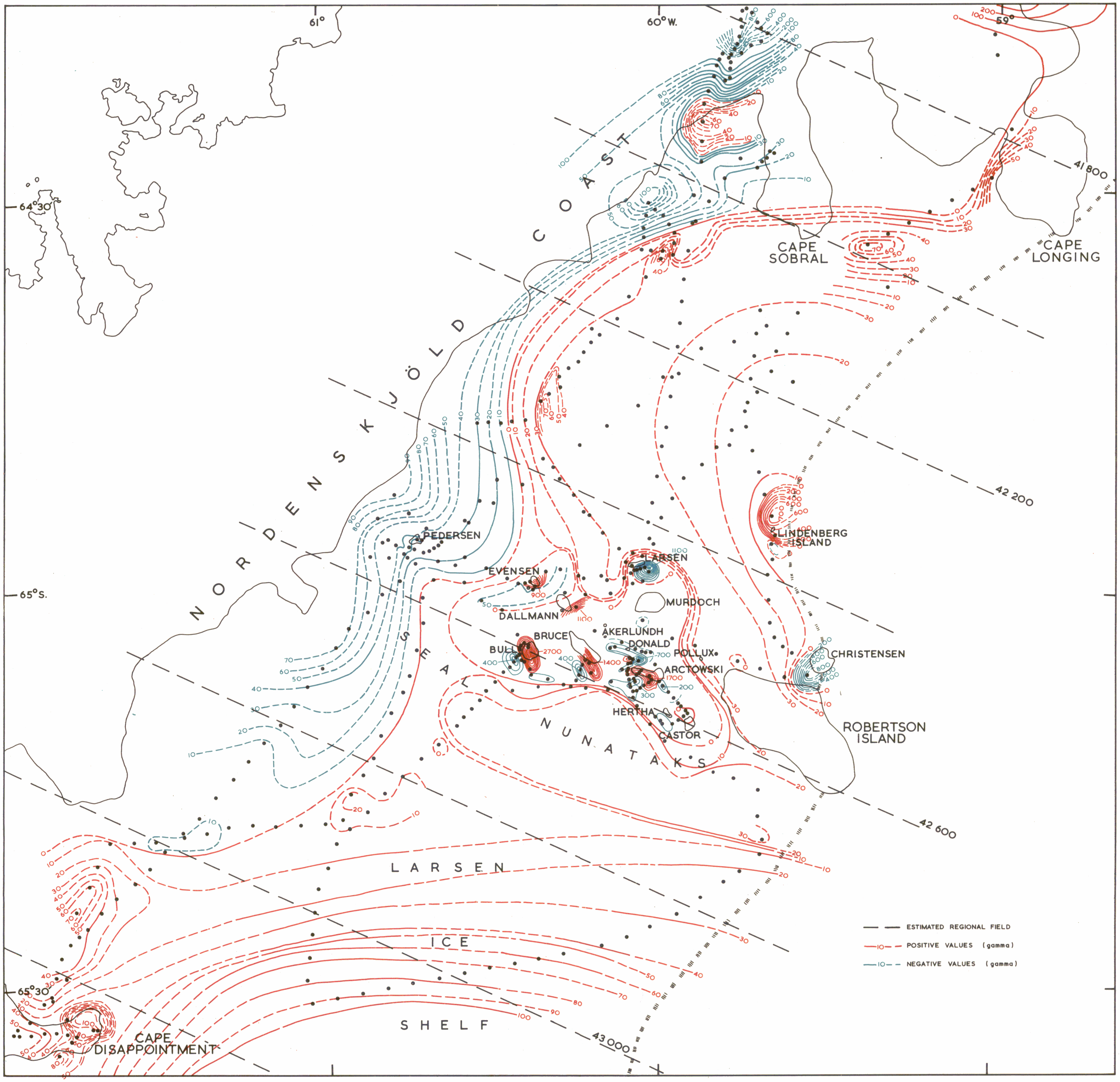

FIGURE 47

Total-field magnetic anomalies over the Seal Nunataks.

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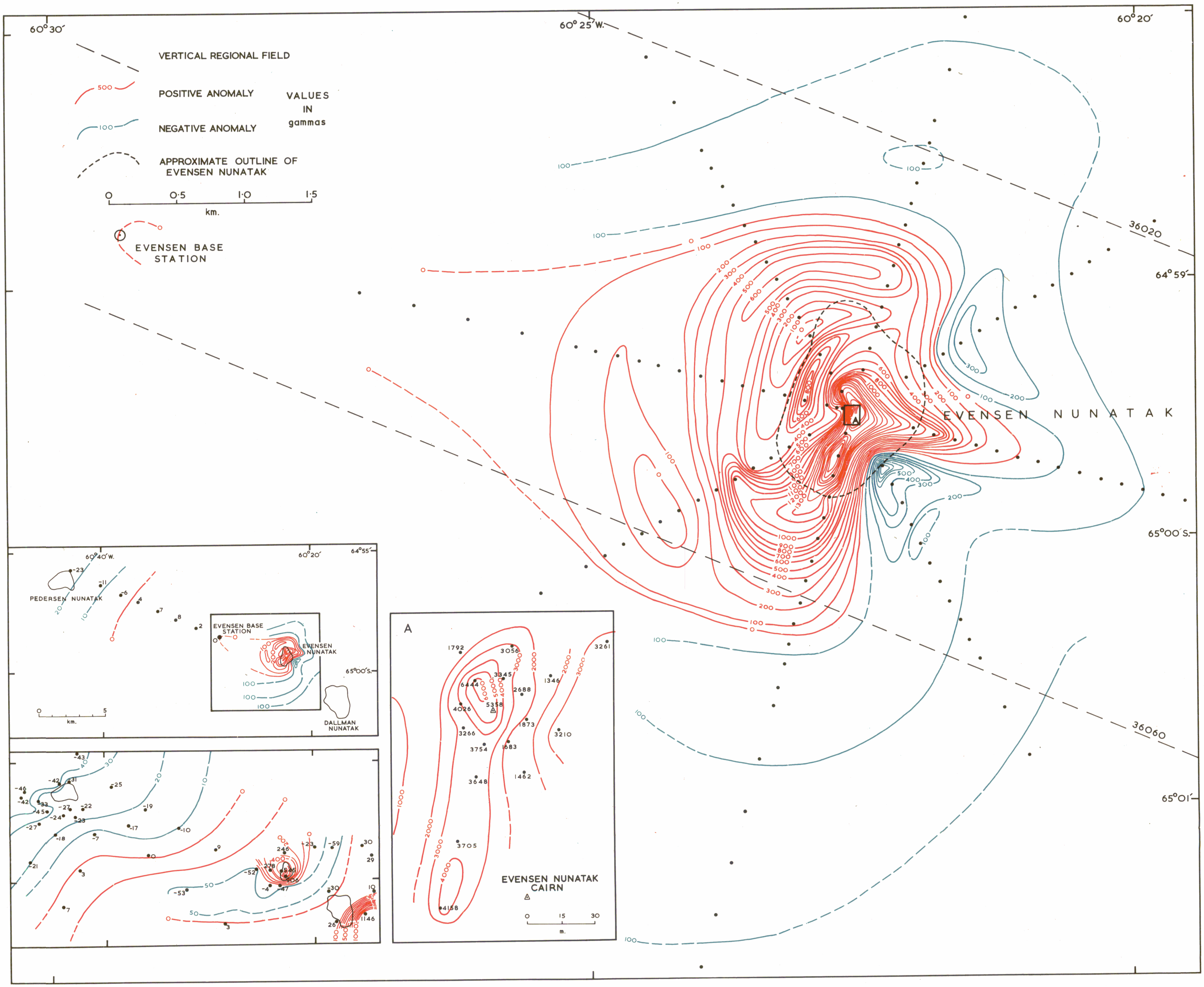

FIGURE 48

Detailed map of the residual vertical-field magnetic anomaly over Evensen Nunatak.

drawn on the sections are entirely speculative and are based on the surface-elevation values at Larsen Nunatak and the ice-shelf edge.

West of the nunataks, irregular surface terrain and heavy crevassing are noticeable off Drygalski Glacier but the absence of similar features to the north could be attributed to the severe nature of the adjacent coastline with no direct influx of glacier ice. Within the congested area the gravity data suggest the ice shelf to be grounded but beyond 8 km. from the coast glacial overdeepening is implied outward from Drygalski Glacier.

2. *Interpretation of the magnetic survey*

Regional total-field magnetic traverses (Fig. 47) have been run over the Seal Nunataks area and a more detailed vertical-field survey has been completed over Evensen Nunatak. It is apparent that the Seal Nunataks form a distinct magnetic feature on the otherwise magnetically quiet Larsen Ice Shelf. Along the Nordenskjöld Coast, the isogams parallel the mainland and decrease towards its axis, a characteristic noted along the length of Graham Land. The slight easterly swing of the contours north of Pedersen Nunatak may be due to glacial overdeepening, whilst the small (100 gamma) but gradual increase in magnetic intensity towards the Seal Nunataks could be partially due to a rise in bedrock. Superimposed on the small positive regional anomaly are larger distinct anomalies which are associated with all but one of the nunataks visited, the exception being the sedimentary Pedersen Nunatak. These anomalies strike parallel to the trend of the nunataks, are not detectable much beyond their individual geographical boundaries, have amplitudes in excess of 3,000 gamma, and exhibit a normal magnetization. Only Bull Nunatak has a total-field coverage adequate enough to be considered representative but the author believes that anomalies of a similar order can be expected from all of these nunataks. The 3,000 gamma anomaly measured at Bull Nunatak is probably in itself a conservative value. A detailed vertical-field survey completed over Evensen Nunatak (Fig. 48) shows an anomaly greater than 6,500 gamma, whilst the more regional total-field survey recorded a maximum of only 950 gamma.

It was considered necessary to use a three-dimensional computation to interpret the total-field survey of Bull Nunatak. Using various models, a best fit (Fig. 49) was obtained which indicated that the anomaly is caused by a pipe-like body dipping at 70° in a direction 050°. Although this is the same as the dip measured by Fleet (1968) for the olivine-basalt dyke which crops out along the summit of Bull Nunatak, the interpreted body did not favour a dyke-like origin. Examination of the field station positions, however, shows that no measurements were taken directly over the dyke exposure and, had this been the case, then the magnitude and trend of the anomaly pattern may have supported a composite geological form consisting of a pipe-like body and dyke. The susceptibility used in the calculation was 0·01 c.g.s. units, which is of the same order as that derived by Ashley (1962) from the resultant magnetizations of the olivine-basalts on Tabarin Peninsula.

The anomaly pattern in the vertical component of 6,500 gamma over Evensen Nunatak (Fig. 50) was the result of a detailed systematic survey. The maximum occurs as a sharp peak rising abruptly from the broader main anomaly. It has a wave-length of less than 30 m. and strikes almost parallel to the anomaly trend of 020°. It is suggested that this part of the anomaly is probably due to an olivine-basalt dyke striking in a direction 020° and dipping steeply to the north-west, similar in fact to that observed on several of the other nunataks. Field identification could be hindered by a snow and scree covering. Applying, as a first approximation, the maximum vertical-intensity formula of Nettleton (1942), a vertical dyke of width 9 m. and a susceptibility of 0·01 c.g.s. units would, if within a few metres of the surface, give an anomaly of the required order. No detailed quantitative interpretation has been carried out on the main vertical anomaly but a similar origin to that at Bull Nunatak is suggested.

