GRAVITY SURVEY IN NORTHERN MARGUERITE BAY, GRAHAM LAND

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ABSTRACT. The results of several years' gravity survey in northern Marguerite Bay are compiled and interpreted. Particular study is made of an igneous complex on Horseshoe Island and of several glaciers near Stonington Island.

OVER recent years, geophysical surveys incorporating gravity and magnetic methods have been carried out along the coast and hinterland north from Stonington Island towards Arrowsmith Peninsula and Adelaide Island. The first gravity survey was conducted by P. Kennett (1965), who occupied 17 stations in traverses from Stonington Island to Horseshoe Island and Bourgeois Fjord. This work was continued by the author in 1968–69 and later extended by F. M. Burns and P. F. Butler. Both regional and local gravity traverses have been completed and they have been used in the compilation of this paper. The results of a survey on Shoesmith Glacier, Horseshoe Island, have already been published (Smith, 1973).

PHYSIOGRAPHY AND GEOLOGY

The area surveyed has a very irregular topography. Steep-sided islands to the west give way to a deeply indented coastline with fjords penetrating the mainland by as much as 20 km. Some of these fjords are partially or wholly filled by ice of unknown thickness and this may show heavy crevassing locally. The area to the south of Square Bay is typified by such ice-filled fjords, whereas to the north they are only ice-filled at their heads. The sub-parallel rock walls up to 2,000 m. in height which border the fjords and their associated glaciers lead up to the ice-capped plateau of Graham Land.

Detailed descriptions of the geology of this area have been published previously and a brief outline only is given here. In the south of this area, gneisses and several suites of granitic intrusive rocks are overlain by generally flat-lying Upper Jurassic volcanic rocks. Farther north, the gneisses are less common but the volcanic rocks become widespread and they are cut by gabbroic and dioritic intrusions which are followed by the intrusion of granites of a late Andean age.

GRAVITY SURVEY

Worden gravimeters Nos. 556 and 743 were used for the survey; their calibration factors were 0·2297(6) mgal/scale division at 78° F [25·5° C] and 0·2038(9) mgal/scale division at 84° F [29° C], respectively.

Several different procedures were adopted in the collection of the data. The first method involved back-packing the gravimeter between closely spaced stations arranged along profiles. At the beginning and close of each day the gravimeter was read at a base gravity station, either at a fixed camp or a station hut. Station positions were determined by compass resection on to fixed points and tied in to the end station of the profile, while station elevations were measured concurrently by tachymetry.

For the longer geophysical traverses over the sea ice, the gravimeter was suspended between the handle-bars of a lightly laden dog sledge. Readings were taken whenever a landfall was made, since tidal oscillations usually prevented gravity readings from being made on the ice. The gravity loops were closed at night to a base station. Station elevations were estimated from sea-level at Mean High Water and positions were again determined by compass resection.

The third method involved travelling with a laden dog sledge on long journeys and readings were taken at intervals of 1–2 km. on land ice or rock. The closure period between base-station readings was up to a month. Station positions were determined by the compass–sledge-wheel method and their elevations were determined by using barometers.

In the Horseshoe Island and Northeast Glacier areas, as shown in Fig. 1 (locations of

Figs. 2 and 3), the gravity data were collected systematically. Detailed profiles were measured over a small granite body surrounded by gabbro in the vicinity of the station hut at Horseshoe Island. The gabbro body was surveyed in sufficient detail to enable its form to be determined. On the Northeast Glacier survey, which was undertaken jointly with a reconnaissance topographical survey party, measurements were taken along profiles enabling the depth to bedrock to be determined.

Data Reduction and the Errors Involved

The instrumental drift was removed by assuming a linear drift rate between base-station readings and allowing, where appropriate, for static drift. Elevations were corrected to a sea-level datum using an average crustal density of 2·67 g. cm.$^{-3}$, and terrain corrections were applied by using the graticule method described by Bible (1962). Latitude corrections were carried out as normal.

