Reprint Norges Geologiske Undersøkelse nr. 214.

Norges Geologiske Undersökelse

Nr. 214 B

Granite Studies: I. A Gravity Investigation of Two Precambrian Granites in South Norway.

By

Scott B. Smithson

With 19 Text-figures and 2 Plates

(p. 53—140)

OSLO 1963

Granite Studies:
I. A Gravity Investigation of Two Precambrian Granites in South Norway.

By

Scott B. Smithson

With 19 Text-figures and 2 Plates

Contents.

Abstract.

The Bouguer gravity anomaly map, based on 778 stations, covers the Precambrian rocks along the southeast coast of Norway from the Oslo region to Kristiansand. The principal features of the map are the low gravity gradient over the Telemark rocks and the steep gradient over the Bamble rocks, upon which are superposed the anomalies of smaller geologic bodies. The Bamble anomalies, which increase toward the coast, attain a value of + 50 mgal along the coast. Negative gravity anomalies occur over the late-to postkinematic Herefoss and Grimstad granites and the synkinematic Oddersjå granite, and a positive gravity anomaly occurs over the nickeliferous Evje amphibolite.

The gravity gradients over the Telemark and Bamble rocks change at the trace of the "great friction breccia" that separates the Telemark and Bamble rocks. A wedge of more dense material that thickens from the breccia to the coast is postulated as the source of the steep Bamble gravity gradient. The wedge corresponds to the succession of Bamble supracrustal rocks overlying the Telemark granitic gneisses. The throw at the breccia zone is calculated to be a minimum figure of 0.5 km.

A residual gravity anomaly map is presented for the Herefoss and Grimstad granites and vicinity. The Grimstad granite has an anomaly of —13 mgal and is approximated by a cylindrical gravity model that has a moderately dipping north contact and a thickness between 2.6 and 4 km. The larger Herefoss granite has an anomaly of —7 mgal, but its density contrast is smaller. The gravity model of the Herefoss granite is composed of a large disc underlain by a smaller disc that is offset to the east. This gravity model suggests that the Herefoss granite has the shape of a funnel; this interpretation agrees with the attitudes of contacts and the foliation of the granite. The calculated thickness is between 2 and 5 km. Smaller negative anomalies within the north-central part of the Herefoss granite are caused by the large amount of foreign material incorporated in the granite here, and a mappable inclusion causes a small positive anomaly within the granite. Differences in both mean density and variability, which are correlated with depth of exposure of the granites, are found between the Herefoss and Grimstad granites. The Herefoss and Grimstad granites are

conceived to be mobilized portions of the substratum that intruded the Bamble rocks and whose dissimilar characteristics are attibuted to different levels of exposure and tectonic framework.

The synkinematic Oddersjå granite has an anomaly of + 2 mgal and a calculated thickness of 1.4 km. The Evje amphibolite has an anomaly of + 12 mgal and a calculated thickness of 1.25 km.

A comparison between salt dome tectonics and granite emplacement offers the following advantages: (1) The room problem is solved for certain granites. (2) The structural relations of granites are explained. (3) The gravity profiles of granites are explained. Granites have characteristic gravity profiles that appear to be only a function of the mass deficiency of the granite itself. Any mass excess, which could be either the displaced country rocks or mafic differentiates of granitic magma, must be dispersed because it is not reflected in the gravity profile. Granites of subvolcanic regions, however, may be associated with broad positive gravity anomalies that could be caused by a mafic residuum.

Introduction.

Purpose

In any interpretation of the origin of granites, the shape and volume of the granitic body are important properties that should be considered. The geologist, who is normally restricted to either 2-dimensional exposures or 3-dimensional exposures of limited extent, must extrapolate downward from the surface. The interpretation of a gravity survey can furnish 3-dimensional information. A gravity survey cannot only reveal the shape, volume, and mass deficiency of a granite but also, theoretically at least, detect such controversial objects as mafic differentiates and "basic fronts".

If there is a gravity anomaly over a granite, then a hypothetical model that simulates the gravity effect of the granite can be computed. The occurrence of negative gravity anomalies over granites is well known (Reich, 1932; Romberg and Barnes, 1944; Garland, 1950, 1953; Goguel, 1950; Bean, 1953; Bott, 1953; Grosse, 1958; Bott and Masson-Smith, 1960). Bott (1956) has even treated the granite problem strictly from a geophysical veiwpoint. The purpose of this investigation is to determine the gravity field of two closely associated granites, the Herefoss and Grimstad granites; to compute the most probable model for them; and to draw petrogenic conclusions from these results.

Location

The area studied is part of the Precambrian basement that forms the southern tip of Norway and extends from the Permian Oslo graben southwestward along the coast to Kristiansand (Fig. 1). The area lies between 58° 00′ and 59° 15′ N. latitude and 8° 00′ and 10° 00′ E. longitude. The region along the coast is called the Bamble area, and the interior is called Telemark so that the rocks have received the names, Bamble rocks and Telemark rocks respectively.

Previous Investigations

The rocks of this area have become well known through the investigations of Barth (1929, 1935, 1947a, 1947b, 1956), A. Bugge (1928,

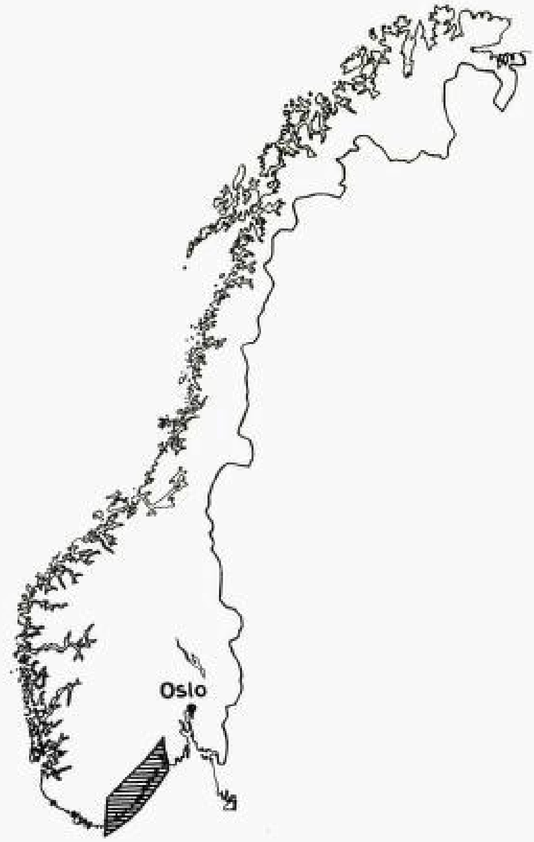


Fig. 1. Map of Norway showing the area studied.

1936), and J. A. W. Bugge (1943). Oftedal has studied field relations and petrography of the Grimstad (Fevik) granite (1938, 1945). More recently under the direction of Professor Tom. F. W. Barth, the tectonics of the Herefoss granite and vicinity has been studied by Elders (1961, 1963), and the petrography and feldspars of the Herefoss granite have been studied by Nilssen (1961). The origin of gneisses between Lillesand and Kristiansand was treated by Dietrich (1959, 1960). Trace-element distribution in alkali feldspars from the Herefoss and Grimstad granites and the surrounding gneisses has been investigated by Heier and Taylor (1959). This area, then, is one of the better studied, better exposed areas in Norway, a fact that makes it well suited for a gravity investigation.

No interpretive gravity survey had been attempted; however, a net of stations, located along major roads, has been laid out by the Geographical Survey of Norway (Norges geografiske oppmåling). Four submarine gravity stations were established off the coast by Collette (1960) as part of a survey of the North Sea. The present survey was intended to locate additional stations in areas of particular geologic interest.

Acknowledgments

The writer is greatly indebted to Professor Tom. F. W. Barth for discussing this project and for the use of the facilities of the Geologisk Museum. Sincere thanks are due to the Geographical Survey of Norway (Norges geografiske oppmåling) and Mr. O. Trovaag and Mr. G. Jelsdrup of this organization whose wholehearted cooperation made this study possible. Mr. Th. Sömod and Mr. O. Mathisen receive thanks for their diligent performance in measuring supplementary gravity stations in the area sudied.

The Geologisk Museum of the University and the Geological Survey of Norway (Norges geologiske undersøkelse) provided financial support for this study. Mr. Thorolf Nordby is sincerely thanked for his excellent help with the field work. The writer thanks Mr. Thor Siggerud for help with the field work and Miss Ingrid Lowzow for drawing the illustrations. Mrs. B. Nilssen and Dr. W. O. Elders kindly placed the results of their investigations at the disposal of the author and discussed the problems. Professor R. V. Dietrich and Dr. K. S. Heier critically read the manuscript. This project was completed with the support of a Postgraduate Fellowship of the U.S. National Science Foundation.

Gravity Procedure and Theory

The section concerning field procedure, computations, and accuracy of the gravity survey appears in Appendix A. In addition, a short explanation of the principles of gravity interpretation and methods is included in Appendix A so that the non-geophysicist reader will be able to follow and understand their development in this manuscript.

General geology.

Introduction

The region studied is part of the Precambrian basement of South Norway that is well known for its interesting rock types and relatively good exposures. The rocks are a series of highly metamorphic crystalline schists, gneisses, and migmatites associated with subordinate amount of plutonic rocks (Plate 2). A fault zone that runs NE-SW parallel to the coast separates rocks of slightly different character and divides the region into two parts, the Bamble area and the Telemark area. The geology of the area studied has been summarized by Barth (1960).

Bamble Area

The Bamble area extends from the Oslo graben on the northeast to Kristiansand on the southwest. The northwest border is delineated by the fault zone, the "great friction breccia", and the southeast border is marked by the Skagerrak.

The rocks of the Bamble area are divided into two broad groups (Bugge, 1943), the older group and the younger group. The older group is composed of diverse rocks, of which a number are of undoubted supracrustal origin, while the younger group arose from the older group and consists of migmatites, charnockites, pegmatites, and granites.

The supracrustal origin of many rocks from the older group can still be recognized. Near Arendal, layers of marble are intercalated with quartzite, amphibolite, and granite gneiss. A series of mica schists that contain sillimanite, cordierite, and corundum extends from the Oslo area southwest to Tvedestrand and is a probable derivative of argillaceous sedimentary rocks. Quartzites and arkosic quartzites are common and widely distributed as is banded gneiss, which is composed of alternating light and dark bands from several centimenters to a few meters in thickness. The individual bands consist of amphibolitic, biotitic, quartzitic, dioritic, and granitic gneiss. The banded gneiss may, in places, include layers of marble, mica schist, and quartzite. The bands must have been determined by original bedding in many cases although the chemical composition of the bands can have changed. Dietrich (1959) has proposed a sedimentary origin for the banded gneisses around Randesund in the southern part of the area. Although any particular rock unit may have been so altered that its original features are no longer recognizable, the older group of the Bamble area is most likely a supracrustal succession of rocks.

Numerous small bodies of hyperite are found in the Bamble area. They seem to have been intruded during a second orogenic phase but are earlier than the granitization. The hyperites form concordant bodies whose intrusive nature is suggested by plutonic texture, fine-grained border facies and transgressive features in places. Many of these bodies have been subjected to metasomatic alteration that resulted in the formation of scapolite, phlogopite, and apatite. The central parts of the hyperites are usually fresh and massive and pass by gradual transition into schistose amphibolite borders. The hyperites are emplaced in a succession of quartzites, mica schists, and nodular granites of the older group and occur scattered throughout the entire Bamble area. They are always intruded parallel to the strike of the gneisses and may occur as phacolithic intrusions. Subsequent to emplacement of the hyperites, the whole sequence was plastically deformed while the hyperites behaved rigidly. The hyperites represent synkinematic intrusions in an orogenic belt.

Some rocks of uncertain age relation occur. These are the albitites, the cordierite-anthophyllite gneisses, and the nodular granites. The albitites are often tourmaline or rutile bearing and are associated with carbonate dikes containing albite and calcite. The cordierite-anthophyllite gneisses are closely associated with amphibolite and occur commonly in zones of sillimanite gneiss and quartzite. They vary from melanocratic to leucocratic. Nodular granites occur in the northeastern part of the area and seem to be genetically connected to sillimanite gneisses and normal granites. They never show intrusive contacts but rather occur as alternating layers in the surrounding gneisses. The occurrence of sillimanite in these rocks is limited to the lenticular nodules, which may give the rock a lineation. The chemical compositon (Bugge, 1943, p. 107) suggests a possible sedimentary origin.

The rocks of the older group are scattered all over the area as lenses, inclusions, schlieren, and digested remnants in the migmatites and gneisses of the younger group. In spite of alteration by tectonism and metamorphism, the vestiges of a supracrustal origin remain.

The younger group of migmatites and granites reworked the supracrustal rocks during a second orogenic period. Processes of ultrametamorphism transformed the rocks of the older group into migmatites while synkinematic granites were emplaced.

Part of these reworked rocks are the arendalites, a group of migmatitic rocks of charnockitic kindred that occur along the coast near the town of Arendal. Three divisions in the arendalites are recognized: (1) The basic division with 48—55 % SiO_2 varying from grabbros and norites to hornblende norites and amphibolites. (2) The intermediate division with 55—65 % SiO_2 containing quartz norites, quartz-hypersthene diorites, and mangeritic to monzonitic rocks. (3) The acid division with >65 % SiO_2 consisting of quartz-hypersthene diorites, charnockites, and biotite or hornblende granites. Rocks of the intermediate division contain antiperthitic plagioclase that is calcic oligoclase and andesine while the rocks of the acid divsion contain mesoperthitic feldspar. The arendalites are placed in the granulite facies. In the northeast, the rocks exhibit a magmatic character but most commonly are migmatitic. The contact between the arendalites and the surrounding rocks is gradational so that Bugge (1943, p. 63) has delineated a map unit called border migmatites which marks the transition zone. Relics of the older rock group occur within the arendalites; bands of amphibolite and dioritic and quartzitic gneiss alternate with typical charnockite. A number of hyperite dikes gives important evidence of the origin of arendalites. The dikes cut the banded gneisses and are clearly younger than these. The dikes also cut the arendalites and seem to be younger; however, the arendalites send numerous apophyses into the dikes and, in the eastern part of the district, exhibit intrusive relations toward the dikes. A metasomatic origin is ascribed to the arendalites because of their transitional nature with the older rocks, their mobilization, and their migmitization.