Adjoining but to the north-west of Evensen Nunatak are two small anomaly belts, an inner one of 900 gamma and an outer one, though still within 600 m. of the nunatak, of 600 gamma. These belts trend parallel to the outline of the nunatak and are probably due to the variable bedrock topography beneath the ice shelf. There is no sign of secondary anomalous belts to the east; this could be expected as the visible eastern boundary is the more precipitous.

It is interesting to speculate on the other nunataks over which there are few or no associated magnetic stations. In the case of Larsen Nunatak, the magnetic data are not conclusive in delineating the strike direction but they do suggest a trend parallel to the observed dyke and hence to the strike of the nunatak itself. In the group including Åkerlundh, Donald, Gray, Arçtowski, Hertha and Castor Nunataks, no dykes

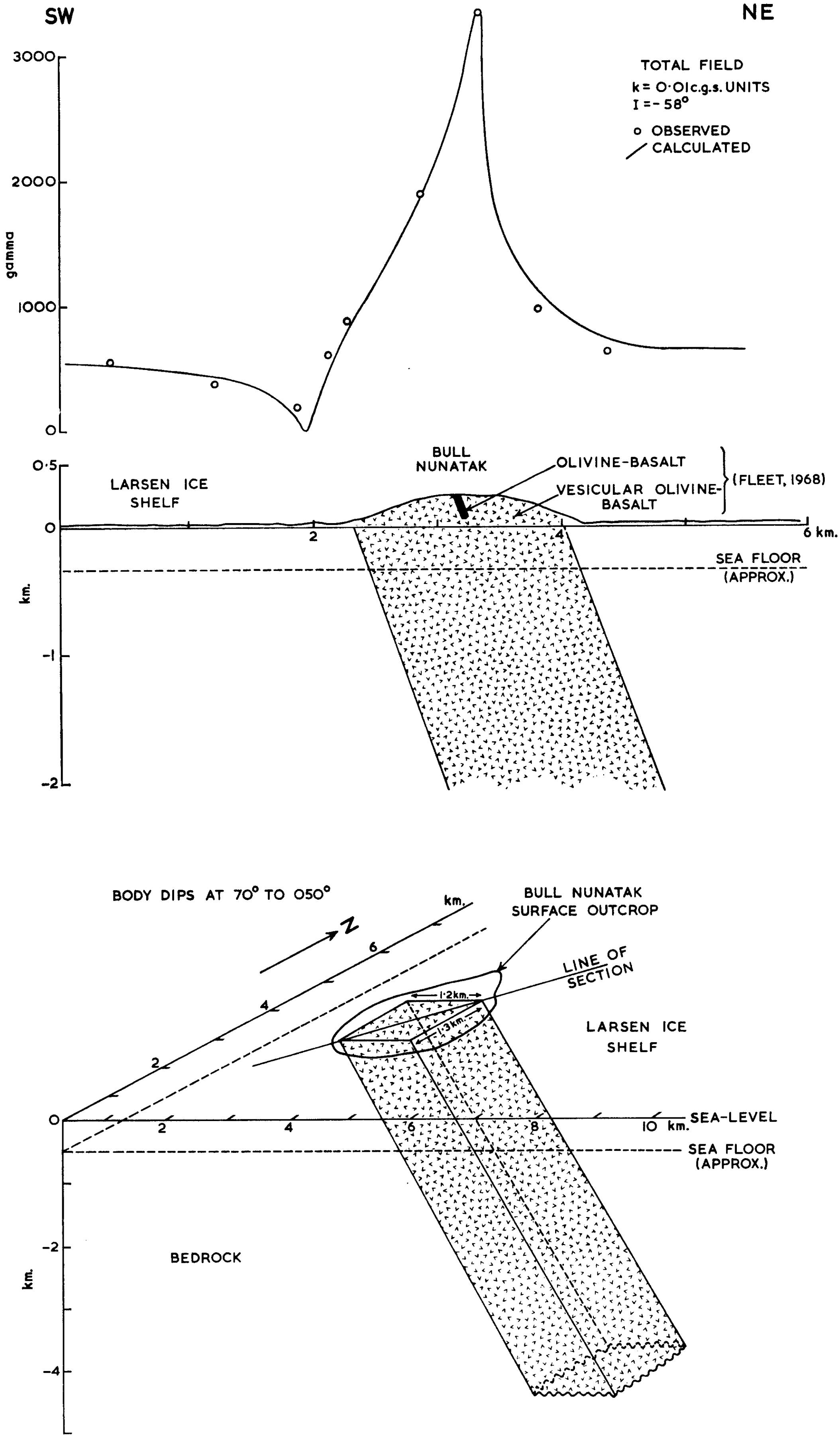

FIGURE 49

A three-dimensional solution for the total-field magnetic anomaly centred over Bull Nunatak.

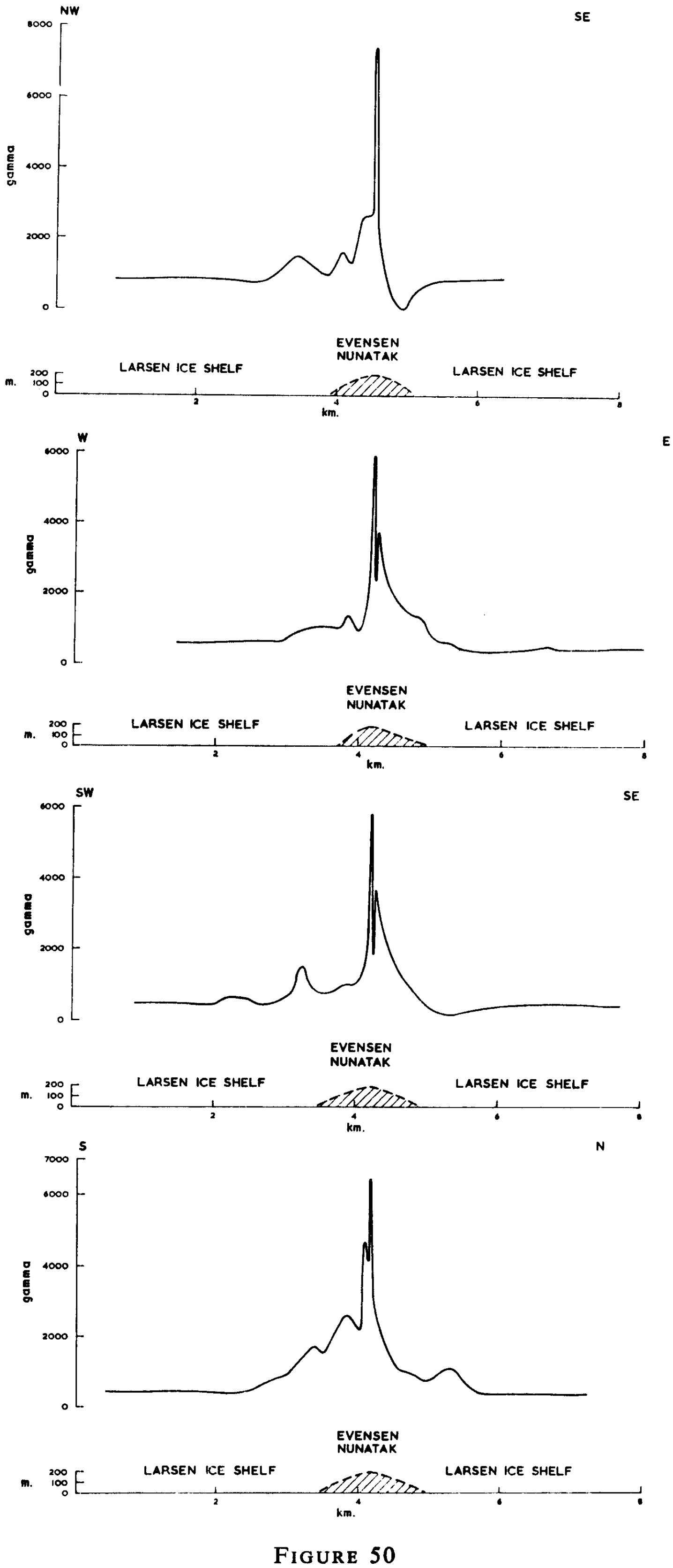

FIGURE 50
Vertical magnetic field profiles over Evensen Nunatak.

have been observed but from the magnetic information available their magnetic strikes again parallel the topographic trend. Christensen Nunatak has an associated basalt dyke with a width of less than 1 m. but the magnetic data do not allow any discussion further than indicating its presence. Of the remaining members of the group, the distribution of magnetic stations does not yet permit further comment. At Murdoch Nunatak a topographical strike of east–west suggests that a dyke, if present, would follow a similar direction. The author suggests that should a detailed magnetic survey be completed over the Seal Nunatak area then all of the nunataks would exhibit similar anomaly patterns to those observed on Bull and Evensen Nunataks.

Disregarding the more easterly Lindenberg Island and Christensen Nunatak, the author extrapolated westwards the magnetic, topographical and geological strike directions of the remaining nunataks. It is found that they all converge in a small area 5–10 km. west of Dallmann Nunatak.

On the mainland to the west there is only one recorded dyke (Elliot, 1966), which shows affinities with the olivine-basalt dykes of the James Ross Island Volcanic Group. This crops out at the head of Drygalski Glacier, is 0·9 m. wide, of olivine-basalt and strikes south-west to north-east. Although its strike is not parallel to that encountered in the Seal Nunataks, it would lie along the westerly projection of their mean strike.

B. Churchill Peninsula–Jason Peninsula Area

The gravity and magnetic maps compiled from traverses along the Oscar II and Foyn Coasts are shown in Figs. 51 and 52. This area has a relatively detailed field coverage due to the siting of the gravity base station and supply depot at the northern end of Churchill Peninsula. The topography of the area contrasts with that farther south in that a static ice piedmont (Marsh and Stubbs, 1969) rises gradually from the Larsen Ice Shelf to a "sub-plateau" at 914 m. before a sharp gradient increase to the Bruce Plateau at 1,524 m. Churchill Peninsula itself is about 50 km. in length and trends north-west to south-east into the Larsen Ice Shelf. Maximum heights of over 300 m. are attained and they are coincident with rock exposures both towards the northern and southern ends. Elsewhere, the surface is snow-covered and gently undulating with a distinct saddle joining Cape Alexander to the remainder of the peninsula. North-east of Churchill Peninsula is the more extensive and irregularly shaped Jason Peninsula. It reaches a maximum elevation of 470 m. and projects eastward for 65 km. to the edge of the Larsen Ice Shelf. The very existence of the ice shelf can largely be attributed to the eastern extent of Jason Peninsula. The surface topography of Jason Peninsula is essentially one of gentle snow-covered undulations and, because of the gradual geomorphological gradients, it is often very difficult to determine the junctions between the land ice and the ice shelf.