The error in the instrumental drift for the back-packing method was computed from the variation from the mean gravity value of a field station which was re-occupied on seven occasions during the course of the survey. From this the error was calculated to be ±0·3 mgal. Where no field stations were re-occupied, as in the second and third methods described above, the errors could not be determined in the same way. Therefore, the error in the assumption of linear drift was estimated from the standard deviation of the drift rates of 11 separate days' work when closures to Horseshoe Island had been obtained. The standard error obtained by this method was ±2·8 mgal over a 10 hr. working day. This figure indicates the variable and rough travelling conditions experienced during the survey rather than the error arising from drift on any one day, which is suggested to be about ±0·5 mgal.

It was possible to estimate the positions of the stations on Horseshoe Island to such an accuracy that the error in latitude correction was negligible. On the other surveys, however, the uncertainty in position depended on the size of the triangles of error produced by the compass resections. This was estimated to be ±0·1 km. which is equivalent to an error of ±0·12 mgal.

Errors in the elevation correction resulted from incorrect station heights and from unknown density distributions in the underlying rocks. Where the heights were found by tachymetry and closed to sea-level, the error was ±4 m., which is equivalent to a maximum error of ±0·43 mgal. If found by barometric methods, the estimated error was 15 m. which corresponds to an error of ±2·0 mgal. It was not possible to estimate the error introduced in the Bouguer anomaly by using the mean crustal density, because there was insufficient information on the relative distribution of the local rocks whose densities would otherwise have been used in the data reduction.

Errors in the terrain correction arose from the uncertain representation of the topography as read from D.O.S. 1 : 200,000 sheets W6766 and W6866. These were less accurate in the north than in the south of the area and corresponding errors ranged from ±0·8 to ±0·2 mgal. Most of the error arose from the near zones but this was reduced as much as was possible by careful positioning of the field stations.

Thus the total standard error is very variable over the area depending on how the data were acquired, but it is estimated that the minimum standard error is ±0·6 mgal and the maximum is ±3·6 mgal.

Interpretation and Results

The Bouguer anomaly map compiled from the corrected data is shown in Fig. 1. The contouring is based largely on stations on or near rock, since stations on ice would tend to confuse the regional picture conveyed by this map. Where few rock stations are available, ice-station information has been considered as on the traverse on Adelaide Island (Fig. 1) and in the area shown in Fig. 3.

Interpretation is concentrated over the areas shown in Figs. 2 and 3.

Fig. 2 shows a gravity high associated with the outcrop of a gabbro intrusion on Horseshoe Island. The gabbro body was delineated from the observed geology and magnetic anomalies, and it is seen to have a sub-circular outcrop of radius 2·5 km. centred 1 km. south of the station hut.