Synkinematic granites and granite gneiss occur commonly in the Bamble area. The granites are almost always concordant with the surrounding rocks and may either exhibit sharp contacts or contacts that are gradational over a distance of over 100 m. by the appearance of feldspar porphyroblasts within the surrounding gneiss and schist. The granites show all gradations from fine-grained equigranular to porphyroblastic with feldspars up to 10 cm. Some granitic augen gneisses occur along the breccia zone that separates the Bamble and Telemark areas; these granites show marked mechanical effects. The synkinematic granites are attributed to palingenic or metasomatic processes.

The numerous pegmatites of the Bamble area comprise two different generations. Small concordant pegmatites that contain quartz, feldspar, and biotite but lack rare minerals were formed during the orogeny. Large, lenticular, crosscutting pegmatites are mainly composed of quartz and microcline but may contain abundant rare minerals. Since these pegmatites are massive and undeformed, they must be either late- or postkinematic. In addition to the above pegmatites, small, pillow-shaped plagioclase-rich pegmatites are scattered throughout the mafic rocks and were probably formed by secretion.

The postkinematic granites of the Bamble area form the principal objects of study of this investigation. These are the Herefoss and Grimstad granites, which crop out in the south-central part of the area and are located about 10 km. apart. The Herefoss and Grimstad granites are approximately circular in plan and have diameters of 20 km. and 8 km. respectively. Although the proportions of the minerals vary considerably, several chemical analyses of the two granites are quite similar, and the two granites resemble each other greatly in both megascopic appearance and texture. The chief minerals of the granites are microcline perthite, plagioclase (*ca.* An_{10}), quartz, and biotite. The granites are typified by subhedral megacrysts of micocline perthite but fine-grained facies are also found. The Herefoss granite is strongly foliated in places while foliation in the Grimstad granite is elusive at best. The foliation of the surrounding gneisses wraps around the Herefoss granite so that this granite presents pseudoconformable contact relations; the Grimstad granite is conformable on the southwest and northeast where the contact runs parallel to the regional strike and is transgressive on the northwest. Breccias occur at the contacts of both granites.

The rocks of the Telemark and Bamble areas are separated by a fault zone called the "great friction breccia", which has been studied by A. Bugge (1928) and Selmer- Olsen (1950). The breccia forms a zone of varying width that is usually followed by a steep-walled valley. The breccia cuts through the Herefoss granite, and, in its vicinity, the southeast side (Bamble rocks) has moved downward and northward with respect to the northwest side (Telemark rocks). Northeast of

Herefoss, the displacement is distributed over a series of faults to give a step-like profile. The breccia is marked by mylonites and sheared quartz veins, which indicate that the movement was recurrent. The vertical displacement of the fault is undecipherable from the stratigraphy, but a gravity study could give the approximate minimum value.

Telemark Area

The Telemark rocks lie to the northwest of the "great friction breccia" and have long been distinguished from the Bamble rocks. More recent opinion presupposes that the Telemark and Bamble rocks are nearly contemporaneous but have undergone somewhat different histories.

The Telemark rocks consist of a sequence of metamorphosed but well preserved (crossbedding, ripple marks, *etc.)* supracrustal rocks lying on a gneissic basement. The supracrustal rocks do not occur in the area investigated; the present study is concerned only with the gneissic basement. The rocks of the Telemark basement are well studied only in the southwestern part of the area (Barth, 1929, 1947a) and are virtually unknown further northeast. Granites and granite gneiss predominate; in these, bodies of amphibolites and rocks of probable supracrustal origin are swimming. The area indicates a process of homogenization in which initial compositional differences have been evened out; metasomatism, which has locally resulted in anatexis, is believed to be the process that has effected the present condition of this monotonous gneissic terrain. A glance at the geologic map (Plate 2) quickly reveals the striking difference between the Telemark and Bamble areas. Within the Bamble area numerous separate rock units have been mapped with sharp contacts, but, just northwest of the breccia zone, a repetitious succession of gneiss and granite with indistinct borders occurs. Another point of interest is the fact that, in spite of the granitic nature of the rocks, no large homogeneous granites have been recognized.

The largest distinctive body mapped in the area is the Evje amphibolite. The Evje amphibolite extends 30 km. in a N-S direction and ranges from 2—10 km. in width. The amphibolite is nickeliferous, and, since relics of an original igneous hypersthene occur rarely, the amphibolite is most likely the transformed derivative of a norite. It is probably related to the marginally metamorphosed norites of the

Bamble area, but, at Evje, the metamorphism took place at a deeper level so that the original igneous body was completely amphibolitized. The contacts of the amphibolite are transitional with the surrounding gneisses and indicate a progressive granitization (Barth, 1947a).

Large lenticular pegmatite dikes are unusually common in the amphibolite while the surrounding gneisses contain none whatsoever. The quartz and feldspar crystals in the pegmatites attain lengths of eight meters; the pegmatites are famed for their variety of rare minerals. No feeder channels have ever been found even though a large number of pegmatites have been quarried. An origin by secretion of ions into potential cracks formed in the more rigid amphibolite has been evoked (Barth, 1947a).

The Oddersjå granite is one of the few bodies that has been studied immediately west of the breccia zone (Barth and Bugge, 1960). This synkinematic granite is an elongate conformable body of augen gneiss, which exhibits N-S trending lineation and foliation. The contacts of the granite are gradational, either through a gradual transition or through a migmatite zone. This granite is regarded as the product of granitization.

Structure

The structure of the area investigated has not been studied extensively, but, as is the usual case with areas of the crystalline basement, the structure can be safely assumed to be complex. The studies that have been made confirm this supposition. Barth (1947a) recognized that the Evje amphibolite and the surrounding gneisses have undergone two periods of folding, an older one around NW-SE axes and a younger one around N-S axes. Wegmann (1960) distinguishes three periods of deformation. The oldest deformation occurred at a fairly shallow level and was accompanied by metamorphism and the emplacement of ophiolitic and granodioritic rocks. During the second deformation, which involved deeper levels within the migmatite zone, the first folds were again folded and warped so that structures of great complexity evolved. Widespread granitization accompanied this deformation and gave rise to different granites that range from augen gneisses to massive plutons. The mobilizing effect of this granitization on the substratum resulted in intrusive movements. In the Risør-Arendal-Grimstad region, the first folding around N-S axes was followed by the second folding around NE-SW axes. The substratum of the supracrustal series seems to have been mobilized everywhere so that the original basement is nowhere preserved. The Telemark supracrustal rocks may be younger than the Bamble rocks; however, since the Telemark supracrustal rocks are deformed around domes of Telemark granite in the Tørdal-Drangedal district, they are older than the second deformation. The structures of this district of Telemark reflect deformation at an upper level while the characteristics of a lower level are exposed in the tract between the coast and the inner districts of Telemark. The movements of the third period of deformation took place in a rigid basement at a rather high level and resulted in the numerous fault zones, marked by mylonite, that dissected the area into a series of rhombohedral slices. Movements along these zones probably recurred up to the Permian.

The previous description gives a general survey of the geology of the area studied. Because of the complementary nature of known geology and gravity interpretation, a detailed discussion of each geologic feature will be given in conjunction with the interpretation of its gravity field.

The gravity maps.

Introduction

Although the Bouguer gravity map often reveals anomalies which can be attributed to geological features, the true shape and amount of the local anomalies are often masked and distorted by the regional anomalies. A residual anomaly map was, therefore, constructed for the region including the Herefoss and Grimstad granites. In order to do this, a net of NE-SW and NW-SE profiles of the Bouguer gravity anomalies was drawn in the vicinity of the two granites. Smoothed curves were then drawn through the Bouguer gravity profiles, and the smoothed curves were mutually adjusted until they were in agreement at their intersections. Thus the smoothed curves represent the regional gravity profile, and the difference between the values on the smoothed curve and the Bouguer gravity profile at any particular point becomes the residual gravity anomaly. The smoothed profiles are used to construct a map which is the regional gravity map (Fig. 2). If this map is subtracted from the Bouguer gravity map and the values obtained are contoured, a residual anomaly map results (Fig. 3). The Bouguer anomaly map is simply the sum of the regional anomaly map and the

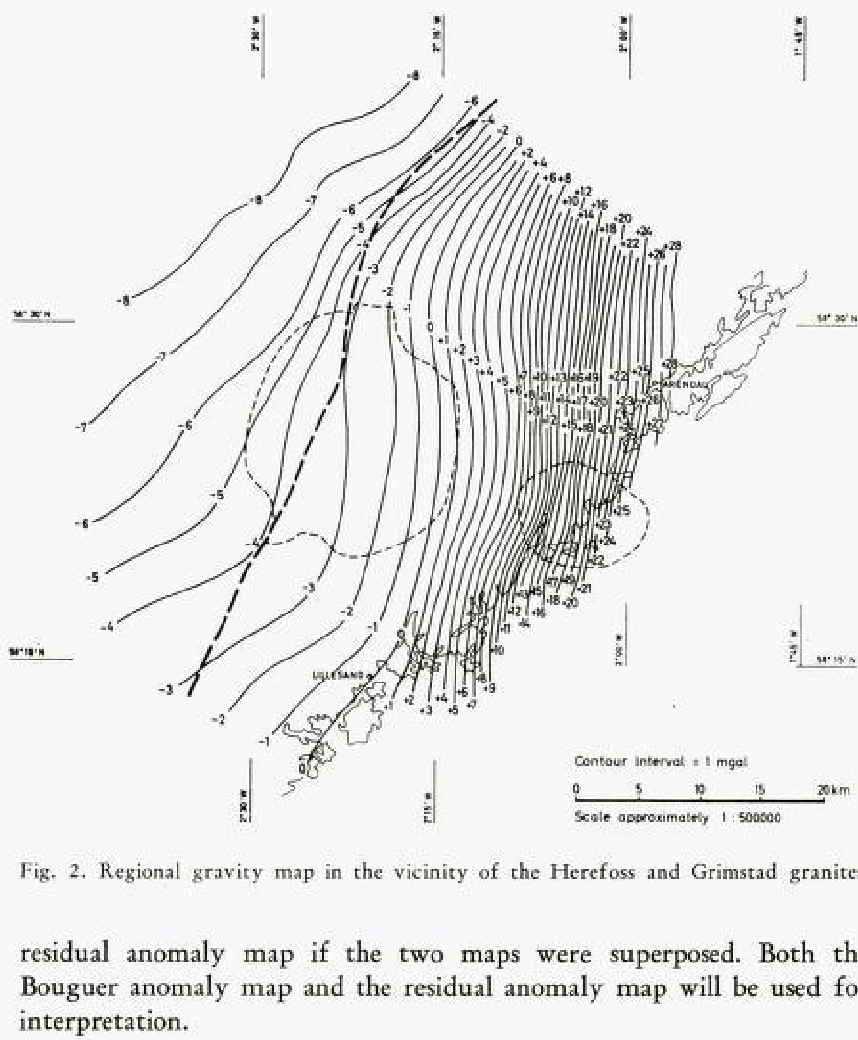


Fig. 2. Regional gravity map in the vicinity of the Herefoss and Grimstad granites.

residual anomaly map if the two maps were superposed. Both the Bouguer anomaly map and the residual anomaly map will be used for interpretation.

Bouguer Anomaly Map

The Bouguer anomaly map (Plate 1) reveals a series of anomalies of considerable interest. In viewing the anomalies, however, the reader must pay particular attention to the distribution of stations upon which the anomaly in question is based. Although the original intention was to treat only the Herefoss and Grimstad granites, the map was extended to include a much larger area as other anomalies were revealed.

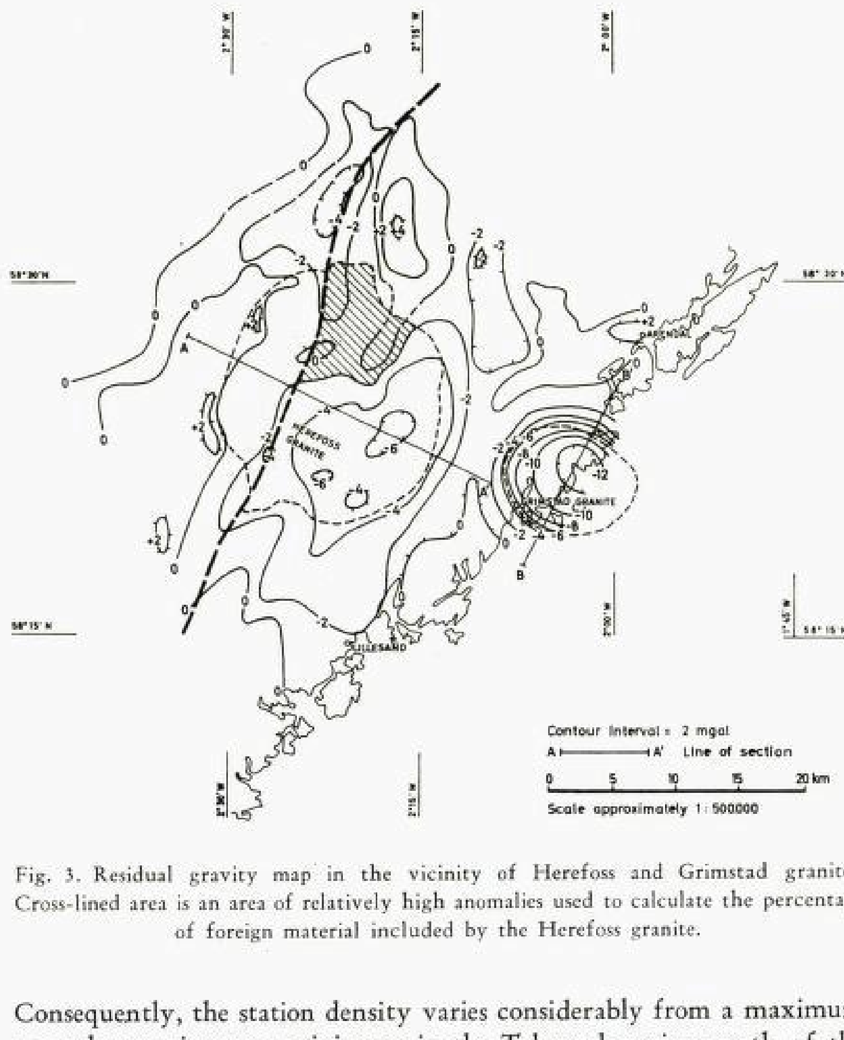


Fig. 3. Residual gravity map in the vicinity of Herefoss and Grimstad granites. Cross-lined area is an area of relatively high anomalies used to calculate the percentage of foreign material included by the Herefoss granite.