Figs. 51 and 52 show that the most significant gravity and magnetic anomalies are situated over the northern part of Churchill Peninsula. The geology of this area has been described by Marsh (1968) as consisting of a gabbro–microgranite laccolith intruded into volcanic rocks. He estimated that the laccolith has a lateral extent of at least 19 km. and a thickness of over 610 m., of which 450 m. are gabbroic. A modal analysis of a gabbro specimen yielded 11·6 per cent of iron ore, though over 50 per cent was recorded in the contact rocks. Density measurements by the author on one gabbro specimen gave a value of 2·86 g. cm.$^{-3}$.

The magnetic anomaly (Fig. 52) is elongate, strikes approximately north–south and has two main peaks, each in excess of 1,200 gamma, separated by a col of lower magnetic intensity. On the evidence of the magnetic data, Kennett (1966*a*) concluded that a steep-sided body approximately 5 km. in width and approaching the surface could be a possible source for the anomaly. Using a susceptibility of 0·01 c.g.s. units, the author applied a two-dimensional computer interpretation, the result of which favoured a laccolithic origin (Fig. 53). The computed shape of the body was asymmetrical with the side towards Cabinet Inlet being almost vertical compared with that towards Adie Inlet, where the upper surface slopes gradually beneath the ice cover. At the highest point of the section, where the body is not exposed, it probably lies within a few metres of the surface.

Accuracy of the Bouguer anomalies is again limited by the unknown ice thicknesses but, after the removal of a regional gradient and allowance for ice thickness, a maximum anomaly of + 20 mgal is indicated. Depth estimates (Nettleton, 1942; Skeels, 1963) from the Bouguer anomaly suggest the body lies within 500 m. of the surface and has its lower surface at about a depth of 7 km.

Over the remainder of Churchill Peninsula the residual Bouguer anomalies have been interpreted as due

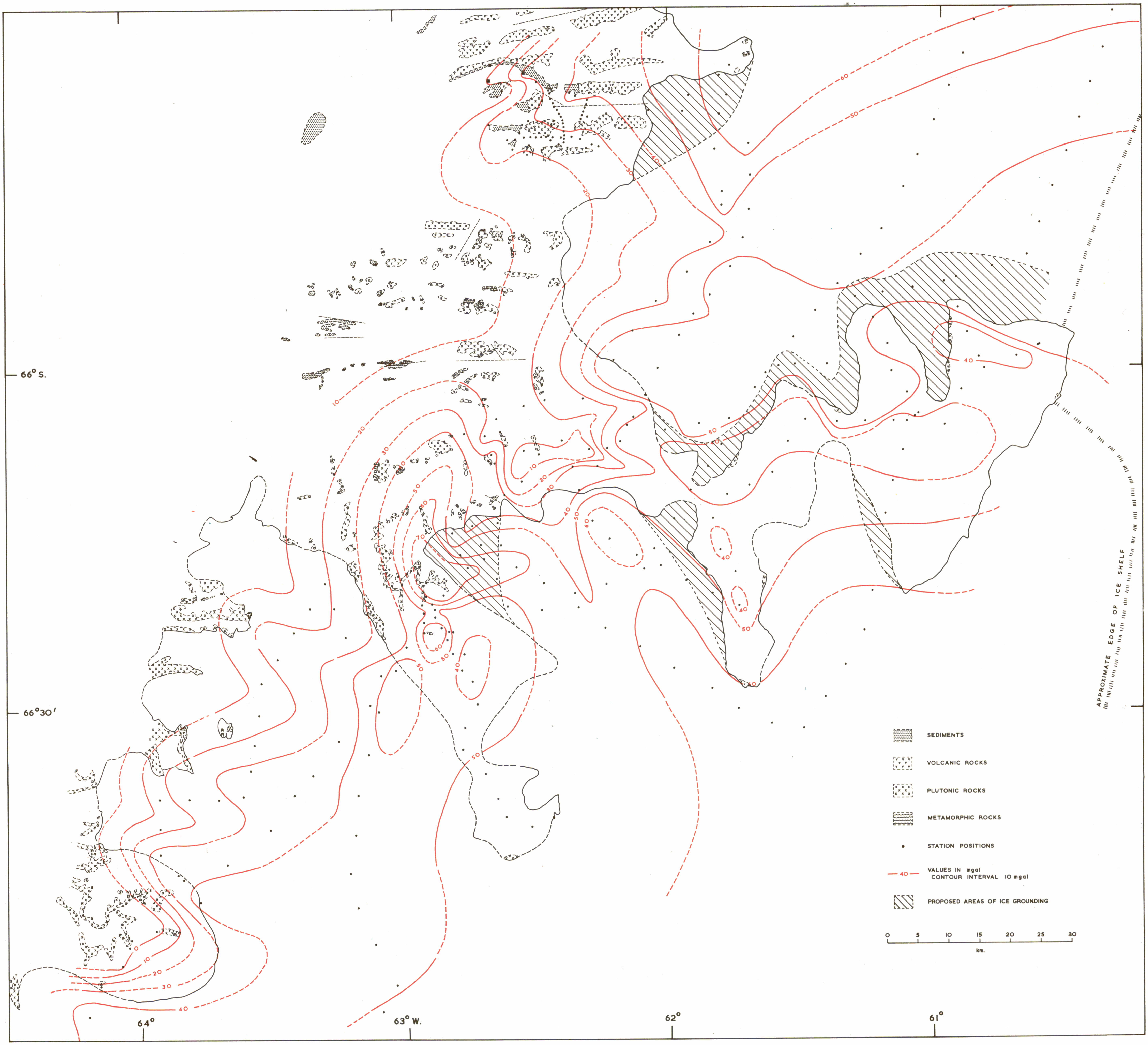

FIGURE 51
Map of the gravity anomalies and geology (Marsh, 1968) of the Churchill Peninsula area.

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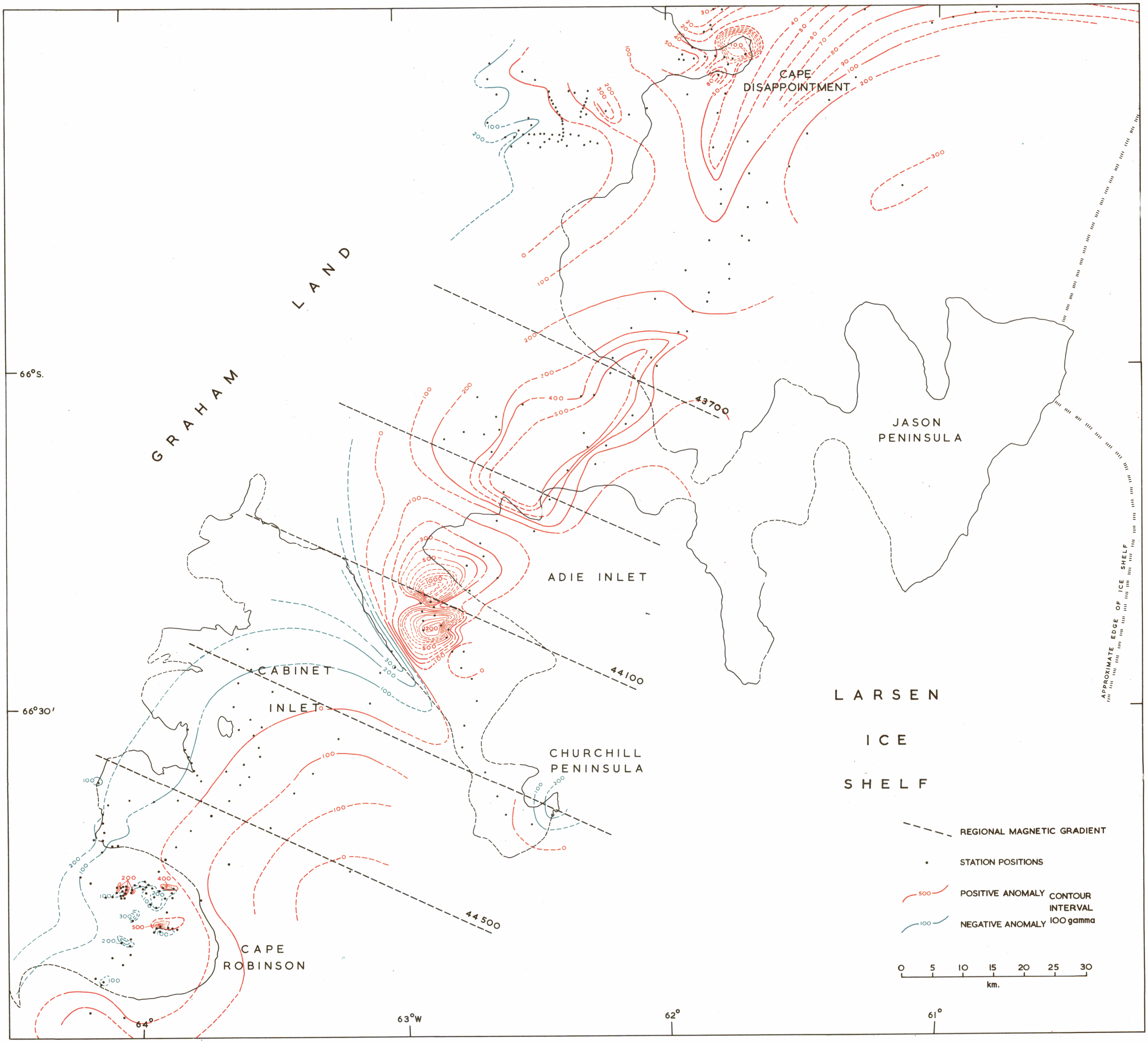

FIGURE 52
Total-field magnetic anomaly map of the Churchill Peninsula area.