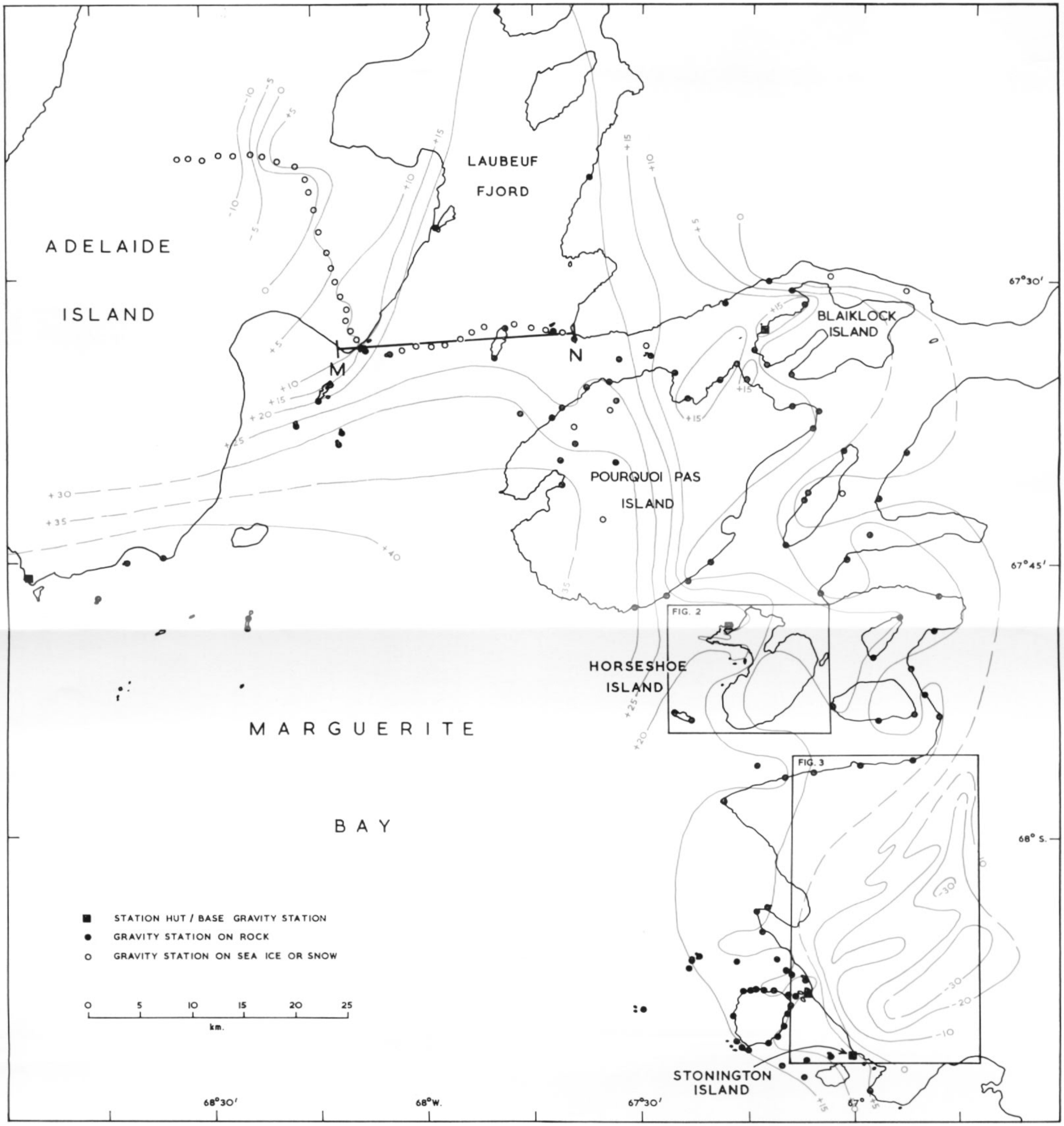


Fig. 1. A Bouguer anomaly map of northern Marguerite Bay. Contours are at 5 or 10 mgal intervals. The locations of Figs. 2 and 3 are shown.