Consequently, the station density varies considerably from a maximum near the granites to a minimum in the Telemark region north of the breccia zone. Any evaluation of the anomalies should take into consideration the reliability of the isoanomaly lines themselves.

The most striking feature on the Bouguer gravity map is the strong gradient in the gravity field that decreases from + 50 mgal along the coast on the northeast to about 0 mgal along the breccia zone. The + 50 mgal line follows the coast from Brevik to Risør and then runs out into the Skagerrak where the approximate positions of the isoanomaly lines at sea are determined from stations measured by Collette

(1960) in a submarine (off the coast the lines are drawn on a 10-mgal interval). The isoanomaly lines can be seen to run almost parallel to the coast and to the breccia zone. It is likely that with a dense station net innumerable curves due to local anomalies would be superimposed on the contours; however, their present straight, parallel course certainly reflects the regional trend. Between the Herefoss granite and the vicinity of the Oslo region, the location of the breccia zone is marked quite closely by the 0-mgal line. The effect of the Herefoss granite causes the 0-mgal line to swing out southeastward, and, to the north, the effect of the Oslo igneous province produces a northward bend in this isoanomaly line. The + 50-mgal line would seem to continue from the coast near Risør out to sea so that a positive anomaly of 50 mgal is still present in the Skagerrak *ca.* 34 km. south of Kristiansand. The gravity gradient over the Bamble area is about 2mgal/km. and decreases northwestward; the gravity gradient over the Telemark area just northwest of the breccia zone is less than 1 mgal/km. and decreases northwestward. The decrease in the Bouguer anomaly toward the center of the country and the Caledonides is due, partially at least, to isostatic compensation of the mountain range. It may be also due to the still unequalized sinking of the land under the weight of the Pleistocene ice sheet. That both the gravity gradient and gravity anomaly itself over the Bamble area exceed normal values can hardly be disputed. A regional anomaly which will be called the Bamble anomaly exists, therefore, over the Bamble rocks between the border of the Oslo province and area offshore from Kristiansand.

The isoanomaly lines swing northward before they enter the Oslo province just north of Skien. A recently published Bouguer anomaly map (Norges geografiske oppmåling, 1960) shows that large positive anomalies are generally found over the Oslo igneous province. The northward swing of the isoanomaly lines indicates that a mass excess is causing the anomaly. No faults have been mapped along this border, and a ready explanation for this deflection cannot be surmised from the surface geology. One can, however, state with certainty that this deflection is closely related to the contact between the Precambrian basement and the Oslo igneous province and that a mass excess in this region is the cause.

Superposed on the strong gravity gradient over the Bamble rocks are two gravity "lows" that correspond to the outcrops of the Herefoss and Grimstad granites. The "low" over the Grimstad granite is marked by a regular pattern that corresponds closely to the granite outcrop. A slight northwestward displacement and distortion is effected by the large gravity gradient, while the actual amount of the anomaly remains concealed. The Herefoss granite, on the other hand, is the site of a gravity "low" whose outline and location do not conform very well to the outcrop of the granite. Small gravity "highs" occur in the granite, and "lows" are found outside it on the south. The true form and amount of the negative gravity anomalies over the Herefoss and Grimstad granites are camouflaged by the regional gravity gradient.

The embayment in the + 42-mgal line *ca.* 4 km. south of Kragerø deserves mention. This embayment reflects a gravity "low" in a gravity profile that was measured across the Levang granite. This is a particularly interesting feature because the surface pattern of a doubly folded structure can readily be followed through the granite (T. Elder, oral communication). The Levang granite will be the subject of further geological and geophysical studies.

Although the station density is low in the Telemark rocks, one feature stands out. It is the positive anomaly that occurs over the Evje amphibolite. Even though the exact shape of the anomaly remains in some doubt, the minimum value of the anomaly can scarcely be less than 10 mgal. Naturally a positive anomaly over an amphibolite is not surprising. As far as can be ascertained, the gravity values west of the breccia zone show little variation and reflect the homogeneity of the rocks, but in a little known area north of Evje and in Tørdal irregularities occur in the profiles.

Residual Anomaly Map

The Bouguer anomaly map has been reduced to a regional anomaly map and a residual anomaly map, Figs. 2 and 3 respectively. The regional anomaly shows the regional gravity gradient which decreases westward to be *ca.* 2.4 mgal/km. near the Grimstad granite. After the regional effect is removed, the gravity fields of the two granites stand out more distinctly. In addition, the gravity minimum of the Grimstad granite occurs directly over the granite. The minimum value of the Herefoss granite has been shifted, and the small highs within this granite have changed their shape. The isolation of residual anomalies has resulted in a clearer picture that will facilitate interpretation.

Interpretation of the gravity anomalies.

Grimstad Granite

The Grimstad granite could be described as the ideal granite for gravity interpretation. Accurate elevations are available, the terrain effects are negligible, and both the outcrop pattern and gravity field are simple. The only complication is that much of the granite is covered by either water or farmland.

The geology of the Grimstad granite has been studied by Oftedal (1937, 1945). The typical granite is coarse grained to porphyritic and is principally composed of microcline perthite, plagioclase An_{10}-An_{30}, quartz, and biotite. The large red phenocrysts of microcline perthite give the granite a distinctive appearance. Microcline perthite is the dominant mineral constituent while plagioclase and quartz each make up about 20 % of the rock and biotite 5—10 %. Normative calculations by Elders (1963) show that the Grimstad granite is a quartz monzonite. An aplitic facies of the granite is found, mostly as dikes, in the country rock and at the contacts. Two more mafic varieties also occur as small bodies within the main granite. Both of these are older than the main granite; the one is a hornblende-biotite adamellite and the other is a biotite adamellite. Oftedal considers the origin of the Grimstad granite to have been through either magmatic differentiation or more likely palingenesis.

The general contact relations of the granite can be seen on the detailed geologic map, Fig. 4. The map shows sharply trangressive contacts except on the northwest where the contact parallels the strike of the gneiss. On the north and south sides of the granite, the contact cuts sharply across the regional strike of the gneisses. At one place on the north side, however, the strike of the gneiss runs parallel to the contact of the granite and swings back to the regional trend a few hundred meters north of the contact (Elders, 1961). The Grimstad granite is more discordant than the Herefoss granite. During emplacement, the Grimstad granite seems to have cut across the gneisses without appreciably deforming them.

The gneiss at the contact is interwoven with aplitic dikes in many places so that the contact is difficult to place exactly. These dikes are found up to several hundred meters outside the contact. Toward the gneiss, the fragments seem to be in place, and, toward the granite, they assume diverse attitudes. The gneiss inclusions are angular but exhibit various stages of assimilation. Amphibolite inclusions are fresh and have sharp borders; inclusions of mica schist and gneiss are either partly assimilated or may only be gray shadows in the granite. Where gneiss fragments are uncommon, the contact between the granite and surrounding gneiss is razor sharp. The texture of the granite remains the same right up to the contact, and thermal metamorphic effects are completely lacking. The contacts themselves are either vertical or dip very steeply outward. The emplacement of the granite gave rise to an eruptive breccia, but has not caused thermal metamorphism.

The contact effects can be better described as small scale migmatitic rather than thermal metamorphic. Felsic material penetrates intimately along the foliation planes so that the adjacent gneiss assumes the red color of the granite. This migmatization is strongest in well foliated mica schists and is missing in the massive amphiboltes. It is also strongest where the granite cuts directly across the foliation of the gneiss and is only of local extent where the contact parallels the foliation of the gneiss.

Flow structure has not been reported in the Grimstad granite. The writer has noticed a faint preferred orientation of microcline phenocrysts on favorable exposures, but it appeared to be highly variable. Possibly, detailed investigation of good exposures would reveal a coherent pattern in the flow structure.

The trace-element distribution in a small number of the microcline perthites from the granite has been analyzed by Heier and Taylor (1959, p. 291). The study indicated an increasing differentiation toward the center of the granite. The microcline from an aplite dike at the border of the granite has a K/Rb ratio similar to the center of the granite and is probably a late differentiate.

The Grimstad granite can be regarded as an almost circular, homogeneous granite that has near vertical margins which are marked by an eruptive breccia and migmatite zone without thermal metamorphism. It is a granite diapir that has not appreciably deformed the surrounding mica schists, banded gneisses, and amphibolites of the Bamble area.

The residual anomaly map (Fig. 3) shows a semicircular negative anomaly of 13 mgal lying over the Grimstad granite. The isoanomaly lines follow the contact of the granite and are displaced slightly to the northeast. The density determinations show that there can be no doubt that this negative anomaly is directly due to the mass deficiency of the granite.

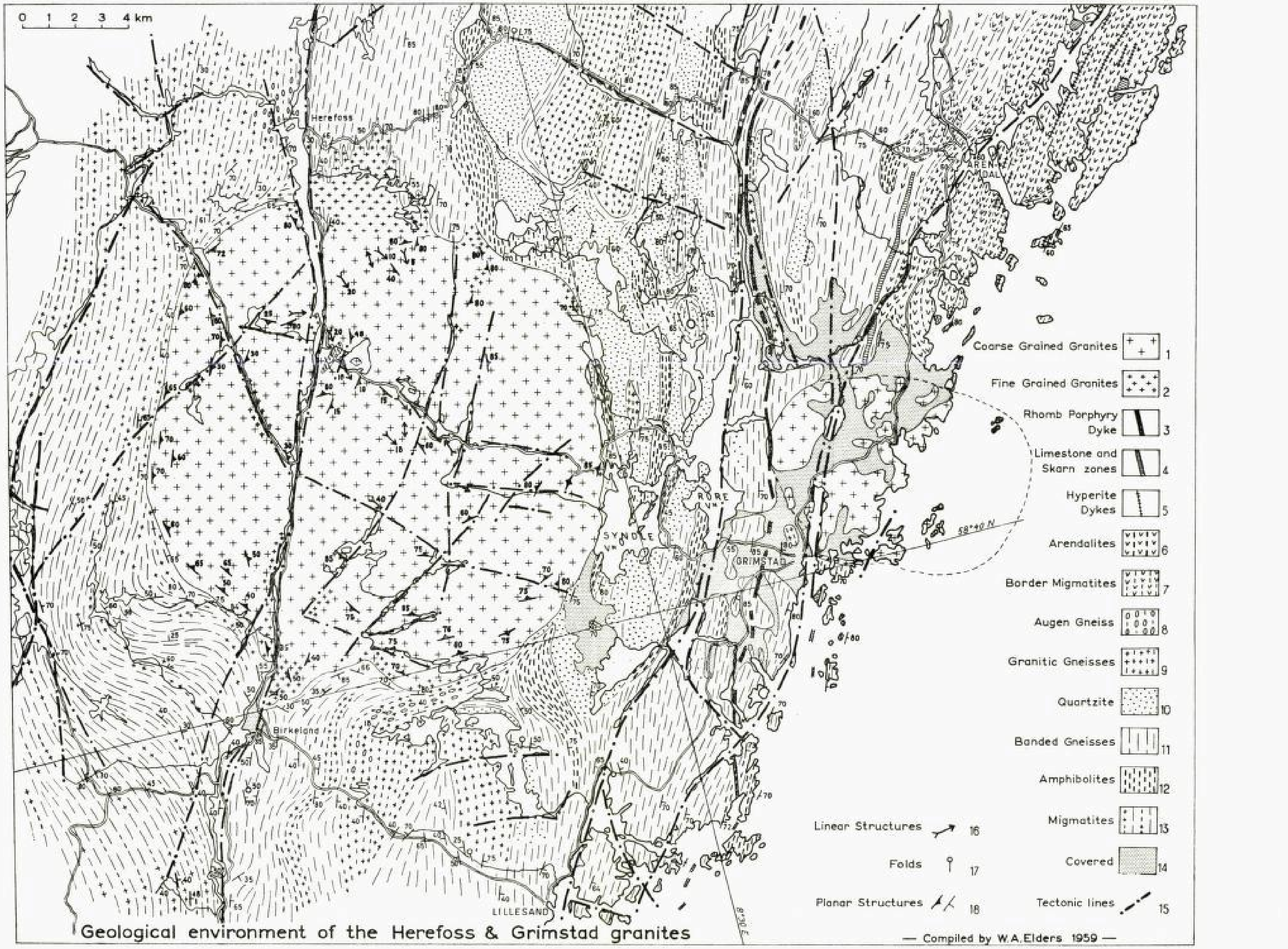

0 1 2 3 4 km
Herefoss
Birkeland
LILLESAND
SYNDLE
RORE
GRIMSTAD
AREN-DAL
58°40 N
8°30 E
Coarse Grained Granites 1
Fine Grained Granites 2
Rhomb Porphyry Dyke 3
Limestone and Skarn zones 4
Hyperite Dykes 5
Arendalites 6
Border Migmatites 7
Augen Gneiss 8
Granitic Gneisses 9
Quartzite 10
Banded Gneisses 11
Amphibolites 12
Migmatites 13
Covered 14
Tectonic lines 15
Linear Structures 16
Folds 17
Planar Structures 18
Geological environment of the Herefoss & Grimstad granites
— Compiled by W.A.Elders 1959 —

Figure 4.