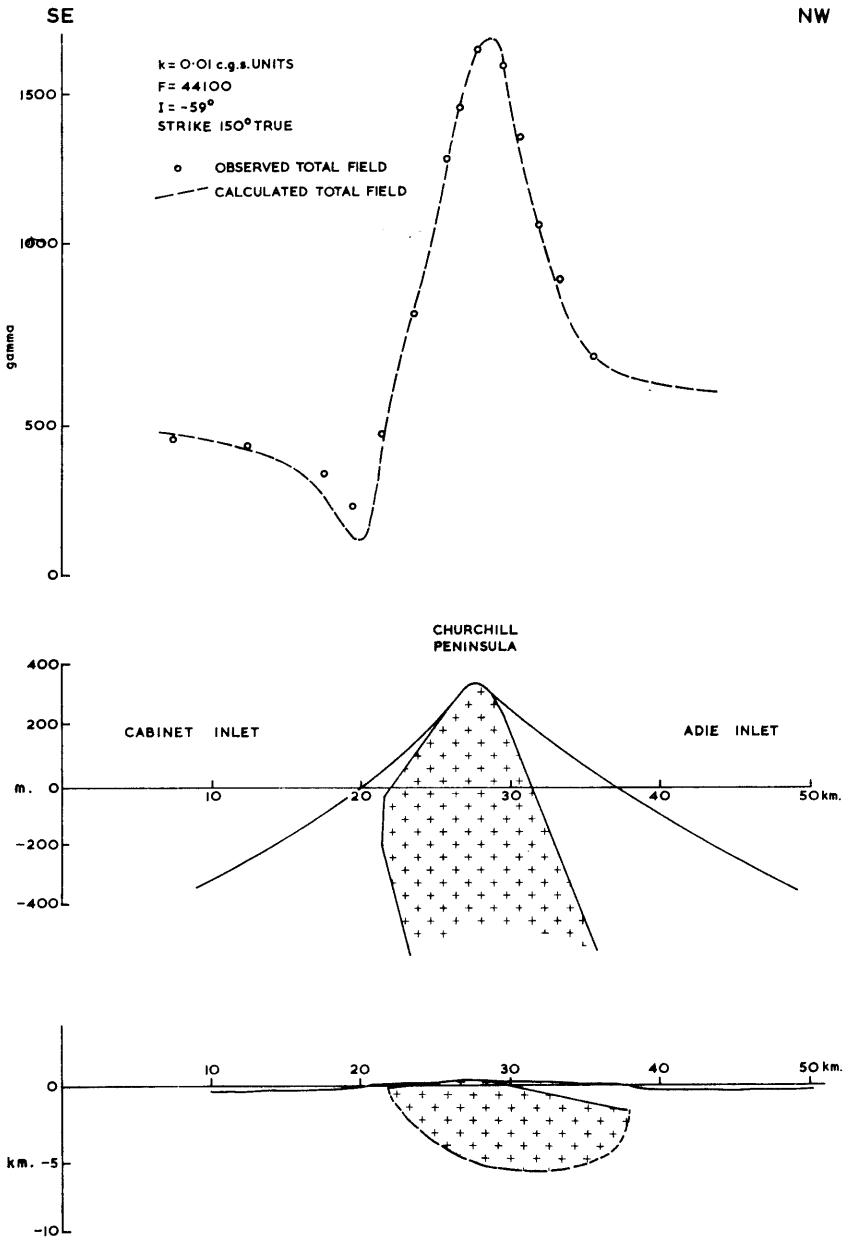

FIGURE 53
Calculated sub-surface distribution of the gabbro–microgranite laccolith beneath the northern end of Churchill Peninsula.

to variation in ice thicknesses. Although it is not possible to give accurate quantitative estimates of thicknesses, some indication as to the relative distribution has been attempted. Fig. 54 illustrates the ice thicknesses estimated using a simple two-dimensional relationship with a density contrast of 1·77 g. cm.$^{-3}$ between ice and bedrock. Over the central areas of Churchill Peninsula, the ice has been calculated on a grounded-ice model as lying near or below sea-level, which would involve ice thicknesses of the order of 250 m. At the southern end of Churchill Peninsula, and provided that the anomalies are again caused solely by variable bedrock topography, the ice-thickness values must approach 150 m. The low-density rhyolitic rocks, which are known to be exposed at Astro Cliffs on Cape Alexander, may partially contribute to the anomaly. Recent radio echo-sounding ice-thickness profiles (Smith, 1972) provide evidence which substantiates the above interpretation. Over the area at the southern end of Churchill Peninsula, ice-thickness values of 89–240 m. are given, while over the central area thicknesses around 400 m. are quoted. The latter value is of considerable importance for it does support the geophysical interpretation in that much of Churchill Peninsula would lie below sea-level should the ice be removed.

The gravity anomaly at the northern end of Churchill Peninsula has been attributed to a geological source but, as with many of the local anomalies in this area, it is difficult to differentiate between the contribution due to the sub-ice topography and that due to density variations within the bedrock. Because of this, the values given in Fig. 54 must be considered as first approximations.

On the mainland west of Jason Peninsula there is a broad north-east to south-west magnetic feature exhibiting a positive anomaly of 500 gamma amplitude. It is roughly coincident with a gravity low, which, after allowance for the ice–rock distribution, is estimated to have a maximum value of −25 mgal. The nunataks in the vicinity have been described by Marsh (1968) as consisting of Upper Jurassic dacites with subsidiary rhyolites and rhyolitic tuffs. Density measurements on such samples from the Oscar II and Foyn Coasts have given values of $2 \cdot 63 \pm 0 \cdot 06$ g. cm.$^{-3}$ while the mean of three rhyolitic-dacitic specimens from localities close to the anomaly gave $2 \cdot 60$ g. cm.$^{-3}$. Density contrasts to the bedrock could account for all of the anomaly but considerable thicknesses of volcanic rocks would be required. As the gravity minimum does not extend over all the area known to be underlain by the volcanic rocks, it is considered that variations in the sub-surface depths to bedrock are, again, in part responsible. On the assumption that the anomaly is caused solely by topographical variations, then depths to the bedrock of 300 m. below sea-level can be expected (Fig. 55). This would give on removal of the ice a sub-sea-level channel between Adie Inlet and the ice shelf to the north of Philippi Rise. Jason Peninsula, which is now joined to the mainland by a 50 m. high snow col, would become isolated. Examination of the 1:200,000 map (sheet W6662) shows that the area of the proposed channel follows a topographical feature which may be traced inland and which may form a possible glacial outlet. The estimated western limit of this feature is determined by the 300 m. contour between Borchgrevink Nunatak and the group of nunataks north of Gulliver Nunatak.

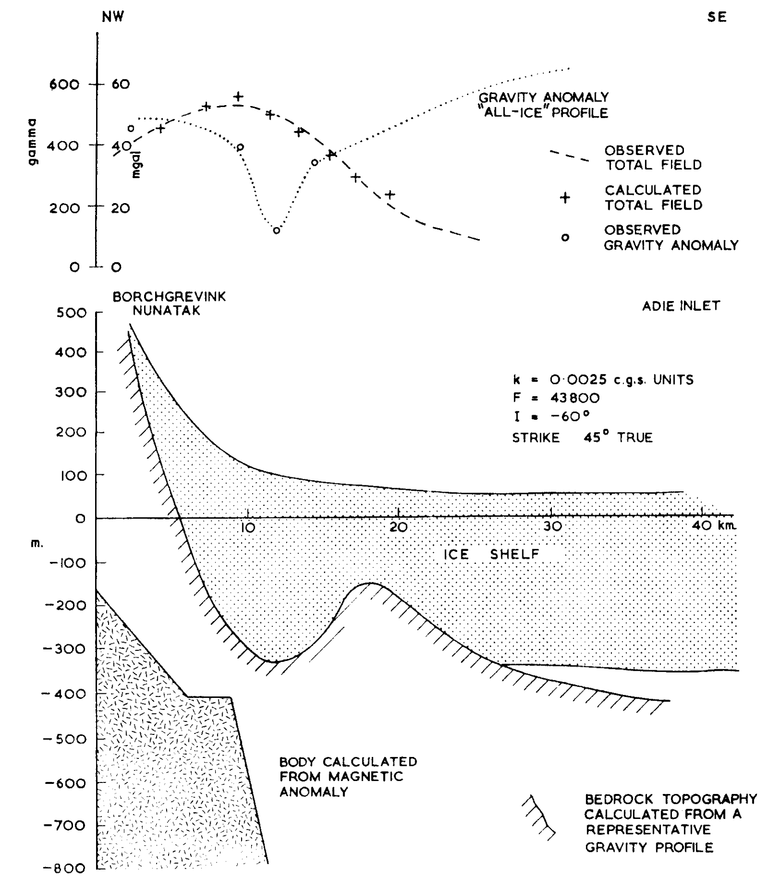

FIGURE 55

Gravity and total-field magnetic profiles west of Jason Peninsula.

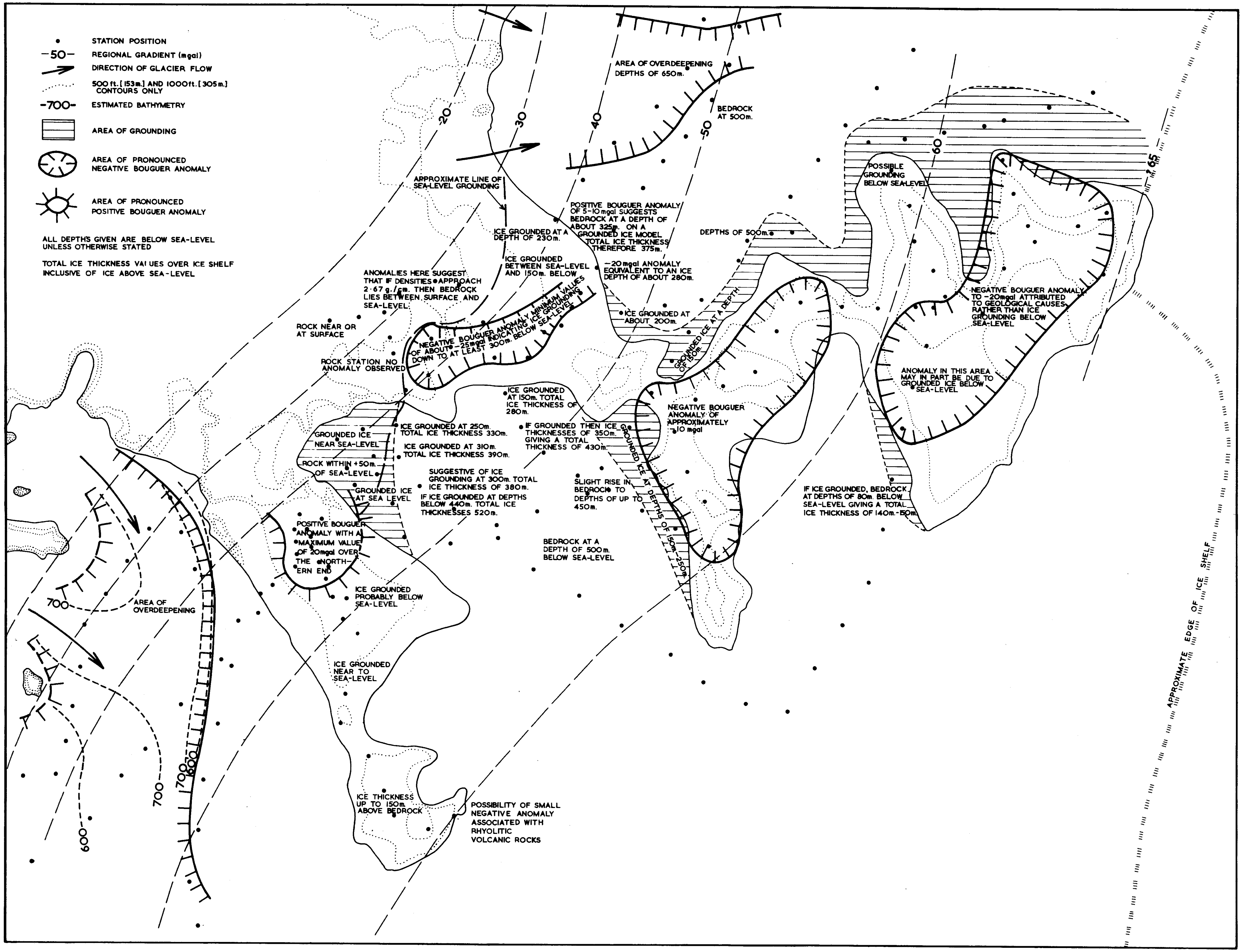

FIGURE 54

Diagram of the Churchill Peninsula–Jason Peninsula area, showing approximate depths to bedrock as calculated from the residual gravity anomalies.