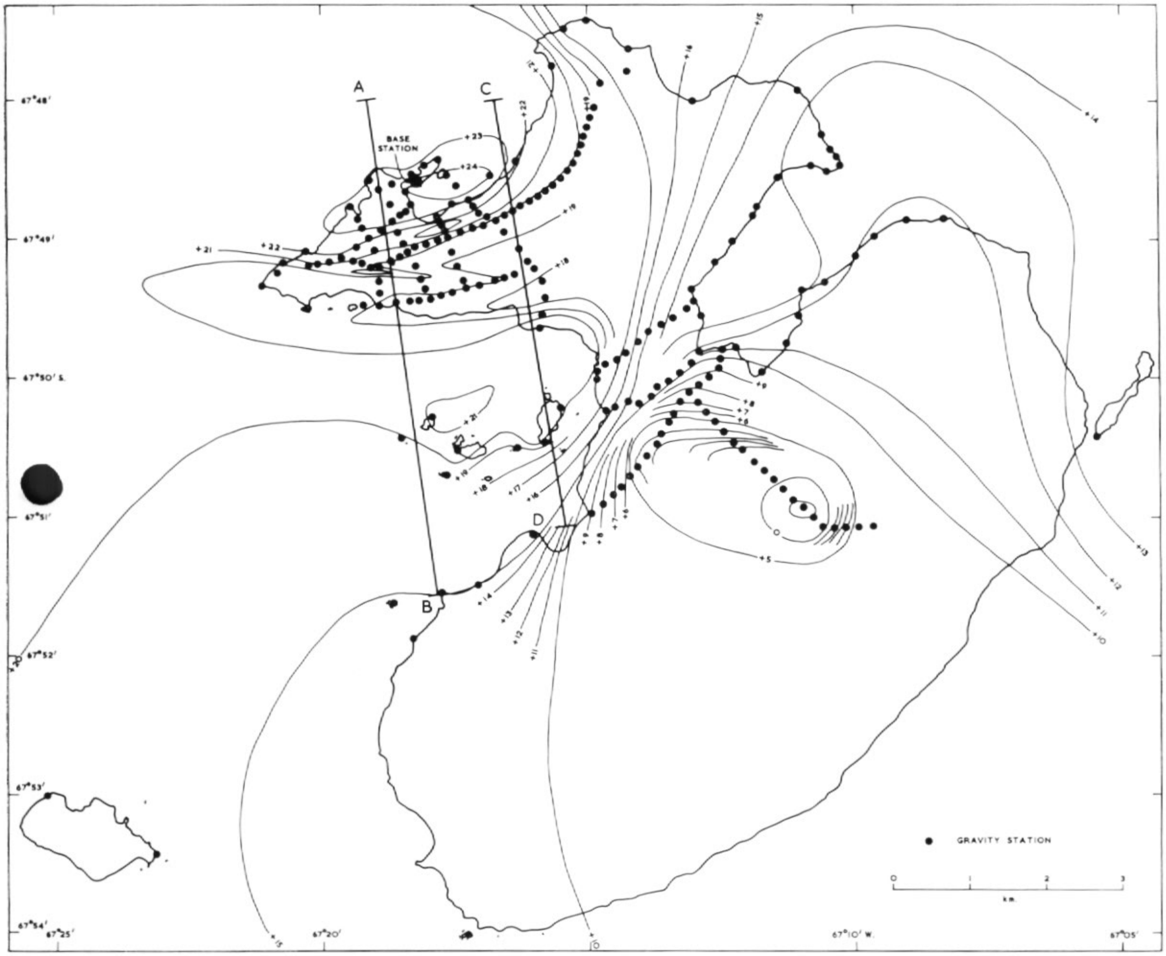


Fig. 2. A Bouguer anomaly map of Horseshoe Island. The contour interval is 1 mgal. Profiles A–B and C–D refer to the text and they are shown in Figs. 4 and 5.

The densities as determined from samples collected during the survey are: gabbro 2·93 ±0·02 g. cm.$^{-3}$ from 15 specimens, and granite 2·57 ±0·02 g. cm.$^{-3}$ from four specimens.

It was assumed that the high Bouguer anomaly values are due to the gabbroic body and the regional residual separation was made on this basis. A gravity profile was drawn across the centre of the body parallel to the strike of the regional field. The resulting residual anomaly was symmetrical with a maximum value of 15 mgal.

It was apparent from this evidence that the gabbro body might be approximated to a vertical cylinder of radius 2·5 km., and the effect of this was computed at several points on the profile using the standard formula (Heiland, 1946, p. 147). Agreement was reached to within 1 mgal for a cylinder with its base at about 2·25 km. depth using a density contrast of 0·26 g. cm.$^{-3}$, which represents the difference in density between the mean crustal density and that of the gabbro.

There were insufficient data for a quantitative interpretation to be carried out on a profile perpendicular to the strike, but it does appear to be compatible with the first interpreted profile.