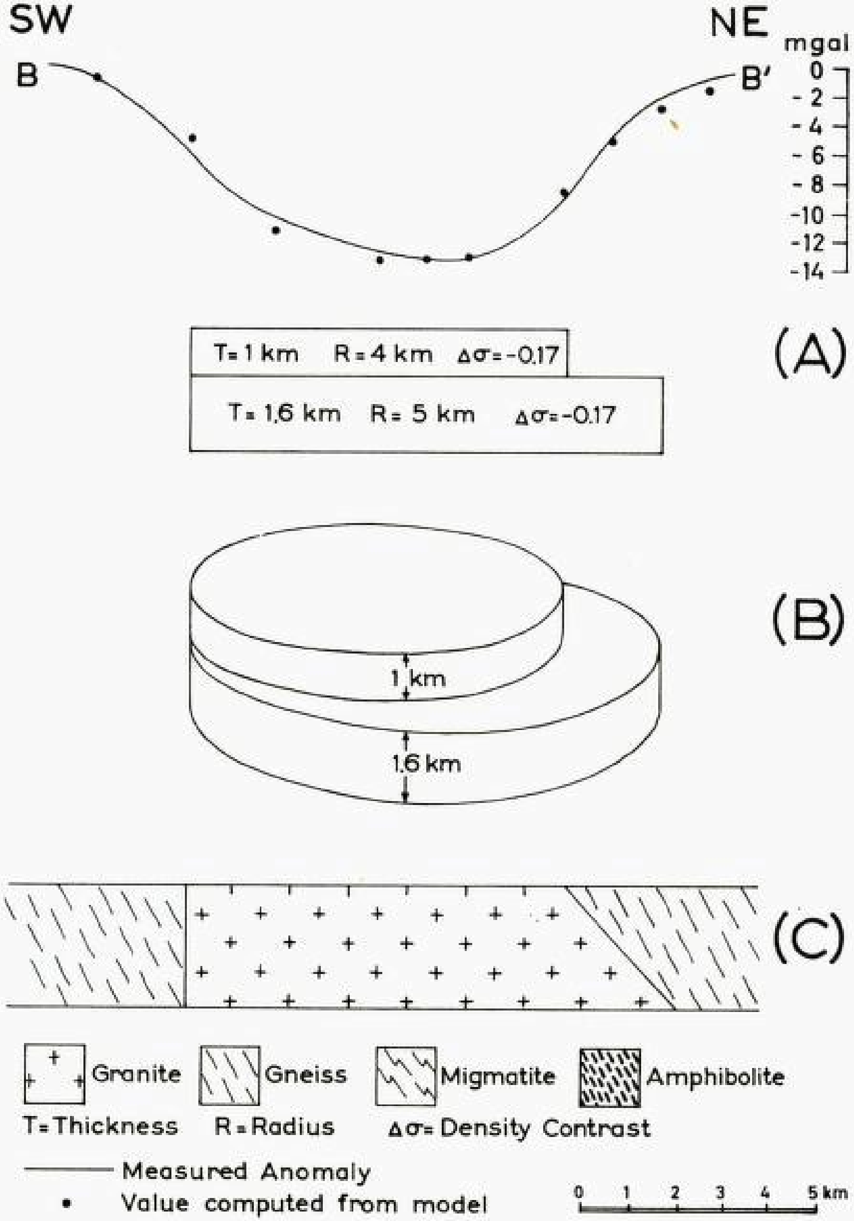


Fig. 5. (A) Profile of the gravity field from the residual anomaly map and the model calculated to simulate the gravity effect of the Grimstad granite. (B) 3-dimensional view of the gravity model. (C) Geologic interpretation of the model. Model based on uniform density contrast of —0.17 gm/cm^3.

The measured rock densities (Table 2) reveal that a large density difference is found between the gneiss and the granite. For 20 samples of Grimstad granite and 54 samples of the Bamble rocks, the mean densities are 2.64 gm/cm^3 and 2.81 mg/cm^3 respectively, giving a density difference of —0.17 gm/cm^3.

Use is made of this density difference to calculate a probable model that will simulate the effect of the granite. A model of a vertical cylinder suggests itself immediately because of the contact attitudes and the plan view of the granite. The solid-angle method of Nettleton (1942, p. 304) was used to compute the two following models that are proposed for the Grimstad granite.

The first hypothetical model presupposes that the granite can be represented by two superposed vertical cylinders each of which have a mass deficiency of 0.17 mg/cm^3. The simplest model would be a single vertical cylinder with the same diameter as the granite; however, the asymmetrical shape of the gravity field precludes this possibility. Two superposed cylinders of unlike diameters were employed, therefore. The upper cylinder has an 8-km diameter, the shorter diameter of the surface exposure, and a thickness of 1 km; the lower cylinder has a 10-km diameter and a thickness of 1.6 km. This model and a profile through the gravity field appear in Fig. 5. The agreement between the calculated attraction and the measured attraction is good; only two points fall slightly off the measured profile. As a consequence of these two cylinders of unequal radius, the northeastern contact dips moderately outward at about 47°, a value that is much less than geological observations would indicate, because only steeply dipping or vertical contacts have been noted.

The second hypothetical model (Fig. 6)is based on the premise that the density difference decreases at depth; i.e., the country rocks become more granitic (migmatitic) or the granite becomes denser. If the Bamble gneisses give way to more granitic rocks at depth, the density contrast between gneiss and granite would decrease. Density differences of 0.17 mg/cm^3 and 0.10 gm/cm^3 (arbitrary) are, therefore, used for the upper and lower cylinders respectively. These cylinders have diameters of 8 and 10.5 km, but the thickness of the lower cylinder is 3 km. The computed profile can be seen to agree almost as well with the measured profile as the previous solution. This model indicates that the northeastern contact dips outward at about 50°, a value that does not agree much better with the observed geology.

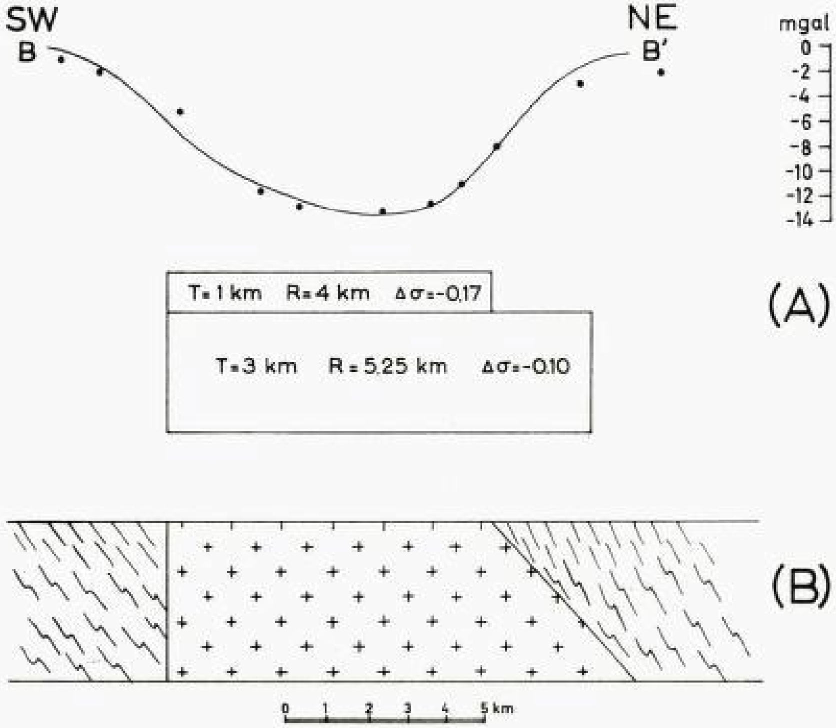


Fig. 6. (A) Profile of the gravity field from the residual anomaly map and the model calculated to simulate the gravity effect of the Grimstad granite. (B) Geologic interpretation of the model. Density contrast decreases at depth; 0.17 gm/cm^3 used for the upper cylinder and 0.10 gm/cm^3 used for the lower cylinder.

The conclusion is that the northeast contact dips under the gneiss at a lower angle than the geological observations indicate. Because the Grimstad granite is associated with small terrain effects and accurate elevations, the asymmetry of the gravity field is real and reflects a contact of moderately high dip. The suggested form of the granite is something between models one and two (Figs. 5 and 6). Superposed vertical cylinders with diameters of 8 km and 10 to 10.5 km closely simulate the Grimstad granite. Although the anomaly of the second model appears to be a little too broad, outside the granite, the choice between the two solutions is about equal if the accuracy of the methods is considered. The thickness of the granite may be from 2.6 to 4 km.

The "thickness" of the granite requires some explanation. The gravity data can only determine the vertical extent or depth to which a density contrast extends; *i.e.*, in the case of a granite, no distinction can be made as to whether the granites "swim" in gneiss or project from a granite layer upward through the overlying gneiss. This point should be remembered throughout this paper. *The word "thickness" applied to a gravity model means the vertical extent of the density contrast.*

Herefoss Granite

The Herefoss granite could, with the exception of rather large terrain effects and locally difficult accessibility, be described as another ideal granite for a gravity study. This granite has recently been the subject of intensive structural studies (Elders, 1961, 1963) and petrographic studies (Nilssen, 1961). The present gravity investigation is intended to complement the above studies.

The Herefoss granite can be conveniently divided into a western and eastern part because the granite is cut by the "great friction breccia". Movement along the breccia was such that the western part represents a deeper section than the eastern part and causes real differences in the rocks on the two sides. The layered supracrustal Bamble rocks occur on the east side and the migmatitic Telemark granitic gneisses occur on the west side. Inclusions of country rocks are common in the granite on the east and rare on the west; the petrography of the granite itself is different on the two sides of the breccia.

As can be seen from the detailed geologic map (Fig. 4), the Herefoss granite is almost circular in plan and has a diameter of about 20 km. In contrast to the Grimstad granite, the contacts of the Herefoss granite are largely concordant in plan. On the southeast, southwest, and northwest sides, the gneisses are deflected from their normal northeasterly strike and swing around the granite. On the north, the contact is partly migmatitic and interfingering and partly crosscutting. Because the granite transects the gneisses in a vertical section, the map view of the granite gives the false impression that the granite is less transgressive than it is in reality.

The Herefoss granite is composed of two main facies, porphyritic and fine-grained granite. The most common facies is a red porphyritic granite that closely resembles the Grimstad granite, particularly in the southeast quadrant. The large red phenocrysts of microcline give the rock its distinctive appearance. The rock is composed of microcline, plagioclase An_{10-20}, quartz, and biotite. The granite west of the breccia contains the same mineral assemblage but is considerably richer in CaO so that three times as much anorthite appears in the norm. In addition to the usual sphene and opaque minerals, the western granite contains about 1 % hornblende. The mineralogy of the fine-grained facies is similar to the porhyritic varieties. This rock is, however, richer in silica.

The fine-grained granite occurs in small bodies that often appear to be younger than the porphyritic granite. The fine-grained granite exhibits both sharp and gradational contacts toward the porhyritic granite. The largest exposure of fine grained granite occurs in the migmatitic northern tongue of the Herefoss granite.

A gray variant of the prophyritic granite, which is richer in mafic minerals and forms small patches in the northeast quadrant of the granite, is considered to derive from digested inclusions.

Some interesting observations regarding the petrography of the granite can be made from the density data. The results of the density determinations are listed separately for the eastern and western sides of the granite (Table 2). The eastern granite possesses a lower mean density (2.68) than the western granite (2.70) but a larger standard deviation (0.054) and range (2.58—2.80) than the western granite (0.033 and 2.64—2.78 respectively). These values express the fact that the eastern granite is less dense than the western granite; however, the western granite is more homogeneous than the eastern granite.

The contact relations of the Herefoss granite have been described by Elders (1961, 1963). On the east side of the granite, the gneiss dips 70°—80° eastward, and the granite contact dips 80°—85° eastward so that the granite is gradually transgressive. Apophyses from the granite cut sharply across the foliation of the gneiss. The strike of the gneiss, however, parallels the strike of the granite with only local discordances. Here and there, an eruptive breccia was formed; the inclusions are locally derived but seem to have moved relative to each other.

In places, the south contact is gradational over a distance of 2 km as the granite passes into migmatite and finally gneiss. To the east, a sharp contact is marked by an agmatite. The gneiss is parallel to the contact, and many gneiss fragments in the agmatite have their long dimension parallel to the contact. Here the contact dips under the granite at 70°.

Again on the west side, the contact parallels the strike of the gneiss and dips at 30°—40° under the granite. The migmatites of the Telemark formation west of the granite dip more steeply so that the granite appears to transect the migmatites in vertical section. The lack of inclusions along the western contact is notable.

West of the breccia, the north contact of the granite and the surrounding gneisses both strike E-W. The contact dips from 90° to 80° under the granite. Some inclusions here are probably rotated. The granite is transitional into the gneiss over 200—300 m.

The northern contact east of the breccia is the most unusual one because fine-grained granite interfingers with the surrounding banded gneiss and forms a large scale migmatite. Just east of the breccia, the gneiss strikes E-W but changes abruptly to N-S where the fine-grained protuberance of granite projects into the gneiss. Long tongues of granite extend into the banded gneiss with a lit-par-lit relationship. Some veins of fine-grained granite cut across the foliation of the gneiss. Most inclusions of gneiss do not appear to be rotated but some are disoriented. The emplacement of the offshoot from the main granite seems to have been replacement caused by late emanations from the granite rather than by forceful intrusion. The fine-grained granite of this protuberance either exhibits a sharp interdigitation with the main porphyritic granite or grades into it.

Statistically constructed β_1 and β_2 fold axes from the gneiss around the granite give interesting results (Elders, 1961). On the southwest side, the axes of a syncline and anticline strike northeast and plunge under the granite at 52° and 30° respectively. South of the granite, a fold axis strikes E-W parallel to the granite contact and plunges 20° west. On the southeast side, two β axes plunge southeast at 65°—75°. On the northeast side, a β axis plunges 65° east. The regional axial trends are NE-SW and the plunge varies. These lineations either plunge subparallel to the dip of the granite contact or strike parallel to the contact. Although the tectonics of the area is complex, these steeply plunging lineations were most likely influenced by the emplacement of the granite. As in the case of the Grimstad granite, thermal metamorphic effects are missing at the border of the Herefoss granite; however, either border migmatites or agmatites are common.

Inclusions are commonly found scattered throughout the eastern part of the granite but are rarer in the western part. The inclusions in the western part usually occur as partially digested patches of gray granite.

Inclusions on the eastern side usually occur with sharp outlines, but the larger ones may have a sharp border in one place and a gradational border in another place. One inclusion was large enough to appear on the geologic map of Elders (Fig. 4) and contains several large pegmatites. Diffuse patches of gray fine-to coarse-grained granite up to 1 km across are believed to be almost completely assimilated inclusions. One such area that holds great interest because of its coincidence with a gravity high within the granite will be mentioned further. Near the northeast border of the granite at Birkedal, a patch of gray granite which grades into the normal eastern facies lies along the strike of an amphibolite layer outside the granite.

The Herefoss granite contains a distinct flow structure, either a foliation, a lineation, or both. The flow structure is formed by the alignment of microcline phenocrysts and may be very strong. The flow structure seems to be more common near the border of the granite. The strike of the foliation generally conforms to the strike of the adjacent granite contact but dips more steeply toward the center of the granite.

Structural evidence indicates that the Herefoss granite is an intrusive body, a granite diapir. The granite has brecciated contacts with rotated inclusions, but migmatization and replacement have played important roles locally so that the contact may be either knife-sharp or gradational. Inclusions that may be large (1.5 km) are scattered throughout the granite, particularly on the east side, and assimilation may have occurred on a rather large scale. Due to displacement along the breccia, two different levels of the granite are exposed. The granite west of the breccia is more calcic, more dense, and more homogeneous than that east of the breccia. Contact attitudes and flow structure suggest that the granite has a funnel-like shape steeply inclined to the southeast.