The distribution of the total-field magnetic data does not permit detailed anomaly contouring but by using the stations available certain limiting dimensions can be calculated. Using an assumed susceptibility of 0·0025 c.g.s. units, it was calculated that a body with its upper surface at a depth of 400 m. below sea-level could produce an anomaly of the observed order. This would place the upper surface less than 100 m. below the bedrock as calculated from the gravity data. Fig. 55 shows a possible solution for an intrusive body. The calculated body terminates sharply to the east but it rises beneath the ice cover and bedrock to the west. In addition to the volcanic rocks, there are also exposures of hornblende-biotite-granodiorites. These rocks crop out parallel to the Upper Jurassic dacites yet they occur 8 km. to the west where they are in contact with the volcanic rocks and members of the metamorphic complex. The only susceptibility measurement from this area was 0·0004 c.g.s. units for a sample of granodiorite from Gulliver Nunatak (Kennett, 1966*a*).

Apart from a general increase away from the axis of Graham Land, the Larsen Ice Shelf has been found (p. 57) to be magnetically quiet. Off Cape Disappointment the contours do, however, take on a more east–west trend as do the gravity results, indicating the probability of a considerable variation in the bottom topography beneath the ice shelf, or of a major structural disconformity. Also, it is apparent that between Philippi Rise and Cape Disappointment, and in Cabinet Inlet, there is overdeepening which can be attributed to glacial action. Both of these areas are on easterly projections of existing major glacier systems. In Cabinet Inlet, depths down to at least 750 m. are calculated which approach the 820 m. estimated by Kennett (1966*a*) on a grounded-ice model; east of Flask and Leppard Glaciers depths of 650 m. are indicated. In both areas the depths shallow to the east but the gravity data do suggest that the overdeepening associated with Flask and Leppard Glaciers may continue for at least 60 km. from the mainland. Immediately to the north of Philippi Rise and before the overdeepening associated with Flask Glacier, the depth to bedrock is estimated at around 250–350 m., which is less than the average depth of 500 m. to the continental shelf. The ice is considered to be floating but probably on a very thin water layer. The transition from the ice shelf to Philippi Rise is extremely gradual. Grounded-ice thicknesses on the land in this vicinity are about 200 m. below sea-level which conforms to the sub-sea-level link with Adie Inlet. The author considers that in Adie Inlet there must be an area of grounding and ice thinning perhaps extending as far as 20 km. from the coast. Within this limit the ice thicknesses decrease, thinning to an estimated 200 m. at the head of the inlet where grounding at sea-level is proposed. A marked rise in surface elevation at the head of Adie Inlet is further evidence of a rise in bedrock. Adie Inlet thus contrasts with the majority of inlets so far investigated in that there appears to be no associated glacial overdeepening. This would appear to correlate with the present geomorphology of the area, since the head of the inlet does not act as a significant glacial outlet. The ice appears to be deflected from the vicinity of Gulliver Nunatak towards the area where the sub-sea-level channel has been proposed. Measured ice-thickness values (Smith, 1972) in central Adie Inlet are within the maximum calculated by the author (Fig. 54).

Covering the greater part of Jason Peninsula are two negative north-east to south-west trending Bouguer gravity anomalies. These may be the result of Upper Jurassic acid volcanic rocks known to be exposed locally. An area on the north coast where dark vesicular basalt crops out is devoid of any such anomaly. The gravity results suggest that much of the ice adjoining Jason Peninsula is grounded (Fig. 54) and that areas which have been recognized as "land" are of ice grounded at and below sea-level. This would also account for the gradual merging of the land ice of Jason Peninsula and the ice shelf. Removal of the ice cover would probably show that Jason Peninsula consists of a few isolated rock outcrops severed from the mainland with any geological continuity occurring below sea-level.

Local geophysical traverses carried out over Cape Robinson show several short wave-length anomalies with amplitudes of about 500 gamma. These are directly associated with the doleritic and microdioritic dyke rocks identified by Stubbs (1968).

C. West–East Traverses across the Antarctic Peninsula

From the available data, three transverse sections (Fig. 56) have been constructed across the axis of the Antarctic Peninsula. Despite the errors inherent in the data, it is clear that there is a gradual southerly decrease in the Bouguer anomalies, the latter becoming increasingly negative. This must be allied to crustal structures at depth though nothing further can be added to that previously discussed. Of considerable

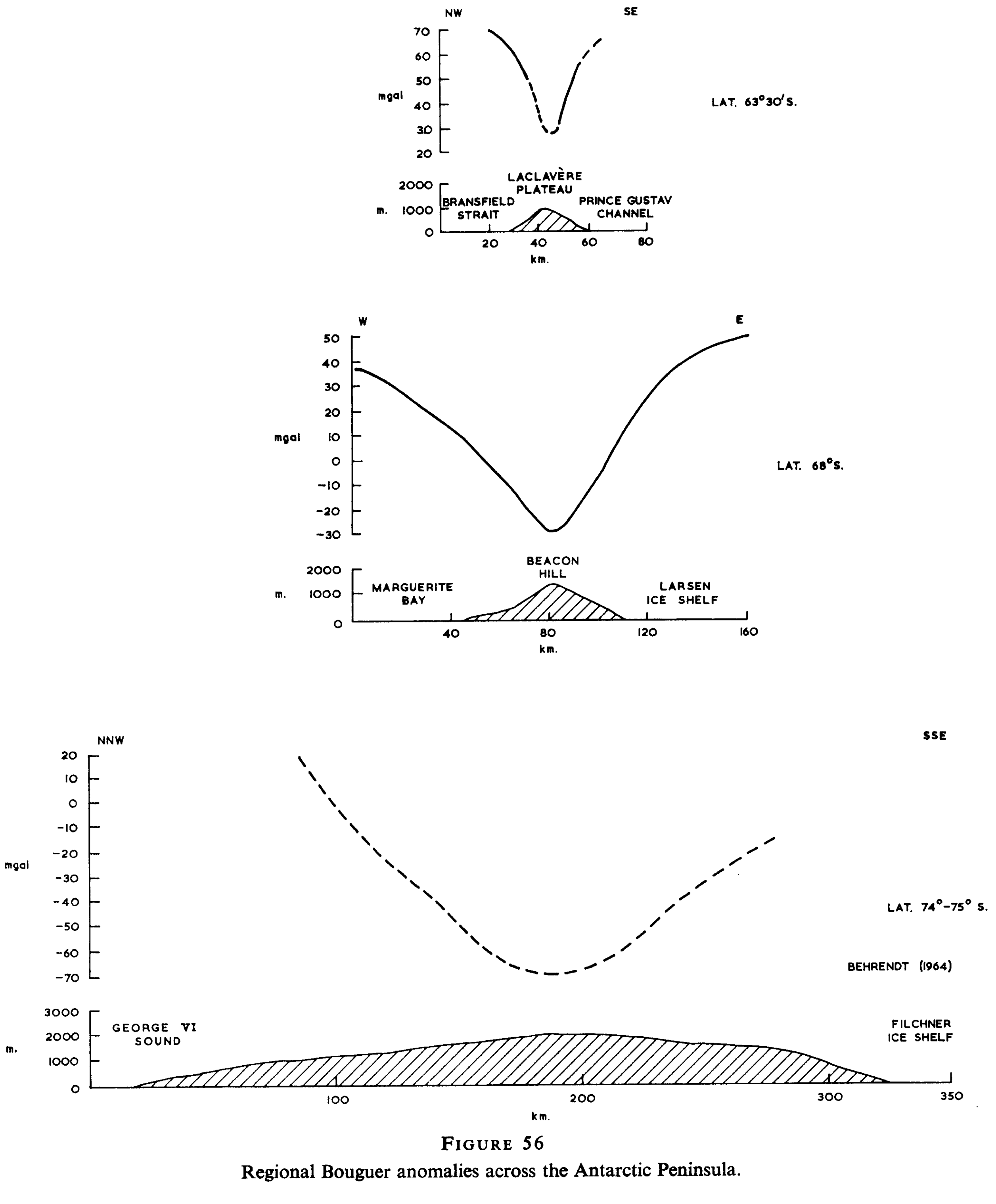

FIGURE 56

Regional Bouguer anomalies across the Antarctic Peninsula.

interest, however, is the area around lat. 68° S., where there are two west–east traverses (p. 45) within 50 km. of one another and which exhibit marked differences in their anomaly profiles. Neny Trough, along which the more southerly traverse follows, is a north-west to south-east structure which averages about 5 km. in width, is over 60 km. in length and, reaching a maximum elevation of 1,000 m., lies about 500 m. beneath the average height of the surrounding plateau. Unfortunately, heavy crevassing over the western section of Neny Glacier did not permit geophysical traverses throughout the full length of Neny Trough, hence the inclusion of the alternative Snowshoe Glacier route. Neny Glacier and Snowshoe Glacier are separated by 10 km. at the point of outflow into Neny Fjord but they have a common origin at the head of Gibbs Glacier.

Two basic assumptions have been made in carrying out the interpretation. The first concerns the regional gradient selected over the Beacon Hill traverse. The initial Bouguer anomaly profile (Fig. 57) shows two distinct anomalous areas: one at either side of the peninsula and each coincident with complex topographical areas. The one on the west side is associated with steep gradients at the head of Northeast Glacier and the one on the east coast with the junction between the land ice and the ice shelf. The associated anomalies were attributed to the use of incorrect ice–rock distributions in the data reduction. The regional Bouguer

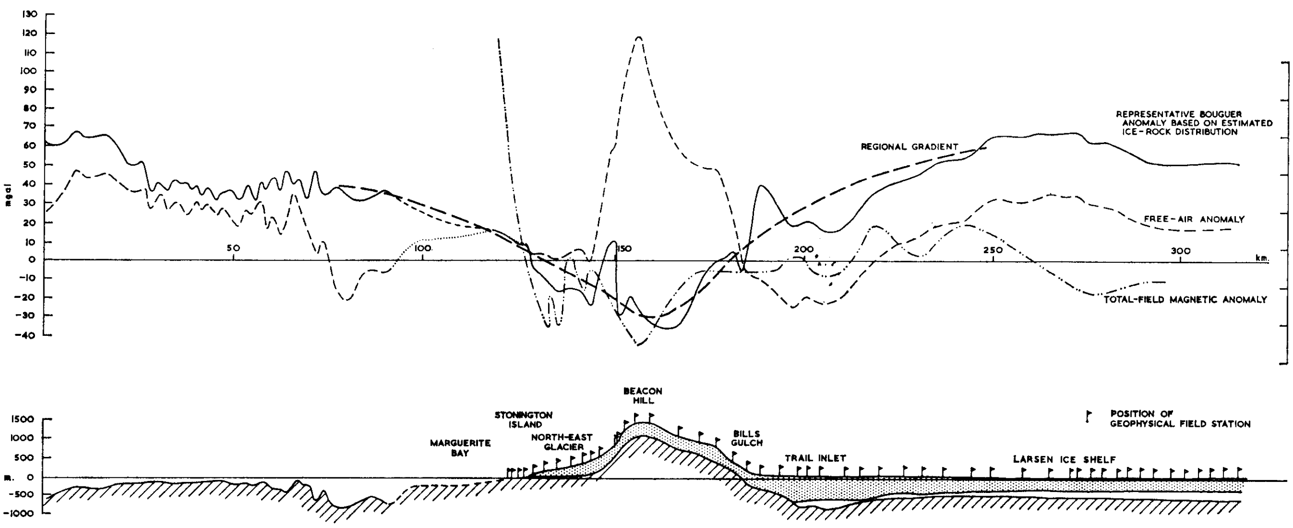

FIGURE 57

Gravity, magnetic and topographical profiles along the Stonington Island–Beacon Hill–Trail Inlet traverse. The grounded ice–rock distribution was calculated from the residual Bouguer anomalies. The bedrock elevation beneath Trail Inlet was compiled from gravity data and total ice-thickness values.

anomaly which was obtained by smoothing through those disturbed areas was used to give a residual anomaly on which was based the recalculated ice–rock distribution shown in Fig. 57. The regional gradient considered to be representative of the Beacon Hill traverse was also assumed to be representative for the area of Gibbs Glacier.