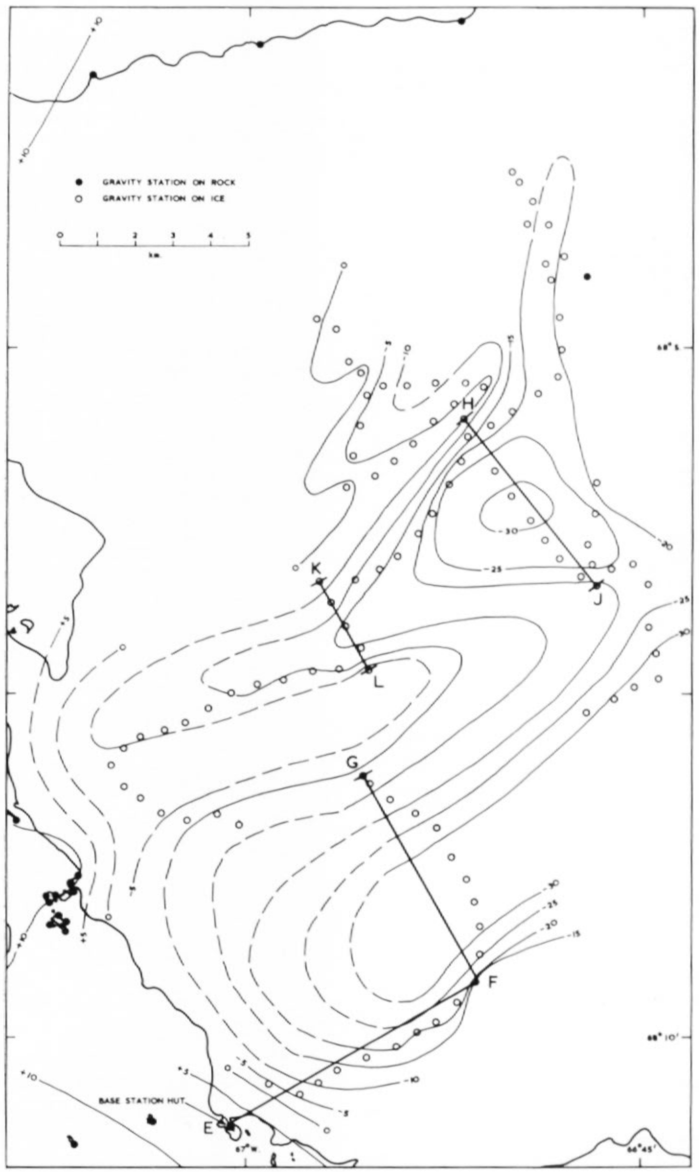


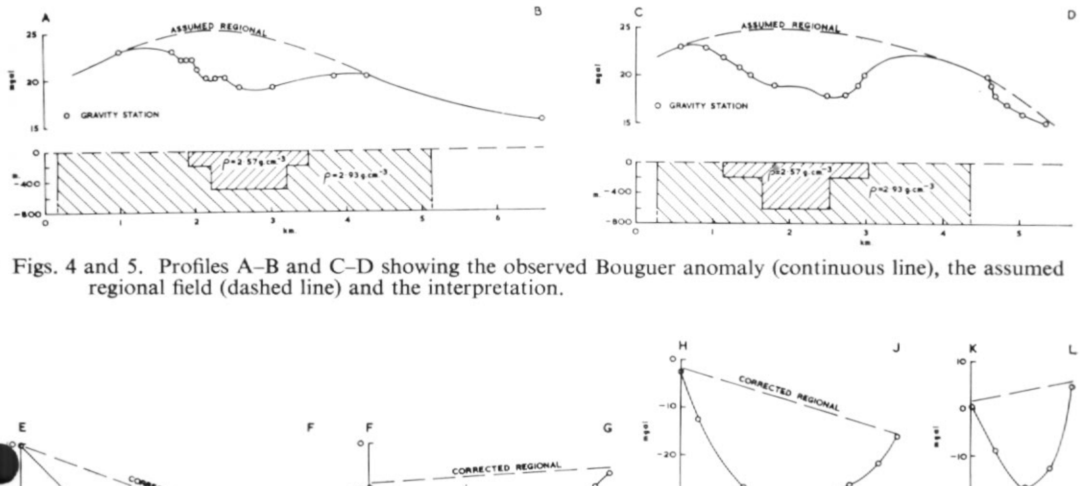
Fig. 3. A Bouguer anomaly map of part of the hinterland. The contour interval is 5 mgal. Profiles E–F, F–G, H–J and K–L refer to the text and they are shown in Figs. 6 and 7.