The residual anomalies over the Herefoss granite are strikingly different from those over the Grimstad granite (Fig. 3). The maximum anomaly is smaller, numerous irregularities appear in the gravity field, and the isoanomaly lines do not parallel the granite contact everywhere. The gravity anomalies suggest that the Herefoss granite seems to have undergone a somewhat different evolution than the Grimstad granite.

That the granite is poorly defined by the isoanomaly lines can be best explained by the petrography of the rocks themselves. The density difference between the granite and surrounding rocks is less than that of the Grimstad granite. The Herefoss granite itself is decidedly more dense, and its enveloping rocks are highly variable. In contrast to the rather uniform banded gneisses and biotite schists that surround the Grimstad granite, the Herefoss granite is encased by quartzites, amphibolites, banded gneisses, augen gneisses (granitic), migmatites, and granite gneiss. The insignificant anomaly that occurs along the western border is due to the low density of the granite gneiss and migmatite that borders the granite here. An elongate positive anomaly of 2 mgal just outside the contact of the granite on the west side occurs over an outcrop of amphibolite that was noted by the writer during the field work. The broad negative anomaly of 2 mgal between the Herefoss granite and Lillesand follows the geologic contacts somewhat and partly coincides with outcrops of augen gneiss. The geologic map, however, shows banded gneisses in the area so that other, unexposed causes may have to be invoked in explanation. In the area just west of Lillesand, Dietrich (1959) has mapped innumerable pegmatites. It may be that granitic rock underlies this area at a shallow depth. The positive anomaly north of the Herefoss probably results from the positive effect of the banded gneisses with which it coincides.

The zero-to-small negative anomalies in the north-central part of the granite confirm gravimetrically what the geologist might intimate from the field relations. The area contains one mappable inclusion, numerous smaller ones, and patches of gray granite that were probably formed by the digestion of inclusions. The small anomalies here indicate that the average density of this part of the granite is only slightly less than that of the country rock. The effect of masses of gneiss enclosed within this part of the granite is great enough to give a comparatively large increase in the measurements and to almost neutralize the mass deficiency of the granite in this area.

The anomaly of the Herefoss granite itself shows a maximum value of -7 mgal which occurs in the southeast quadrant of the granite. This indicates that the maximum thickness and/or maximum density difference is found in this area. The model used to simulate the gravity effect of the granite is composed of two superimposed discs, which were suggested by the outcrop pattern and contact attitudes. A model comprising only one disc of uniform thickness would not account for the gravity "low" enclosed by the -6-mgal line. Because of the irregularity of the gravity field, particularly within the northeast quadrant of the granite, a different method was used to caluculate the gravity effect of the model. A circular diagram divided into compartments by concentric circles and radii was constructed. This diagram resembles a terrain correction chart (Hammer, 1939); the diagram is placed on the point in question and the attractions of all the compartments are read from a table and added together. The method has been described by Bean (1953, p. 525) and is readily applicable to a problem such as the Herefoss granite.

The first model that approximates the gravity field of the granite comprises two discs of unlike density. The upper disc has a diameter of 18 km, a thickness of 1 km, and a density contrast of 0.13 gm/cm^3. The density difference of 0.13 gm/cm^3 is the difference between the mean rock density of the Bamble formation (2.81) and that of the east part of the Herefoss granite (2.68). For the lower disc, however, a density contrast of 0.07 gm/cm^3 which is the difference between the mean rock density of the Telemark formation (2.77) and that of the west part of the granite (2.70) was employed. Although the fact that the granite is cut by a fault is a complicating factor both geologically and gravimetrically, it also reveals information that would not normally be available. Because the west side of the fault is upthrown with respect to the east side, a deeper level is exposed. The assumption that the rocks exposed west of the breccia occur at depth east of the breccia is reasonable; *i.e.*, in depth the granite becomes more dense, and the Bamble supracrustal rocks are underlain by the Telemark migmatites and granite gneisses. The density contrast of 0.07 gm/cm^3, therefore, is justified for the lower disc, which has a thickness of 1 km and diameter of 8 km. The density contrast of 0.07 gm/cm^3 is also used for the disc west of the fault because both the density determinations and the gravity anomalies indicate that the density contrast is lower here. The western lower part of the upper disc is indented, and the eastern edge of the upper and of the upper and lower discs coincide so that the model presents contacts similar to those observed in the field.

The profile of the gravity model and the computed points appear in Fig. 7A. As can be seen, the computed points fall on the measured field near the center of the profile and at the eastern contact but differ east of the eastern contact. Points computed for the western side using a density contrast of 0.13 gm/cm^3 fall well off the measured profile (Fig. 7A). Figure 7C shows a profile of the geological interpretation of the gravity model, a faulted granite with a maximum thickness of 2 km, a western border dipping moderately inward, and a vertical

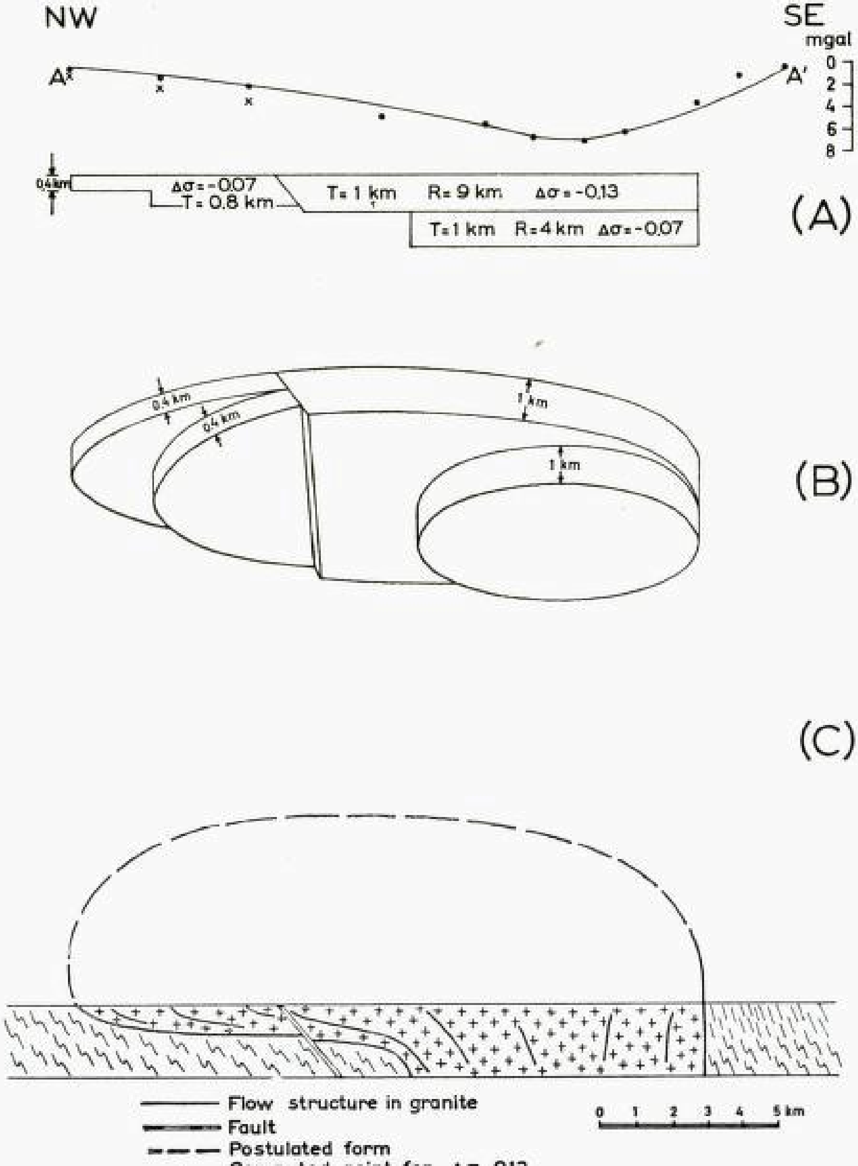


Fig. 7. (A) Profile of the gravity field from residual anomaly map and model calculated to simulate the gravity effect of Herefoss granite. (B) 3-dimensional view of gravity model. (C) Geologic interpretation of model. Smaller density contrast on west side of granite and in lower cylinder. Broken line shows postulated extent of granite before erosion.

eastern border. The model shows a displacement in the granite of 0.2 km at the breccia. Although better agreement with the measured profile is achieved by using a smaller thickness on the western side, the gravity anomalies within the granite are not directly influenced by the faulting; *i.e.*, there is no "step" in the gravity profile within the granite. The value of 0.2 km is not regarded as a useful calculation of the throw along the breccia zone. The effect of faulting is indirectly expressed by the difference in the anomalies over the eastern and western parts of the granite. The maximum thickness of 2 km occurs near the eastern border of the granite; the Bamble rocks are presumably underlain by the Telemark migmatites in this area.

A granite 18 km in diameter and only 2 km thick presents a surprisingly thin cross section (Fig. 7), particularly for one with diapiric characteristics. Let us then calculate the "limiting case" for the maximum thickness that we might possibly expect to occur. Although any one value for the Bouguer anomaly may be in error by as much as 1 mgal, there is no reason to suspect any bias in the Bouguer anomalies that would cause them to be consistently positive. On the contrary, if a bias exists, it is probably a negative bias which is indicated by the consistently small terrain corrections using Hubbert's method in Table 1. The logical value to alter, therefore, is the value of the density contrast. Accordingly, we will assume the density contrast of 0.07 gm/cm^3 that was determined between the Telemark gneisses and the western Herefoss granite exists all around the granite and at depth. The density contrast west of the breccia is arbitrarily taken as 0.04 gm/cm^3, approximately half of the value that was used previously.

Again a model composed of two superposed discs simulates the gravity effect of the granite (Fig. 8). The upper disc is 18 km in diameter and 2 km in thickness except for the western part which has a general thickness of 1.6 km and a 0.8 km indentation along the border. The lower disc has a diameter of 8 km and a thickness of 3 km. The eastern edge of the lower disc is vertically aligned with the eastern edge of the upper disc; the attitudes of contacts of the model again coincide closely with those of the granite. The calculated points show good agreement except for the area just outside the eastern border. By assuming a density contrast of 0.07 gm/cm^3, a gravity model with a maximum thickness of 5 km occurring in the eastern part of the granite results.

Unless one postulates that the granite is only several hundred meters thick on the west side, is is impossible to account for the small anomalies

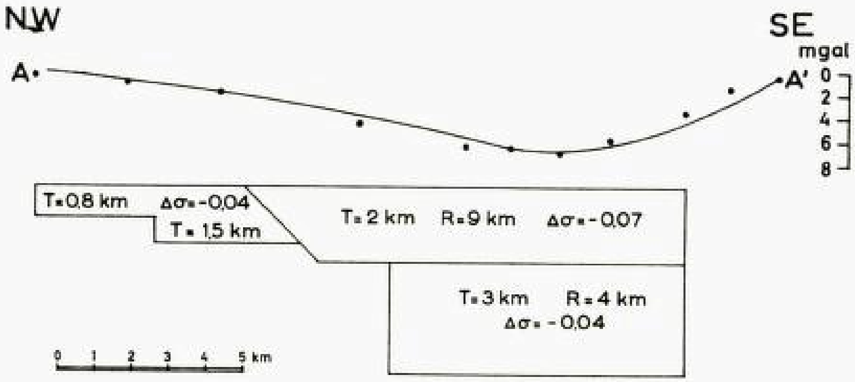


Fig. 8. Profile of gravity field from the residual anomaly map and the alternate model calculated to simulate the gravity effect of the Herefoss granite. Density contrast has been arbitrarily assumed to be low so that this represents the «maximum probable thickness of the granite».

here without assuming a small density contrast. A granite only several hundred meters thick appears unreasonable. The more likely explanation lies in the low density contrast on the west side of the granite. The rocks that lie just outside the western granite contact are not mapped in detail, but qualitative observations indicate that migmatite and granite gneiss occur commonly along this border. The only "dark" rocks that were observed during the field work were the amphibolites mentioned previously that coincide with the elongate anomaly on the west side of the granite. The mean density of the Telemark rocks was determined from rock samples scattered over a large area so that the density of rocks just west of the granite could deviate somewhat from the mean. The small anomalies over the west side of the granite are presumably attributed to the very low density contrast between the granite and Telemark rocks.

The anomalies are too low outside the Herefoss granite on the east end of cross section A-A', as can be seen where the calculated points fall above the measured profile. In addition, a shallow trough of negative anomalies extends between the Herefoss and Grimstad granites. This feature cannot be explained by the surface geology. A granite saddle which is about 1 km thick and 1 km deep and connects the two granites could cause this anomaly, but stations are sparse in this area so that control is poor.

Gravity anomalies offer a rare opportunity to calculate how much foreign rock is presently incorporated within part of the granite as has been done previously by Bott (1960) for the border zone of British granite. The north central part of the Herefoss granite is assumed to be 1 km thick; *i.e.*, the thickness of the upper disc in the first model. Any difference in the anomaly between this area and the "normal" area to the south is a function of the amount of denser foreign material included within the granite.

For the north-central part of the granite, the area within the —2 mgal line is considered to be that of abnormally high density. The area considered is the lined area of Figure 3. The average anomaly within this area is about —1 mgal, and the anomaly for "normal" granite that has a density contrast of 0.13 gm/cm³ with the country rock is —5 mgal. Since the attraction of a given body varies linearly as the density contrast between that body and its surroundings, we can write

$$\frac{x}{0.13} = \frac{1}{5}$$

$$x = 0.03$$

where x is the density contrast between the north-central part of the granite and the country rock. If the density contrast within this area is 0.03 gm/cm³, the mean density of this rock is 0.10 gm/cm³ greater than the mean density of "normal" granite. Further

$$\frac{y}{100} = \frac{.10}{.13}$$

$$y = 77$$

where y is the percentage of the granite that is foreign material. Therefore, *ca.* 77 per cent of the granite in this area is composed of inclusions with a mean density of 2.81 gm/cm³. Of course, the actual distribution could be quite different if considerable amphibolite (sp. gr. = *ca.* 3.00) or gray granite (sp. gr. = *ca.* 2.75) were present.