The second assumption concerns the residual anomaly obtained through Neny Trough by the removal of the regional gravity gradient. The interpretation of the residual anomaly, which in places exceeds −50 mgal, has again been attributed to variation in the ice–rock distributions and not due to local density changes within the underlying geological formations. Kennett (1966*a*), who has already drawn attention to the difficulty of obtaining a true Bouguer anomaly across the Neny Trough section, produced variable profiles dependent on different ice–rock distributions beneath the gravity stations. Fig. 58 shows topographical and geophysical sections along the traverse with a Bouguer anomaly calculated from an ice–rock ratio above sea-level of approximately 2:1.

Interpretation of the residual anomaly was completed using the two-dimensional iterative gravity programme. The eventual profile showed an agreement to within 0·01 mgal of the observed profile but the calculated depths are likely to be an underestimate of the true depths. This is because the profile does not represent a two-dimensional feature, since the glacier walls restrict its extent at right-angles to the traverse.

Along the length of the traverse the ice thicknesses of Snowshoe Glacier appear to be less than 150 m. This apparently low figure is understandable when one considers that the line of section is not central but follows along the side of the glacier. The most important results of the interpretation are the values for ice thicknesses in the Gibbs Glacier–Mercator Ice Piedmont area. In Mobiloil Inlet, the greatest calculated depth for the bedrock (a little over 700 m. below sea-level) is only several kilometres to the east of the junction between the land ice and ice shelf (as mapped on the D.O.S. 1:500,000 sheet 13 of the area).

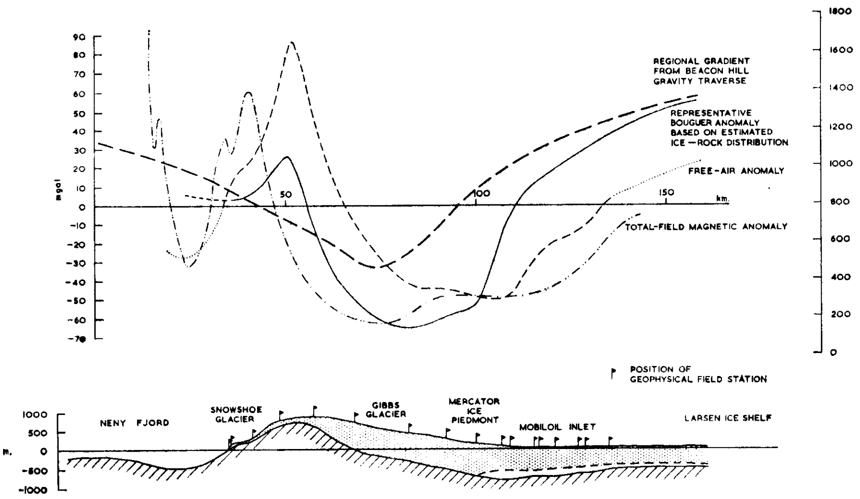

FIGURE 58

Gravity, magnetic and topographical profiles along the Neny Fjord–Gibbs Glacier–Mobiloil Inlet traverse. The bedrock elevation was calculated from residual gravity anomalies.

Depths greater than the 500 m. average are found as far as 30 km. east from the junction and this is in accordance with similar inlets to the north where glacial overdeepening has been suggested as the primary cause. To the west, however, Gibbs Glacier is seen to harbour considerable thicknesses of ice (Fig. 58), values up to 800 m. being calculated and with bedrock below sea-level for as much as 30 km. from the junction with the Larsen Ice Shelf. Despite the attitude of the bedrock, the author estimates that the junction between the grounded and floating ice still occurs in the vicinity of the surveyed coastline. The traverse by Kennett (1966*a*) along the northern side of Mobiloil Inlet suggests that bedrock rises by about 180 m. above that found to the south but, since this is likely to be associated with ice thinning, it does not automatically follow that there is grounding. On the other hand, a wide belt of heavy crevassing is clearly visible along the southern part of Mobiloil Inlet. Here the ice may be grounded, though an alternative explanation is that the rock bluffs beneath Werner Peak restrict the easterly outflow of the ice from the Mercator Ice Piedmont. The calculated ice thicknesses bear witness to the fact that the Mercator Ice Piedmont must act as a vast amphitheatre for the collection of ice from several major glaciers.

Kennett (1966*a*) commented that the free-air anomaly peak lies to the west of the highest point of the traverse. This imparts a certain asymmetry to the profile but examination of the calculated bedrock section (Fig. 57) clarifies the apparent displacement. From the section, it can be seen that the greatest mass of bedrock does lie to the west of the highest point and, as the free-air anomaly reflects the physical map in areas where local isostatic compensation is incomplete, its position would naturally be centred over the main rock mass. On the Beacon Hill traverse any such displacement is not so obvious though a similar explanation could apply to the slight apparent westerly shift.

Discussion of the total-field magnetic traverses over Trinity Peninsula and south of lat. 69° has been given on p. 42–45, where the anomalies have largely been interpreted as due to an irregular subglacial relief or to the distribution of the Andean Intrusive Suite. On a regional basis, the surveys over Graham Land have shown that the magnetic intensity decreases towards the central axis of the peninsula, values of −250 gamma having been recorded over Trinity Peninsula with more negative values farther to the south. In the vicinity of Neny Trough (Figs. 58 and 59), the magnetic pattern shows a marked variation and the traverse from Gibbs Glacier and the Mercator Ice Piedmont to the south illustrates this point (Fig. 21). On the plateau area, anomalies of 600 gamma amplitude are found but over the ice piedmont an abrupt change in magnetic character is evident for anomalies of only 60 gamma amplitude are superimposed upon an undisturbed magnetic background. Similarly, very little magnetic activity is observed up to the head of Gibbs Glacier. However, along Wyatt and Snowshoe Glaciers, anomalies of greater amplitude are again observed. The magnetic strike thus parallels the topographical and gravity trends.

An apparent mean susceptibility of the order of 0·0036 c.g.s. units has been calculated for the rocks at the southern end of the Antarctic Peninsula (Behrendt, 1964). Here the high-amplitude anomalies are

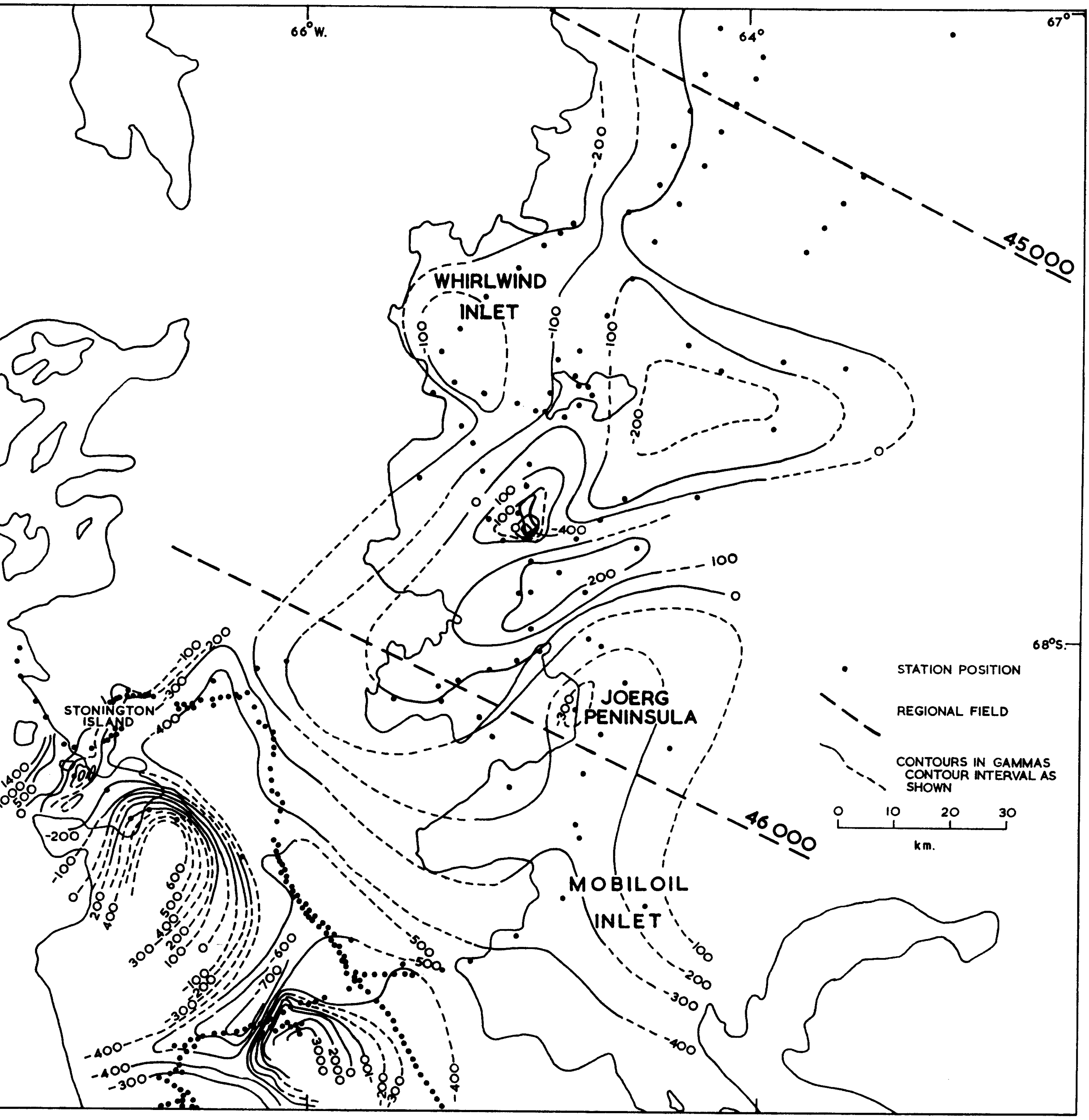

FIGURE 59
Total-field magnetic anomalies over Graham Land between lat. 67° 00′ and 68° 45′ S.

thought to be caused by the shallow extrusive rocks with 52 per cent of the depths to the surface calculated at within 0·5 km. of the ice–rock interface. A similar susceptibility was used for the interpretation of the southern plateau anomalies and by assuming that the source depths lay at the ice–rock interface then some indication of ice thickness could be obtained. The resulting depths (Fig. 21) show, in general, close agreement with those derived from the gravity data, particularly over Gibbs Glacier and the western end of the Mercator Ice Piedmont. In addition, the bedrock profile beneath Wyatt Glacier is, as expected, much steeper, whilst the ice gradually thins at the southern edge of the Mercator Ice Piedmont as it transgresses into Weyerhaeuser Glacier. The central area of the Mercator Ice Piedmont is the thickest as far as ice depths are concerned but values of 1,500 m. obtained at its central eastern end are far in excess of those derived by other means.