The Bouguer anomaly high associated with the gabbro is modified by an east–west trough which closely follows the outcrop of the gabbro–granite contact, and therefore the granite was assumed to be the sole cause of the anomaly. To obtain the regional field, a smooth curve produced from the interpreted gabbro body was drawn through and removed from two profiles across the granite feature. The residual anomaly was symmetrical to within 0·5 mgal and, since the northern contact was exposed and seen to be vertical, it was possible to estimate the width of the body which was found to be 1·6 and 1·9 km. on profiles A–B and C–D, respectively. Various two-dimensional models were constructed and using a gabbro–granite density contrast of 0·36 g. cm.$^{-3}$ the associated anomalies were calculated using a graticule method. The models giving the closest fit to the observed anomalies are shown in Figs. 4 and 5. The calculated and observed profiles agree to within 0·5 mgal.

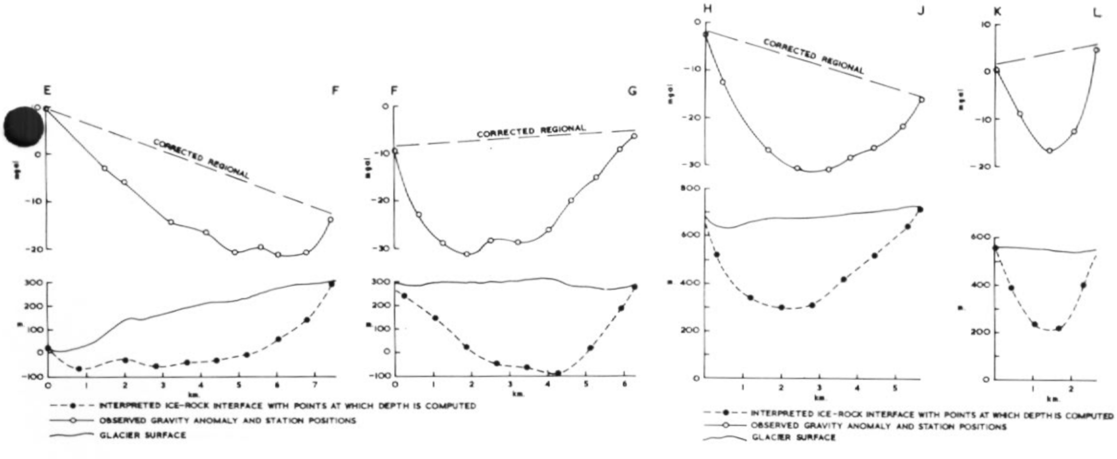
An enlargement of part of Fig. 1 is given in Fig. 3, and this shows the Bouguer anomaly map of the area to the north and east of Stonington Island. Several traverses were measured across glaciers from stations very close to the rock walls. Unfortunately, in this area rock stations were inaccessible but it was considered justifiable to estimate the thickness of ice beneath the end station of a profile by downward extrapolation from the adjacent exposed rock. Thus, knowing the thickness of ice, a Bouguer anomaly was calculated which was considered equivalent in value to the station being on the rock.

It was assumed that the variation in Bouguer anomaly along the profile was solely due to the thickness of ice in the glacier. A linear regional gradient was drawn between the equivalent rock stations at each end and subtracted from the observed profile. The residual anomaly was then used to determine ice thickness based on the method described by Bott (1960). In the original method, depths were computed below a horizontal plane but this was modified to enable depths to be computed beneath stations of any given heights.

The results from these traverses are shown in Figs. 6 and 7. Fig. 8 gives the results from the traverse across Laubeuf Fjord. The echo-sounding profile obtained by R.R.S. *John Biscoe* has been corrected for temperature and salinity using Matthews's (1939) tables, and this is compared with the profile determined by the Bott method on data from which a straight-line regional has been removed.



Figs. 4 and 5. Profiles A–B and C–D showing the observed Bouguer anomaly (continuous line), the assumed regional field (dashed line) and the interpretation.