The reader should note the amphibolite layer that abuts against the north end of the granite (Fig. 4). This amphibolite coincides approximately with the outline of the 0-mgal line that extends into the granite on the north side (Figs. 3 and 4). The amphibolite may well be the cause of the zero anomaly within the granite here. The amphibolite could either continue into the granite at a shallow depth or be broken up into numerous fragments. The gravity meter can only confirm the occurrence of considerable dense material within the granite; it cannot resolve the physical coherence of this material.

The order of thickness of the mappable inclusion near the breccia can be determined. The inclusion causes an anomaly of slightly over 1 mgal. This anomaly corresponds to a disc of gneiss (sp. gr. = 2.81) 1 km in diameter and 400 meters thick.

Although the model calculated for the Herefoss granite appears somewhat peculiar, the foliations in the granite (Fig. 4) can be interpreted quite well in respect to the proposed form. The flow structure is near vertical around the "root" or "feeder". Near the breccia, however, it dips or plunges only moderately toward the "root". These same gentle attitudes are found in the thin part of the granite west of the breccia, but, near the western border of the granite, the flow structure becomes steep again to parallel the western contact of the granite. The mapped flow structure tends to follow the contacts of the model of the granite determined gravimetrically (Fig. 7C).

A similar occurrence has been described by Steenland and Wollard (1952). The foliation in the Cortlandt igneous complex delineated a basin-shaped structure. Steenland and Woollard were able to demonstrate from the gravity model that the magma must have flowed from feeders horizontally into a shallow basin in order to develop the mapped flow structure. Flow structure and gravity studies are complementary in the study of plutonic rock bodies.

The writer prefers the first model as the better approximation of the Herefoss granite because the densities used were those that were determined from the field samples. According to this model, the Herefoss granite has a general thickness of 2 km occurring near the east side. If we calculate the "maximum plausible thickness" using the minimum likely density contrast, a general thickness of 2 km and a maximum of 5 km on the east side result. Naturally, if we wished to select a density distribution whose density contrast decreased linearly with depth, we could arrive at an infinite thickness for the granite; however, this assumption would be quite speculative. Likewise the gravity interpretation cannot preclude the occurrence at depth of granite surrounded by granite gneiss, whose megascopic appearance would be quite different to a geologist; *i.e.*, the calculated "thickness" of the granite is the vertical extent of the density contrast, not necessarily of the textural contrast. The north-central part of the granite includes about 77 per cent dense material within the borders of the granite itself. The combined geological and geophysical information suggests an asymmetrical, funnel-shaped diapir granite that has incorporated a considerable amount of the country rocks; the gravity interpretation strongly confirms the geological observations.

The "Great Friction Breccia"

Since the "great friction breccia" is a large-scale geologic feature, its geophysical importance was not appreciated until a regional gravity map that encompassed a much larger area than the environs of the two granites was compiled. The breccia divides two geophysical entities as well as two geological regions. A detailed gravity study of the breccia is beyond the scope of this paper; nevertheless, enough gravity coverage is available to reveal the general features. The geophysical interpretation of the gravity anomalies associated with the breccia holds great interest because, as we shall see, it is intimately connected with the interpretation of the Herefoss and Grimstad granites and *vice versa.*

The "great friction breccia" was described by A. Bugge (1928) as a fault zone separating the Bamble formation from the Telemark formation. The breccia zone extends from Porsgrunn on the northeast where it runs into the Oslo province to Kristiansand on the southwest where it continues out into the Skagerrak. Recurrent movement has been noted along the breccia zone; other parallel breccia zones are found in the Bamble rocks to the east. Bugge believed this breccia zone, which was from 50 to 300 m wide, dipped southeast at 30°—60°.

Selmer-Olsen (1950) stresses the irregularity of the breccia zone, which may consist of one main fault zone or many adjacent small faults such as at Gjerstad where the displacement that is relatively downward and northeastward on the southeast side is distributed across a series of parallel faults. He verified dips of 45°—60° southeast along the fault plane of the breccia. Most of the movement along the breccia was Precambrian but some occurred as late as Permian. Besides the main breccia zone, a series of other faults which either parallel the rock contacts or transect them at high angles divides the Bamble rocks into blocks which have moved differently relative to each other.

Elders (1961, 1963) has also found the same sense of movement along the breccia in the vicinity of the Herefoss granite. Here, accurate mapping of the granite contacts and their attitude demonstrates that the east side of the granite must have moved downward and northeastward with respect to the west side. He also comments that the breccia strikes in two prominent directions, N25°—30°E and N50°—65°E; first one direction predominates then the other to give a zig-zag trace. The amount of displacement across the "great friction breccia" has not been determined.

It was hoped that the gravity survey would reveal an anomaly across the breccia so that an approximate value for the throw of the fault could be computed. In the interpretation of the Bouguer anomalies, however, another unexpected feature came into view. This feature is the large gravity gradient, the Bamble anomaly, over the Bamble area that culminates in a known anomaly of + 50 mgal along the coast. The large positive anomalies and high gravity gradient are seen to form a belt that runs from the Oslo province south to the Herefoss and Grimstad granites where it runs into the Skagerrak and seems to continue of the coast south of Kristiansand (Collette, 1960). Further off the coast the anomalies decrease again *(ibid.)*. Near Porsgrunn, the isoanomaly lines are deflected inland by the positive effect of the Oslo province, and, to the south, the Herefoss and Grimstad granites appear to form a buttress that forces the isoanomaly lines into the Skagerrak. The problem of the gravity interpretation of the breccia zone becomes, therefore, closely knitted to the whole exposure of Bamble rocks and possibly even to the Herefoss and Grimstad granites themselves.

A consideration of the gravity gradients, which are about 1 mgal/km over the Telemark rocks and 2 mgal/km over the Bamble rocks, on both sides of the breccia is illuminating; furthermore, the hinge between these two gradients is localized at or near the breccia zone. The Bouguer anomaly map (Plate 1) shows that the trace of the breccia zone is delineated rather closely by the 0-anomaly line except for the area southwest of the Herefoss granite. Although stations in the Telemark rocks are not numerous, the measured profiles suffice to establish that the gradient over these rocks varies from 1 mgal/km in the northeast to almost zero in the southwest. Similarly, the gradient over the Bamble rocks decreases from 2.5 mgal/km to 1.5 mgal/km. The one detailed traverse, which was measured across the breccia zone at Hynnekleiv 10 km north of the Herefoss granite, shows a positive anomaly of 2 mgal at the breccia (Fig. 9) and will be used for the gravity interpretation.

Normally the large regional gradient and positive anomalies that occur over the Bamble rocks would be attributed to a deep source; *i.e.*,

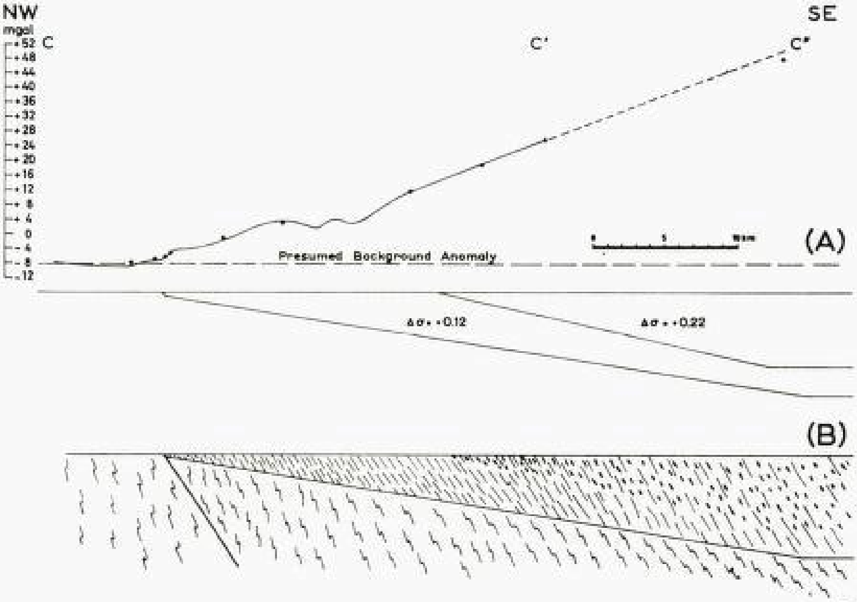


Fig. 9. (A) Profile of the gravity field from the Bouguer anomaly map and the model calculated to simulate the gravity effect of the «great friction breccia» and Bamble rocks. Model is 2-dimensional; *i.e.*, it extends to infinity perpendicular to this page. (B) Geological interpretation of the model. A wedge of more dense supracrustal Bamble rocks overlies the granitic Telemark gneisses. An increase in density of Bamble rocks must be postulated near the coast.

variations in crustal layering. The fact that the trace of the breccia forms a "hinge line" for the gravity gradient shows that the breccia zone controls the gravity gradient to a large extent. The cause of these anomalies can hardly even be located at moderate depth and still produce such a marked increase in the anomalies and gradient right at the breccia itself. The deeper a density contrast lies, the more gradual the gradient; if the source were a number of kilometers deep, the gradient would begin to increase well out in the Telemark area northwest of the breccia and would exhibit a broad step instead of the rather abrupt step measured at Hynnekleiv. The source of these anomalies occurs, therefore, at or immediately subjacent to the surface, but concomitant effects from a deep source cannot be excluded.

Since the breccia zone separates the Telemark and Bamble rocks, the source of the anomalies should be sought within these same rock units. At Hynnekleiv, granite gneisses of the Telemark area are separated from banded gneisses of the Bamble area by the breccia zone. The granite gneisses have a mean density of 2.67 gm/cm^3 for four samples, and the banded gneisses have a mean density of 2.79 gm/cm^3 for five samples; consequently, a density contrast of about 0.12 mg/cm^3 exists at the breccia. It is not surprising, therefore, that a "jump" in the gravity profile occurs at the breccia. Since the gravity gradient increases abruptly at this point also, the assumption that they are due to one and the same cause is reasonable.

The gravity model used to simulate these conditions consists of a horizontal plate of denser material that is underlain by lighter material. The northwestern contact dips under the plate at 45°, the approximate dip of the fault plane. The model (Fig. 9) was assumed to extend to infinity in a direction normal to plane of the page and was computed by Hubbert's line-integral method (1948b). Because the gravity increases, a thin wedge of denser material that thickens away from the breccia is used. Fig. 9 shows this model, for which a density contrast of 0.12 gm/cm^3 was taken. The wedge thickens toward the southeast from a thickness of 0.5 km at the breccia to 7 km off the coast. Another wedge of greater density was necessitated by the increase in gradient near the middle of the profile; accordingly, the upper wedge with a density contrast arbitrarily taken as 0.22 gm/cm^3 was added. The base value of the Bouguer anomaly was assumed to be —8 mgal over the Telemark area. The computed model agrees closely with the measured field; however, this in itself does not prove its validity.

There can be little doubt that the computed model is reasonably correct in the vicinity of the breccia where granite gneiss adjoins banded gneiss, and the anomalies over the western part of the Herefoss granite indicate that the gneiss here has almost the same density as the granite. The overall density of the Telemark formation can hardly be as low as 2.67 gm/cm^3. Also, the density contrast of 0.22 gm/cm^3 used for the upper wedge seems unusually high; nevertheless, the presence of hyperites and mafic members of the arendalites might conceivably raise the density this much. Although part of the anomalies must be caused by a density excess near the surface, combined shallow to moderate and deep sources could cause the anomalies; *i.e.*, a flat slab of denser material underlain by coastward rise in the top of the "basaltic layer" or Moho. A small amount of the gravity gradient is probably caused by the northward thickening of the crust toward the Caledonian mountain range. The difficulty with a deep source still

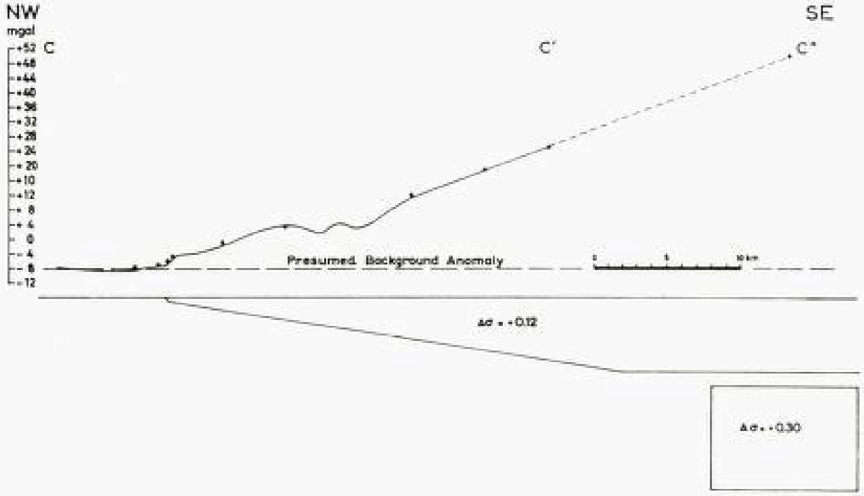


Fig. 10. Profile of the gravity field from the Bouguer anomaly map and the alternate model calculated to simulate the gravity effect of the «great friction breccia» and Bamble rocks. Model is 2-dimensional. Instead of the increased density of the Bamble rocks as in Fig. 10, a mass of high density at moderate depth is postulated.

remains that the interface between the density contrast must either occur at improbably shallow depth or have an incredibly steep inclination; an alternate solution is presented in Figure 10.

In the place of the upper wedge, the second model (Fig. 10) employs an infinitely long mass excess of rectangular cross section that occurs at moderate depth. The density contrast for this prism is arbitrarily selected to be 0.30 gm/cm^3, approximately the density contrast that gabbroic rocks present. The mass is 10 km wide and 7 km deep, and its top lies 6 km below the surface. In conjunction with this model, the dense wedge at the surface attains a maximum thickness of only 5 km.

The two models presented here are extremely hypothetical in the vicinity of the coast; nevertheless they do indicate the type of mass distribution that could cause the anomalies. Thinning of the crustal layers under the Skagerrak could cause part of the anomaly, but most of the isostatic anomalies (Collette, 1960) in the Skagerrak are very small and fail to confirm upwarps of at least the Moho.

Possibly the biggest aid toward a better interpretation would be a large number of density profiles (Nettleton, 1939) together with seismic profiles that would reveal the behavior of and depth to the Moho and "basaltic layer". In addition, a number of detailed gravity profiles across the breccia coupled with detailed geologic mapping are required.