The Neny Trough feature is geologically of great interest and also perhaps of great significance in the structural history of the Antarctic Peninsula. Regionally, the gravity and magnetic data appear to follow the solid topography over the sections visited. Where the landform is obviously depressed beneath the surrounding ice plateau, then negative anomalies are found. Neither the negative gravity nor magnetic anomalies can, however, be traced on Wyatt Glacier nor has any evidence been found of a northerly extension to the Beacon Hill traverse. Likewise, there is no geophysical evidence of the structure swinging through Snowshoe Glacier. Therefore, should it continue to the west coast, its course must lie along the obvious yet highly crevassed Neny Glacier route. The author suggests that any continuation of the Neny Trough feature on to the east coast will be found along the southern side of Mobiloil Inlet towards Poseidon Pass.

V. ISOSTASY OVER THE ANTARCTIC PENINSULA

THE Antarctic continent is an area of considerable interest for isostatic studies due to the ice overload. Gravity measurements on the continent, however, are susceptible to large errors and because of this Bentley (1964) considered that a mean free-air anomaly of at least $\pm$ 20 mgal is necessary before it can be considered significantly different from zero. Woollard (1962*a*) used free-air anomalies to show that isostatic equilibrium does exist on a regional basis throughout Antarctica for areal segments greater than 350 km. in width. Grushinsky and Frolov (1967) averaged isostatic anomalies over a 167 km. radius pattern and showed that substantial positive isostatic anomalies are to be found in the peripheral zones of Antarctica. They also found that, whilst large areas of central Antarctica have isostatic anomalies within the range of $\pm$ 10 mgal, and hence are isostatically compensated, there are in this region few zones of negative anomalies.

Although average free-air anomalies in western Antarctica have shown it as a region of isostatic compensation, the Antarctic Peninsula does exhibit relatively abnormal isostatic conditions. At the southern end of the peninsula, east of Eights station the free-air anomaly averages $+ 60 \pm 30$ mgal (Behrendt, 1964), which is interpreted as an area with a rough subglacial topography with incomplete local compensation, although regional compensation is a possible alternative. To the west of Eights station in Ellsworth Land the average free-air anomaly of $+ 11$ mgal suggests isostatic equilibrium as also is the case over most of the Filchner Ice Shelf, where Behrendt (1962) determined an average free-air anomaly of $- 11$ mgal. However, the trough beneath the eastern edge of the Filchner Ice Shelf was considered only partially compensated. Frolov (1965*b*) has published a diagram which shows over Graham Land two calculated isostatic anomaly values of $+ 42$ and $+ 54$ mgal. Davey (1971) completed preliminary isostatic investigations across the South Shetland Islands, Bransfield Strait and Trinity Peninsula, and likewise showed the area to be one of undercompensation, with a mean isostatic anomaly approaching $+ 50$ mgal.

In considering the isostasy conditions, it may be that two different aspects are involved: those due to the geological effects of crustal down-warping and those caused by the overloading and deglaciation of the ice cover during Pleistocene and Recent times. Due to the once greater ice overload, the area of the Antarctic Peninsula should still provide evidence of the submergence which was necessary to maintain isostatic equilibrium. Geomorphological evidence supporting this has been provided by Adie (1964*b*). A minimum submergence totalling 305 m. was estimated, this being the difference between the 429–490 m. depth of the continental shelf off the Antarctic Peninsula and that of the world average depth of 132 m. for the shelf margin. Cross-sections over the continental shelves from South Georgia to Graham Land show a southerly increase in depths which may be equated with a corresponding thickening of the ice sheet. During and since deglaciation, a gradual recovery to former levels could be expected. The discovery of raised beaches, sea caves and wave-cut platforms along the Graham Land coast (Adie, 1964*b*) provides evidence of isostatic emergence. John and Sugden (1971) have reported raised beaches in the South Shetland Islands which they also attributed to isostatic recovery since deglaciation.

The Antarctic Peninsula, especially the northern area, has provided much speculation on its structural history; thus, apparent isostatic anomalies may be caused by tectonic displacements, local changes in crustal density or by variations in the upper mantle. The dimensions alone of the Antarctic Peninsula may not even qualify for local compensatory phenomena. On the present knowledge, however, it would appear that the Antarctic Peninsula is probably associated with a positive isostatic anomaly indicating an area which is not fully compensated. Further information, in particular from crustal seismic studies, is required before real departures from isostatic equilibrium can be isolated from apparent ones, so that the respective roles played by local and regional compensation can be determined.

VI. SUMMARY

THE results have been presented for gravity and magnetic traverses undertaken over the Antarctic Peninsula prior to 1967. Data from marine gravity, magnetic and seismic profiles have been incorporated and used to investigate possible regional anomalies adjacent to the mainland. Aeromagnetic profiles along the length of the peninsula have also been considered, whilst airborne radio echo-sounding techniques over some areas have made possible greater accuracy in data reduction and provided limiting depths to bedrock.

Despite the errors inherent in the land geophysical data, particularly those arising from the gravity data reduction, much useful information has been gained from the interpretation.

Re-interpretation of previous magnetic results in north-east Trinity Peninsula (Ashley, 1962; Allen, 1966*a*), in the light of additional total-field magnetic data, has supported the conclusions previously reached. The northern 30 km. of Trinity Peninsula appear to be underlain by an extensive batholith system composed of members of the Andean Intrusive Suite. Variations in the topography of its upper surface have been held responsible for the observed magnetic anomalies. Due either to increasing thicknesses of the Trinity Peninsula Series or to the presence of non-magnetic plutonic representatives, there exists south of Fidase Peaks–Camel Nunataks an intervening area almost devoid of magnetic activity. This continues for about 50 km. down the peninsula until magnetic anomalies are again observed on the west coast at Thanaron Point. It is still farther to the south along the western shores of Prince Gustav Channel that the anomalous areas make their re-appearance. In contrast to those due to the northern batholith, these more southerly anomalies appear to be less extensive, and hence their isolation suggests individual plutonic bodies not necessarily connected at depth. On the basis of a vertical-component magnetic survey, Ashley (1962) postulated that at Summit Pass between Trinity and Tabarin Peninsulas there is a laccolithic diorite body. The total-field magnetic results corroborate this conclusion.

Apart from an isolated magnetic anomaly area attributed to a gabbro–microgranite laccolith, it is not until south of lat. 68° S. that significant magnetic anomalies again become apparent over the mainland. However, much of central Graham Land remains unexplored by geophysical field parties and therefore insufficient data coverage must contribute to the apparent absence of anomalies. The anomalies which have been observed on the southern plateau traverses are not dissimilar from those attributed to the Andean Intrusive Suite to the north.

Consideration has been given to a possible structural trend existing between two traverses across the plateau in the vicinity of lat. 69° 30′ to 70° S. There is some indication that a north-west to south-east trend is present which may be a reflection of the underlying bedrock relief. A two-dimensional interpretation of the magnetic anomalies was attempted along one of the profiles on the assumption that variations in ice thicknesses were responsible. Close analogy was shown between the bedrock relief computed by the author and that obtained at the base of the Antarctic Peninsula by Behrendt (1964). A similar geology has already been suggested for both areas.

The only other magnetic anomalies of note over Graham Land are associated with the James Ross Island Volcanic Group at the Seal Nunataks, over Tabarin Peninsula and to the east of Prince Gustav Channel. Ashley (1962) and Allen (1966*a*) have previously discussed the anomalies associated with James Ross

Island and its vicinity, and they attributed them to the olivine-basalt lava flows belonging to the James Ross Island Volcanic Group. No further comment is required at this stage except to draw attention to the postulated northern boundary of the James Ross Island Volcanic Group in Duse Bay (Ashley, 1962). This boundary was tentatively defined by the difference of magnetic character between the north and south. Ashley also remarked, however, that a considerable thickness of lavas would be required beneath Duse Bay to allow their detection at sea-level. With the additional bathymetry calculated from the gravity data, it would now appear that the decrease of magnetic intensity must be partially affected by the increasing distance from the source rocks, and not necessarily due to the presence or absence of the individual rock types.

The Seal Nunataks lie 150 km. south of James Ross Island and they are exposed through the ice shelf as agglomerates, olivine-basalt lavas and dykes. Two noticeable exceptions are Pedersen Nunatak and Robertson Island which are composed of Cretaceous sediments. Up to the present time, all of the volcanic representatives visited exhibit a characteristic anomaly trending parallel to the geographical distribution of the respective host nunataks. They appear to be normally magnetized and the anomalies reach at least 3,000 gamma in amplitude. Detailed vertical-field surveys over Evensen Nunatak have revealed anomalies in excess of 6,500 gamma. A three-dimensional interpretation correlates the anomalies with the olivine-basalt representatives but suggests as a source pipe-like bodies whose upper surfaces are expressed as the nunataks themselves. Although this does not conform to the observed dykes as would have originally been suggested, it does appear that the respective bodies dip at the same angle and must be interrelated. Evensen Nunatak has several anomaly peaks, some of which are outside the visible physical boundaries. It is suggested that these are due to either a variable topography beneath the ice cover or are associated with subsidiary dykes. Gravitational evidence from this area indicates a general rise in bedrock towards the vicinity of the nunataks.

The magnetic character of the Larsen Ice Shelf contrasts sharply with that of the mainland. Small-amplitude, large wave-length anomalies indicate an area devoid of magnetic material and implies a gently undulating bedrock topography possibly covered by a thick layer of low-susceptibility sediments. On the west coast, the much smaller Wordie Ice Shelf exhibits stronger magnetic intensities with amplitudes up to 1,800 gamma. This is in sharp contrast to the situation found on the Larsen Ice Shelf though examination of the surface relief of the Wordie Ice Shelf shows it to be extremely irregular. Ice rises, severe crevassing and rifting suggest a highly disturbed bedrock surface which could account for the anomalous zones.

The gravity data have shown that Graham Land is associated with a negative Bouguer anomaly (Davey and Renner, 1969), the value becoming more negative towards its longitudinal axis and the farther south in latitude. A minimum recorded value of -60 mgal was measured at lat. 68° 30′ S. Crustal structures at depth are suggested as the main contributory cause of this although a major topographical feature is apparent in the area. Until further control is available it is not possible on the present knowledge to estimate crustal thicknesses. The values at the northern end of the peninsula (Davey, 1971) may be given as an indication of the order of increase, although they are based on a minimum of representative data. These values show an estimated crustal thickness of around 21 km. at Hope Bay and this increases to at least 32 km. in a section 35 km. to the south-west. At the base of the peninsula, thicknesses of the order of 36 km. have been proposed by Behrendt (1964).