Figs. 6 and 7. Profiles E–F, F–G, H–J and K–L showing the observed Bouguer anomaly, the corrected regional field, the glacier surface and the interpreted ice–rock interface.

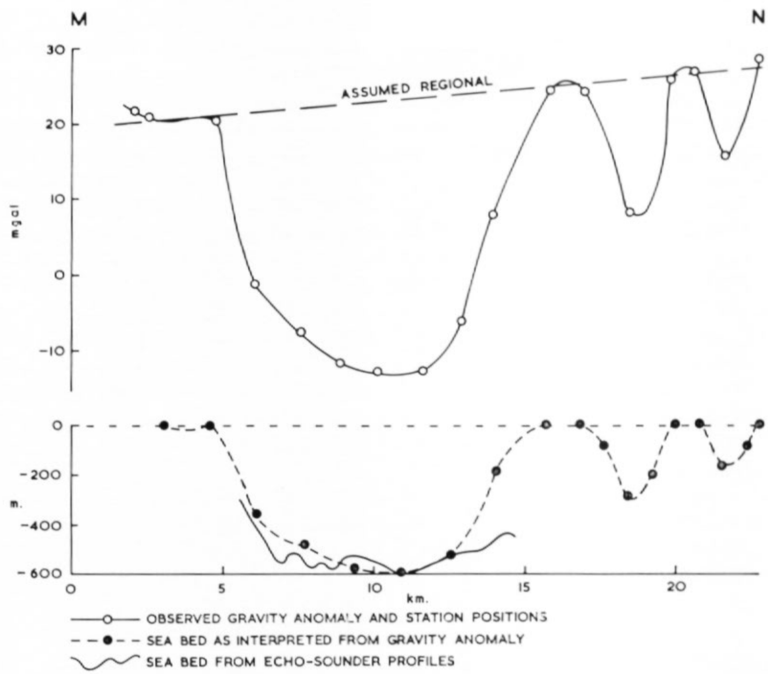


Fig. 8. Profile M–N across Laubeuf Fjord showing the observed Bouguer anomaly, the assumed regional field, the echo-sounder profile and the interpreted sea floor.

Discussion

The regional Bouguer anomaly field over this area closely follows the main trend of the coastline. This suggests that there is an obvious correlation between the major geological structures which are responsible for both the field and the coastline. Further data are necessary before any interpretation on this scale is possible.

Interpretation of the Horseshoe Island survey is consistent with the geology, the cylindrical body of gabbro being a good approximation to a ring boss, possibly suggesting intrusion of the gabbro by cauldron subsidence. The wedge of granite might represent either a roof pendant or a later intrusion depending on the relative ages of the rocks.

Points where the computed anomaly due to the granite deviate from the observed anomaly are closely correlated with stations observed on the small permanent icefields on the island and which give small negative anomalies associated with ice thicknesses.

Profile M–N across Laubeuf Fjord shows that the direct interpretation method is capable of giving results which compare well with a proved interface. However, the errors are rather larger at the edges, since the initial assumption that the profile may be approximated to a set of infinite slabs (Bott, 1960) is invalid if the interface is steeply dipping. Small errors may also arise because of the unknown thickness of morainic deposit on the sea floor, which is ignored in the gravity interpretation; therefore, the depths determined by this method may be too great.

It is therefore expected that the ice–rock interfaces determined for the glaciers are quite reliable. Profile H–J is expected to be less accurate because the glacier is narrow and steep-sided, and the edge effect referred to above may be important.

The results show that the bed of Northeast Glacier is below sea-level for about 6 km. from Stonington Island along profile E–F, and also for most of its width on profile F–G. This suggests that much of the bed of the lower reaches of the glacier may be below sea-level, except where crevassing suggests an uneven bed. This deep channel or fjord may extend well up towards the plateau.

ACKNOWLEDGEMENTS

I wish to express my thanks to F. M. Burns and P. F. Butler as well as our field assistants for their help in collecting the data, and to R. G. B. Renner for his invaluable criticism during the preparation of the paper.

MS. received 22 February 1973

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