The gravity model shows an apparent throw of 0.5 km along the breccia zone. Since the anomaly-producing rocks are surface formations, the value obtained is the minimum throw that could have occurred. Some of the faulted rocks have certainly been removed by erosion so that the throw must have been greater than the model shows.

The gravity model has more interesting geologic implications than the displacement along the breccia zone. The computed model presupposes that the main distinction between the Bamble and Telemark rocks on either side of the breccia is merely different levels of exposure. In other words, the granite gneisses and migmatites of the Telemark area which occur beneath the Bamble rocks and are found at increasingly greater depth as the coast is approached would then continue out under the Skagerak where Collette 1960 has observed isostatic anomalies of + 40 to + 60 mgal off the coast at Risør and Kristiansand. Collette invokes the mafic arendalites as the possible source of the isostatic anomalies. The map (Pl. 1) shows that the anomalies appear to continue from the breccia zone out to the submarine stations in the Skagerak in a belt that overlies the Bamble formation and would seem, therefore, to have a common source. The postulated faulting (Holtedahl, 1940) that is supposed to have formed the "Norwegian Channel" may be a form of local isostatic compensation whereby the dense rocks that cause the large anomalies have been downdropped.

To call these supracrustal rocks of the Bamble formation distinctly heavier than the underlying migmatites may seem surprising because low-density quartzites and synkinematic granites are fairly common in the Bamble area. The Bamble rocks are regarded to be a packet of interfolded supracrustal rocks that thickens toward the Skagerrak (Fig. 9B) and can be simulated by the model of Fig. 9A. That the Telemark migmatites and granite gneisses are probably less dense than the Bamble supracrustal rocks has been demonstrated by the density determinations; an interesting parallel can be drawn from the Mt. Gaustad region of inner Telemark. Here the Telemark supracrustal rocks, consisting largely of interlayered quartzites and amphibolites, overlie the Telemark gneiss granites (Dons, 1960). The outcrop of the supracrustal rocks coincides with a positive Bouguer anomaly. The author has noticed the same phenomenon in gravity measurements on the west side of the Flå granite, but the anomalies are only a few mgal over interbedded quartzites and amphibolites. These facts suggest that metamorphosed supracrustal rocks may commonly be more dense than their migmatitic substratum.

The gravity survey cannot, of course, determine the nature of the contact between the Bamble rocks and the presumed migmatitic substratum. The writer's interpretation is that this contact is the migmatite front of Wegmann (1935) that separates the migmatitic infrastructure from the superstructure while the Bamble supracrustal rocks are in the transitional zone of regional metamorphism. The Bamble rocks are migmatitic in places, but they would hardly be confused with the great area of migmatites and granite gneisses west of the breccia. The geologic map, itself, offers striking evidence of the difference in migmatization because numerous mappable rock units are recognized east of the breccia and very few west of the breccia. Besides the migmatite front, the other possibilities include a depositional contact or a tectonic contact. The nature of the contact between the Bamble supracrustal rocks and the Telemark gneisses has not been investigated other than the fault relation between them. In southern Telemark, the contact between gneiss granite and the Telemark supracrustal rocks, although still inadequately known, is intrusive in places and gradational in others so that anatectic processes are inferred (Dons, 1960, p. 56).

Barth (1947a, p. 9—10) has written concerning the breccia zone and the Bamble and Telemark rocks in the area investigated: "Furthermore at many places a rock trespasses across a fault line . . . Thus it would seem reasonable to assume that the two formations were mutually related . . . Therefore the granitized Telemark formation should represent deeper strata than the less granitized Bamble formation . . . In the direction across the strike, the gneisses become, generally speaking, more granitized landinward; this may correspond to a stepwise elevation of the county — each fault representing a new step — and thus successively deeper and more granitized strata ought to be exposed as one proceeds from the coastal regions in the southeast towards the interior of the country."

The computed thicknesses of Herefoss and Grimstad granites bear noteworthy relationships to computed thickness of the Bamble rocks. The granites have approximately the same calculated thickness as the wedge of supracrustal rocks. In other words, the thinner Herefoss granite occurs near the edge of the wedge, and the thicker Grimstad granite is found near the middle where the wedge is thicker (Fig. 18). In spite of the similarity of the density contrasts used for these calculations, it is interesting to note that the thickness of the wedge of denser material necessary to account for the Bamble anomalies has about the same thickness as the granites that penetrate it. This fact implies that the Herefoss and Grimstad granites are mobilized (and differentiated) Telemark substratum that has penetrated the Bamble supracrustal rocks.

The Bouguer gravity map further demonstrates this point. The west side of the Herefoss granite causes practically no gravity anomaly. The density of this granite is, therefore, almost the same as the density of the surrounding gneiss except for the amphibolite lense that causes the small positive anomaly near the west contact. Also, the isoananomaly lines that follow the breccia swing around the eastern border of the Herefoss granite so that its eastern contact has about the same gravity effect as the breccia north of the granite. South of the Herefoss granite, the augen gneisses that occur with the negative anomalies here may be offshoots from the migmatitic substratum at shallow depth. The gravity effect of the Herefoss granite resembles that of the Telemark gneisses along the breccia zone; therefore, the granites can be regarded as projections of the Telemark gneisses through the Bamble supracrustal rocks.

The arendalites along the coast are granulite — facies rocks that occupy a strange position in relation to the calculated model. They occur in the thicker part of the supracrustal rocks but at what seems to be a higher structural level. This is obviously a point worthy of further investigation.

The interpretation of at least part of the Bamble anomalies as the effect of a wedge-like body of denser supracrustal rocks overlying the migmatitic substratum, though not unique, offers several advantages. The close connection between the gravity gradient and the breccia and the positive anomalies over the Bamble rocks are best explained by this interpretation. Similarly, the deviations from the regional pattern at Gjerstad and south of the Herefoss granite are easily clarified if the mass excess is at or very near the surface. Finally, the anomalies of the Herefoss and Grimstad granites and the breccia and Bamble area fit together to give an integrated picture of the geology of the region.

Evje Amphibolite

The Evje amphibolite is an elongate amphibolite body that is located in the Telemark gneiss area between Evje and Iveland. The area including the amphibolite has been described by Barth (1947a, 1960). The amphibolite extends about 30 km N-S and measures from 2 to 10 km across. The Evje amphibolite is known for its rare-mineral pegmatites as well as its nickel deposits.

The rocks of the Evje amphibolite have been almost completely recrystallized so that amphibolite composed of hornblende, andesine, biotite, and quartz predominates. Massive gabbros and norites that contain relict hypersthene occur within the amphibolite. This body is probably related to the smaller meta-norites of the Bamble area but has been more thoroughly metamorphosed at a deeper level.

The amphibolite is surrounded by a monotonous succession of granite gneisses and migmatites which are attributed to regional granitization. The gneisses have been deformed around north-trending axes and plunge gently northward. These gneisses have reacted plastically in relation to the stiff Evje amphibolite and are wrapped around it so that the foliation in the gneiss conforms to the gneiss-amphibolite contact everywhere.

The contacts between the amphibolite and gneiss are gradational and difficult to place. In contrast to the small norites in the Bamble area, which often have sharp contacts, the original contacts of the Evje amphibolite have been obliterated by an advanced stage of granitization. Lenses and layers of granite gneiss wedge themselves into the amphibolite and form a banded gneiss which, in turn, grades into the common gneiss; concomitantly, the amphibolite may become biotized and pass from a biotite schist into the usual granitic gneiss. In addition, areas in various stages of granitization appear within the amphibolite and complicate the mapping. Barth (1947a, p. 11) remarks that these transitions render cartography of the geology difficult so that is is impossible to place an exact boundary on a map.

Numerous potassium-rich granite pegmatites, which are lenticular bodies, cut the amphibolite. They contain abundant rare minerals and have been quarried both for these minerals and feldspar. No pegmatites occur in the enclosing gneiss.

We would expect that such a large amphibolite body enveloped by granite gneiss would cause a positive gravity anomaly, and this is, in fact, the case. One of the largest anomalies on the Bouguer anomaly

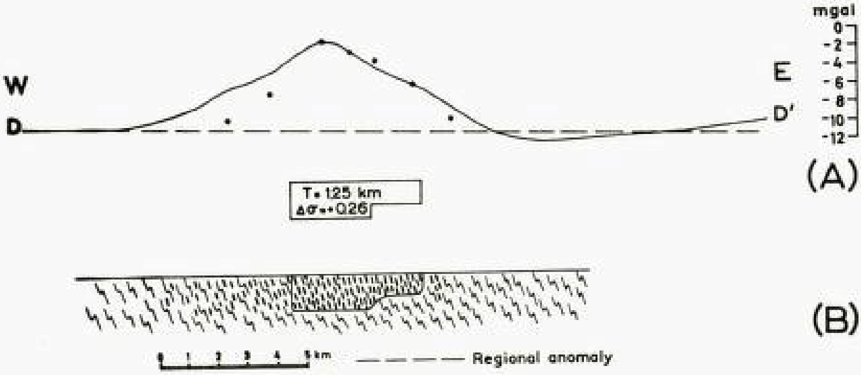


Fig. 11. (A) Profile of the gravity field from the Bouguer anomaly map and the model calculated to simulate the gravity effect of the Evje amphibolite. Model is 2-dimensional. (B) Geological interpretation of the model. Disagreement between the model and the observed gravity curve is caused by gradational contacts.

map occurs over the amphibolite and forms a prominent feature against the monotonous background value over the Telemark gneiss. The maximum relief in the Bouguer anomalies is from -9 mgal to + 2 mgal and yields an anomaly of almost + 12 mgal that is due to the amphibolite. The widely scattered stations indicate that the isoanomaly lines generally follow the mapped outline of the amphibolite.

In order to compute a gravity model, two profiles were drawn through the amphibolite. These appear in Figs. 11 and 12; profile D-D′ is based on far better data. The density determinations (Table 2) show a density contrast of 0.26 gm/cm^3 between the Evje amphibolite (3.03) and the Telemark formation (2.77). The maximum anomalies along profiles D-D′ and E-E′ are about 10 and 12 mgal respectively. The attraction of the model was computed graphically by means of a graticule (Jung, 1927). Simple 2-dimensional models (the model is assumed to extend to plus and minus infinity in a direction perpendicular to the page) are used to approximate the gravity effect of the amphibolite. A plane rectangular cross section suffices in this case where the gravity information lacks detail and the geologic contacts are transitional. As can be seen from Figs. 11A and 12A, the models possessing a simple cross section yield values for the gravitational attraction that agree rather closely near the centers of the profiles but exhibit greater and greater errors as the edges of the model (amphibolite contacts) are approached. This error is very likely due to the gradational

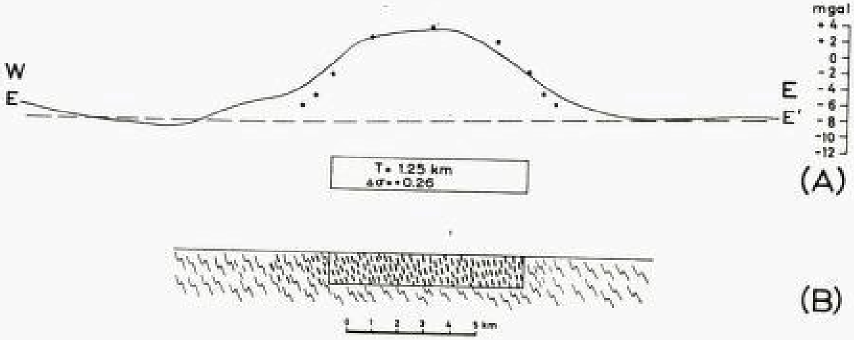


Fig. 12. (A) Profile of the gravity field from the Bouguer anomaly map and the model calculated to simulate the gravity effect of the Evje amphibolite. Model is 2-dimensional. (B) Geological interpretation of the model. Disagreement between the model and observed gravity curve is probably caused by gradational contacts. The larger anomaly in this profile is caused by the greater width of the amphibolite.

nature of the contacts. The cross sections in Figs. 11B and 12B are drawn in order to give a qualitative representation of the probable cause of these errors which are particularly marked on the west side of profile D-D′. The most important result of the two gravity models is the fact that both models give a thickness of 1.25 km for the amphibolite; the difference between the maximum anomaly in both profiles appears to be due to the width of the body rather than the thickness.

Oddersjå Granite

Because of its markedly different character, the Oddersjå granite was selected for a limited study. This granite is cut obliquely by only one road through a valley of such abrupt relief that the terrain effect could well exceed the expected anomaly of the granite; therefore, the gravity meter was carried along part of a traverse that was a compromise between favorable terrain and a necessary location. The elevations along this profile were rendered more accurate relatively because a long, narrow lake that extends across most of the granite was utilized as a common surface for the measurements. Fortunately, this one profile is almost normal to the strike and near the center of the granite so that it should be representative.

The geology of the Oddersjå granite has been described by Barth (1956) and Barth and Bugge (1960). The granite is an elongate body about 2 km wide and 10 km long that is conformable with the surrounding migmatitic gneiss (Plate 2). The rock resembles an augen gneiss that is strongly lineated and foliated. The lineation strikes N-S and is almost horizontal, but foliation dips steeply. The foliation which does not increase toward the borders, was formed by the same orogenic forces that deformed the surrounding gneisses rather than by magmatic flow; the granite is synkinematic. The contacts of the granite may be either sharp or gradational. The gradations occur either by passing from the gneiss through a larger scale migmatite zone into granite or by addition of microcline crystals to an amphibolite. The Oddersjå granite is composed of microcline, plagioclase An_{27-30}, quartz, biotite, and accessory garnet, in places. A fine-grained, massive granite that is called the Sødal granite and occurs at the south end of the Oddersjå granite shows both conformable and intrusive relations. The origin of the Oddersjå granite is ascribed to granitization processes that resulted in anatexis.

The profile measured is designated profile F-F′ on the Bouguer anomaly map. The variations in the gravity field along this profile were so small that they do not influence the isoanomaly lines on the map; therefore, they are receding into normal error of the survey. Fotunately, the fact that the elevation error was minimized by use of the above-mentioned lake increases the usefulness of the small anomaly that was found.