Superimposed on the broad regional gravity trend are several smaller local anomalous areas. These at present are essentially restricted to the east coast of Graham Land, where more detailed land traverses have been undertaken. It has not been possible to isolate accurately residual anomalies because of the uncertainties in the regional field. Therefore, in the interpretation a regional field has been extracted which the author considers to be the most justified over any given anomaly. The majority of the residual anomalies have been attributed to variations in bedrock topography below the ice. No allowance has been made for possible unconsolidated sediments and therefore calculated depths to bedrock are minimum values. Beneath the Larsen Ice Shelf, two basic geomorphological variations are present within the otherwise gently undulating sea floor, here estimated at an average depth of 500 m. The first concerns areas of marked overdeepening, and the other areas of bedrock elevation.

Where the traverses have penetrated the inlets, the gravity values become more negative. Typical of this are Solberg, Trail, Seligman, Whirlwind, Mill, Cabinet and Larsen Inlets. A similar situation exists where major glacier systems enter the ice shelf directly without an inlet phase. The area to the north of Philippi Rise and to the east of Drygalski Glacier are examples of this. Two-dimensional interpretational methods have given depths to bedrock, which in the case of Trail Inlet are of the order of 900 m. This is considerably greater than the surrounding continental-shelf depths of 500 m. and it has been suggested that erosion during the glacial maximum was responsible for the overdeepening. Radio echo-sounding (Renner, 1969; Smith, 1972) over many of the inlets has indicated increased ice thicknesses and the even greater values during the glacial maximum could easily have led to grounding and erosion to the depths calculated.

A similar situation exists in Prince Gustav Channel at the northern end of Graham Land. This area has a debatable ice shelf but the additional presence of sea ice has enabled the continuity of gravity traverses across the channel and has provided land sites along the coast to which the traverses can be tied. As expected, the negative anomalies parallel the topographical trend of the channel and, although the area is geologically complex, the anomalies have been interpreted as due to the bathymetry. At the southern end of the channel, maximum depths of over 800 m. were calculated and they are similar to the values measured 40 km. farther to the north by ship soundings. Duse Bay, at the northern end of Prince Gustav Channel, has yielded a surprising result. A detailed gravity survey completed over this area resulted in unusually large negative gravity anomalies with residual values down to -75 mgal. The anomaly distribution suggested a three-dimensional interpretation was required and this gave depths to bedrock of 2,400 m. below sea-level. Although depths of 1,200 m. have been sounded off the west coast, there are no recorded depths of the magnitude of those calculated in Duse Bay anywhere else on the surrounding continental shelf. The area has had a complex geological history, both stratigraphically and structurally, but, even after considered allowance for these, it would appear that a sizeable negative anomaly would still remain centred over Duse Bay. Unfortunately, fast ice over the bay has prevented marine echo-sounding but where ships have had partial access the sounded depths show good agreement with those calculated from the gravity readings over the same areas. Glacial activity is again a suggested cause for the deepening in Duse Bay.

Adjacent to some of the peninsulas and promontories, the gravity values show an increase over the local background. These have been interpreted as areas of grounding and their extent in the field is often indicated by associated surface irregularities in the ice shelf.

Local anomalous areas on land have been more difficult to resolve due to the elevation uncertainties and the often unknown ice–rock distribution beneath a given station. Over some localities, however, anomalies have been sufficiently large to enable at least some qualitative evaluation to be made.

Joerg and Churchill Peninsulas have associated positive Bouguer anomalies. At the former locality, it has been proposed that the presence of sediments of the Trinity Peninsula Series with a measured density of $2 \cdot 74$ g. cm.$^{-3}$ are responsible. Further work is necessary over this area before more definite conclusions can be reached, but in support of the present idea is the absence of any magnetic anomalies which would be expected if the causative body was a basic intrusion. At Churchill Peninsula, the gravity anomaly is associated with a strong magnetic anomaly of over 1,500 gamma. Interpretation suggests an asymmetrical laccolithic body whose upper surface must lie within a few metres of ground level. Geological evidence (Marsh, 1968) supports this interpretation.

Negative Bouguer anomalies are found over Jason Peninsula and on the mainland to the west, as well as over southern Tabarin Peninsula. On Jason Peninsula their cause is probably the Upper Jurassic acid volcanic rocks which are known to be exposed, whilst on Tabarin Peninsula the answer may lie in the less dense tuffs and agglomerates belonging to the James Ross Island Volcanic Group. The area of low gravity west of Jason Peninsula provides a slightly different situation for here both geological and topographical factors may have some influence. Analysis of the data has suggested that a sub-sea-level channel exists between Adie Inlet and the ice shelf to the north of Philippi Rise. Should the ice melt, Jason Peninsula would therefore be separated from the mainland. The southern part of Churchill Peninsula also exhibits a gravity field slightly lower than the estimated regional. Examination and approximate depth calculations over Churchill Peninsula suggest a variable bedrock relief with much of the rock surface below sea-level. Recent ice-thickness profiles obtained over Churchill Peninsula (Smith, 1972) support these findings.

Starbuck Glacier lies immediately to the south of Cape Disappointment. East of its junction with the Larsen Ice Shelf there is an interesting series of north–south rifts (Fleet, 1965). Some doubt has been expressed over the values of surface elevations in the area as discussed by Kennett (1965*c*) and Fleet (1965). It may be, however, that both estimates are substantially correct as the gravity data suggest that grounding of ice may occur west of the rifting. If this is so, the elevation estimate of Fleet, at first seemingly low, could be due to local variation. This situation could arise if the elevation was estimated in an area near the junction between grounded and floating ice, and where flexing and thinning perhaps due to tidal oscillations has taken place.

Neny Trough, a major structural feature laterally bisects the Antarctic Peninsula and, although some systematic geological research has been completed over part of its length, its true significance has not yet been diagnosed. On the assumption that the residual gravity anomalies are the result of variation in depths to bedrock, values of 800 m. below sea-level were calculated for distances up to 30 km. west from the mapped boundary between the land ice of Gibbs Glacier and the ice shelf in Mobiloil Inlet. Although it is considerably below sea-level, the ice is believed to be grounded inland of the "accepted" boundary. The magnetic results show that the structure has a distinct magnetic character. Total-field magnetic traverses to the north and south reveal a fairly active pattern but over the Mercator Ice Piedmont and Gibbs Glacier anomalies of less than 60 gamma are observed. Gravity traverses in central Mobiloil Inlet show no obvious increased depths to bedrock which would have indicated a possible eastward extension of the structure. It is felt, therefore, that if the topographical feature does continue to the east, its sub-surface development must lie along the southern part of the inlet. Faulting has already been suggested (Adie, 1957) between Kenyon Peninsula and the mainland which strikes along the course of a possible continuation of the Neny Trough feature. Severe crevassing of Neny Glacier has so far restricted geophysical and geological work at the western end of Neny Trough.

Three Project Magnet flight lines over the peninsula have been analysed. On a regional basis, they verify the basic anomaly pattern revealed by the land and inshore marine magnetic traverses. This shows that Graham Land is broadly subdivided into three distinct magnetic provinces with boundaries paralleling its length. The eastern zone consists of a flat, relatively non-magnetic region coincident with the Larsen Ice Shelf. Over the Antarctic Peninsula itself, the anomalies indicate an area of greater magnetic activity reflecting the underlying intermediate to basic members of the Andean Intrusive Suite. This is particularly noticeable along the coastal sections where such rocks are known to be exposed. Along the Beak Island to Cape Legoupil profile, however, little magnetic activity is observed over the peninsula. This was also found on the equivalent land surveys where it was attributed to the increasing thicknesses of the non-magnetic Trinity Peninsula Series sediments. The largest area of magnetic disturbance occurs along and adjacent to the west coast of Graham Land. Over this area the flight traverses revealed a magnetic anomaly pattern which was also identifiable on all of the marine magnetic traverses, here located north of lat. 68° S. The profiles indicate that a sharp increase in magnetic activity occurs off the west coast, and upon this is superimposed a distinctive magnetic pattern. In general, the smaller wave-length disturbances within these zones increase in amplitude southward and they are thought to be associated with the near-surface basic igneous rocks known to crop out on the offshore islands. Geological observations on the islands have recognized a variety of basic rocks including dolerites, gabbros and basalts.

The geophysical evidence from the west coast of Graham Land suggests that the group of offshore islands paralleling the mainland are not simply an integral part of it which has been separated as a consequence of isostatic submergence. The pronounced increase in magnetic activity on all measured profiles indicates that some major disconformity is present. The interpretation of the magnetic anomaly adjacent to Graham Land favours a two-dimensional upfaulted block of basic igneous material striking parallel to the coast. The seismic data of Ashcroft (1972) and gravity results of Davey (1971) have also suggested the existence of such a structure. Examination of the marine magnetic results to the north-east of Graham Land suggests that at least some aspects of the magnetic zoning continues for a further 160–200 km. to long. 53° W. From magnetic, gravity and seismic investigations (Griffiths and others, 1964; Davey, 1971; Ashcroft, 1972), it has already been determined that the north-western boundary of Bransfield Strait is a fault zone with a northerly upthrow against the South Shetland Islands. It appears that the south-eastern boundary of Bransfield Strait has a similar structure, at least along north-western Graham Land where upfaulting to the south against the mainland has taken place. Geological research off Cape Legoupil (Halpern, 1964) has provided further evidence of major north-east to south-west faulting.

Over north-east Trinity Peninsula, composite regional trends associated with Prince Gustav Channel are being clarified. Here, the gravity gradient towards James Ross Island is atypical of that shown over the remainder of the east coast of the Antarctic Peninsula. Structural control predominates but the less dense James Ross Island Volcanic Group rocks must contribute to the easterly decrease in gravity. Similarly, faulting across Tabarin Peninsula, again involving the James Ross Island volcanic rocks, has been indicated from geophysical measurements. Both geological and geophysical observations support a possible line of north-west to south-east faulting through Broad Valley and Misty Pass. At the southern end of Graham Land the Neny Trough feature has been previously discussed as also have the structural implications of the southern plateau magnetic traverses.

It is difficult to formulate an opinion on the isostatic conditions existing over the whole region. Reference to the southern Antarctic Peninsula work of Behrendt (1964) suggests that it is one of incomplete local compensation, while the same author (Behrendt, 1962), working on the Filchner Ice Shelf, concluded that the latter area was one illustrating isostatic balance. At the northern end of Graham Land, more specifically the Bransfield Strait area, Davey (1971) has calculated that this area is not fully compensated. It is debatable to what extent compensation could be attributed to local as against regional control. The dimensions of the area are within those postulated as being under regional influences, but if local compensation has to be considered, either through geological causes or due to the ice overload, it would then appear that this area is one of undercompensation, and therefore submergence should be in evidence if the process is continuing to completion.

In this respect, Adie (1964*b*) has already drawn attention to the physiographical evidence indicating that Graham Land is in fact a region of submergence. Likewise, the removal of the ice cover could be expected to produce some isostatic rebound and supporting evidence such as raised beaches has been identified (Adie, 1964*b*; John and Sugden, 1971). Isostatic movements must still be in operation for the ice overload but it will be difficult to separate these from any imbalance which may prevail due to the solid geology.

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