Profile F-F′ reveals a small negative anomaly of 2 mgal over the Oddersjå granite (Fig. 13). The anomaly is, however, asymmetrical in respect to the granite contacts. According to the geologic description of the granite, the large size of the anomaly is somewhat surprising. Provisional density data (Table 2) give a density contrast of -0.10 gm/cm^3 between the granite (2.73) and gneiss (2.83). Again, the attraction of a simple 2-dimensional model was computed graphically by means of a graticule (Jung, 1927). Since the profile cuts the granite obliquely, a correction was applied to the calculated values and the cross section appears to be too wide. The calculated model has a rectangular cross section with a thickness of 1.4 km and a surficial projection 0.3 km thick on the west side. The writer would rather regard this projection as a function of the density contrast rather than the actual shape of the granite body; *i.e.*, the west contact of the granite grades into the gneiss such that the projection represents a migmatitic border zone whose depth is the same as that of the granite but whose mean density falls somewhere between the densities of

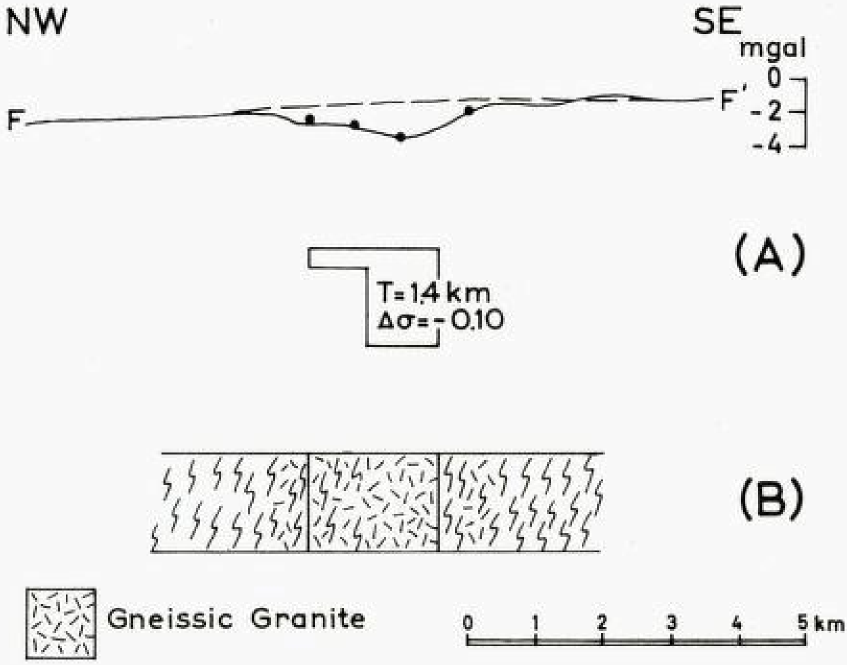


Fig. 13. (A) Profile of the gravity field from the Bouguer anomaly map and the model calculated to simulate the gravity effect of the synkinematic Oddersjå granite. Model is 2-dimensional. (B) Geological interpretation of the model. The thin western projection is interpreted as a function of the gradational contact rather than the true shape.

the granite and the country rocks. Figure 13B shows the writer's geologic interpretation of the gravity model. This study indicates that small migmatitic synkinematic granites may have small anomalies but a "thickness" that is not insignificant.

Gravity anomalies and granite emplacement.

Introduction

In 1930, Wegmann proposed that granitic bodies can be intruded as diapirs and compared the mechanism of intrusion with that of salt diapirs. Since that time, a wealth of geophysical information has revealed a further analogy between granites and salt domes; the gravity effect of granites and salt domes is almost universally negative. This fact further suggests that the analogy between granites and salt domes is a useful one and offers valuable information with regard to granite emplacement.

Granites have been shown to possess characteristic gravity profiles (Bott, 1956). A mode of emplacement similar to a salt diapir is one means of explaining this profile.

Characteristics of Salt Diapirs

The rise of salt bodies through younger strata has long been known; however, considerable disagreement has existed regarding the causes of this phenomenon. The three possible causes are the high plasticity of salt, lateral compression, and the hydrostatic force due to the low density of the salt. These properties have similar counterparts in granite emplacement although the similarity cannot be carried too far.

In folded areas in Europe, Pustowka (1929) and particularly Stille (1925) have emphasized the importance of lateral compression and the greater plasticity of the salt while, in America, Nettleton (1934, 1943) and Parker and McDowell (1955) have stressed the efficacy of the low density of the salt in the Gulf Coast region, which is undeformed except by the salt diapirs. Actually, all three properties must play an important role in which compression and high plasticity dominate in orogenic regions and high plasticity and low density predominate in non-orogenic regions. It is apparent that these three features can become effective in varying degrees during the emplacement of granite.

The source for the salt diapirs has been a salt bed or rather a succession of beds. Stille *(op. cit.)* writes that flat-lying internally deformed salt beds on the border of the North German Basin pass into salt anticlines with thickened cores, and finally into salt stocks that break through the crests of the anticlines in the center of the Basin. All these features may occur along one anticlinal axis. The diapirs may be either elongate or stock-like. The interference of tectonic elements commonly determines both the location and form of salt uprise; salt stocks form at the intersections of differently directed fold axes.

Like the German occurrences, compression has played an important role in the Rumanian salt diapirs (Pustowka, *op. cit.)*. Thick salt beds have been thrown into folds and have continued their uprise in the anticlines until they have penetrated through the suprajacent beds, which were dragged along. These salt diapirs are solely elongate bodies.

Pustowka confirms the importance of tangential stress and high plasticity for the formation of salt diapirs but also places some importance on the low density of the salt.

In the American Gulf Coast region, salt diapirs are exclusively circular bodies either like the frustrum of a cone or cylinders that exhibit a slight "overhang" at the top (Balk, 1953). They occur in strata that are unfolded except in the immediate vicinity of the salt intrusions. These diapirs have domed the overhead strata while the surrounding beds have collapsed into a rim syncline, a depression that is commonly found encircling the salt intrusions. These salt plugs are almost always associated with negative gravity anomalies; in the absence of any regional deformation, the low density of the salt is considered to be the driving force that caused the upthrust, in which the relative plasticity of the salt was also important (Nettleton, 1934; Parker and McDowell *op. cit.*).

The internal structure of Gulf Coast salt domes has been studied by Balk (1947, 1949, 1953). Balk finds a strong vertical orientation of the structural elements. Tightly appressed folds in interlayered salt stand on vertical axes, and crystals of elongate anhydrite crystals within the salt produce a vertical lineation. According to the experiments of Escher and Kuenen (1929), this type of folding develops when interlayered material of different flow properties is extruded through an annulus. The fold hinges are often erased near the contacts of the salt body so that a rather uniform planar structure that parallels the border develops. Circular, oval, and sigmoidal fold traces show that, within the main dome, there must be many small ones.

The salt deposits are remarkably dry (Balk, 1949). If a pore fluid was present during the intrusion, it has escaped. Movement was achieved by repeated internal slip and recrystallization of the halite while the comparatively unyielding gypsum was carried along passively in the flowing halite groundmass.

The internal structure of the Rumanian salt diapirs is described by Pustowka (1929). Where salt is interlayered with more competent rocks, motion is achieved by slip within the salt layers. As movement continues, tension cracks form in the competent layers, and the salt sends apophysis-like offshoots into these layers. Finally if motion continues, the mass of rock attains the appearance of a breccia in which angular fragments of competent rock float in the salt.

Experiments with model tectonics have given further important data regarding salt dome emplacement. Nettleton (1934) used liquids of different density in his early experiments. This presupposes that low density is the driving force for the rise of the salt. Nettleton (1943) later confirmed that his experiments had been consistent with the recently developed theory of scale models (Hubbert, 1937). Nettleton showed that a peripheral sink, analogous to the rim synclines, found in nature, forms around the rising diapir. This peripheral sink may eventually pinch off the flow of material into the diapir and thereby controls the volume of the rising mass. The more viscous the fluid that simulates the overburden, the slower the rim syncline forms, and the more material will flow into the diapir. Nettleton *(op. cit.* p. 1192—93) found that, by selecting a relatively high-viscosity material for the overburden, the underlying low-density layer will flow into a broad, balloonlike dome. He concludes that the physical properties of the overburden are instrumental in determining the size and shape of the diapir.

Parker and McDowell (1955) performed dome-simulating model experiments using asphalt for the low-density layer and granular material that is somewhat too strong to give quantitative results *(ibid.* p. 2402)for the overburden. The asphalt tends to assume a circular cross section as it rises, and the diameter of the base is greater than the diameter of the top. If the rising dome encounters a more rigid layer during its passage through the overburden, then "overhang" develops near the crown of the column. Oval domes, initiated by an elongate ridge in the surface of the asphalt layer, are surrounded by oval peripheral sinks. The diameter of peripheral sinks is controlled somewhat by the strength of the overburden, but maximum sink development occurs where the source layer is thickest. The diameter of peripheral sinks tends to be 6 to 8 times the diameter of the dome. According to these experiments, the width of a diapir is approximately equal to the thickness of the source layer ,but the applicability of this is governed by the above-mentioned reservation.

Considerable attention was paid in the experiments *(ibid:)* to the mechanism by which a diapir is initiated. In general, diapirs start to grow from any positive irregularity in the surface of the low-density layer. These irregularities can form by lateral variations in pressure caused by variations either in thickness or in density of the overburden, tilting of the layers, or folding. Of particular importance is the fact that only very gentle folding is required to initiate diapirism and that the uprise always occurs along anticlines, never along synclines (Parker and McDowell, *ibid.*, p. 2411—12).

The behavior of salt diapirs reveals how granite diapirs might be expected to develop and to act. Certain problems of granite emplacement, moreover, are resolved by a comparision between the tectonics of granite and salt.

Diapiric Emplacement and the Room Problem

In spite of the general controversy that rages around granite emplacement, few geologists would deny the similarities in properties between salt diapirs and at least some granites. In the property of low density, salt diapirs and granites are certainly very similar as has been shown by both gravity surveys and sampling. Like salt, the plasticity of many granites must also be considered to be greater than their surroundings. The great dissimilarity between the two is in the chemical and possibly the thermal effects of granite.

Due to its mineralogy, the density of a granite is almost always less than the average density of its enclosing rocks if they are nonporous. This fact is supported by the common occurrence, cited at the beginning of this paper, of negative gravity anomalies over granite. Granite bodies will, therefore, attempt to rise ,and whether they do or do not ascend depends on the interplay of the plasticity of the granite, the strength of the country rocks, the amount of the density difference between granite and country rocks, and the intervention of tectonic forces. Since the tendency for granite to ascend is universal and particularly effective in anticlines, domal structures would commonly occur.

The immediate consequence of a comparison between salt diapirism and granite diapirism is that the room problem is eliminated. As a salt diapir rises, it thrusts aside the surrounding rocks, which subside into the rim syncline, and it domes the suprajacent rocks. There is no reason why the same process would not be effective in the case of a mobile granite body; the form of the granite diapir could either be round or oval depending on the action of compressive stress.

Naturally, the comparison is valid only where the granite has strongly deformed the country rock. Around the granite, one would expect to encounter the same steeply inclined strata that slope into a rim syncline (see Wegmann, 1930, p. 71) like those occurring around salt diapirs. In plan, the country rocks have been deflected so that they strike, in general, parallel to the adjacent granite contact; however, in a vertical section, the granite contact may or may not conform with the surrounding strata. Such features are common in the plutons of the catazone and lowered mesozone (Buddington, 1959). Macgregor (1951) describes an area of granite domes in which the intervening regions appear to be synclinal. Mantled gniess domes (Eskola, 1948) appear to be an initial stage of a particular type of diapir development, and Eskola has invoked the density contrast as the driving force for the rise of the domes. The salt-diapir solution of the room problem is applicable where analogous deformation of the surroundings can be demonstrated; the analogy cannot resolve the room problem for epizonal granites for which an emplacement by stoping is postulated.

Clearly, the study of granites depends as much on tectonics as on petrology. The room problem necesitates the study of the tectonic relations in the granite and surrounding rocks, and the demonstration of "shouldering" and a rim syncline together with foliation in the granite suggests diapiric emplacement which diminishes the room problem. Experimental investigations of diapirism (Nettleton, 1934, 1943; Dobrin, 1941; Parker and Mc Dowell, 1955) have shown the dominant influence of the overburden on the shape and growth of a diapir. Investigations that reveal the geometry of the rim syncline and the plastic properties and heterogeneity of the country rocks may, through model studies, allow the elucidation of the granite's form and ultimately even permit quantitative estimates of size and depth of the granite's source by means of such formulas as appear in Nettleton (1934, p. 1187).

Source of the Granite

A wealth of information regarding melting in the crust has become available recently. Wyllie and Tuttle (1960) propose that partial melting can be expected within a geosyncline at depths of about 20 km. Wyllie and Tuttle (1958) showed that differential melting of shales of widely different compositions begins at nearly the same temperature and that a granite melt results; similarly, the melting of arkoses would also give a granitic melt at about the same temperature as shales. Wyart and Sabatier (1959) also melted shales hydrothermally to obtain a granitic liquid in a crystal mesh of cordierite, spinel, and biotite and suggest that granitic pore magma must be common at depth.

Winkler and von Platen (1958) have found that clay melts differentially to form a granitic liquid and that, within a temperature range of only 4° C, 45 % of the clay liquified. The residual material consisted of cordierite, biotite, sillimanite-mullite, hematite, ilmenite, corundum, and rutile. If a clay contained less Fe and Mg, more granitic melt was formed *(ibid.)*. Although the initial melt is granitic for rocks of different chemical compositions, the amount of pore melt formed will depend on the chemical composition of the rock.

The source of granite magma can hardly be regarded as a magma "chamber" in light of these experimental studies. Since chemical composition of supracrustal rocks is mainly a function of layering, the occurrence and amount of melting will be determined by the layering. Within limits set by folding, melting should roughly follow a certain horizon that is determined by its chemical composition, fluid content, and heat supply. Instead of occurring in the classical magma "chamber", the magma source would be a layer composed of a pore melt and solid residuum. This layer or sheet of granitic melt can be likened to a salt layer; under the proper conditions, it will begin to flow and form a granite diapir.

Sometime before it is emplaced, the pore melt must be separated from its solid mafic residuum. Tectonic forces are commonly invoked to accomplish this. Various authors (Winkler and von Platen, *ibid.;* Wyart and Sabatier, *op.cit.)*, however, have recognized that this process may be incomplete to give intrusives of "mixed" appearance. By comparision with slag, Sosman (1948, p. 116) has estimated that an intrusive granite may flow when it is 95 % crystalline. If one considers the tectonic environment and the probable mobility of a rock which is 5 to 50 % liquid, it seems remarkable that the pore melt would have a chance to separate from the residuum at all.

As recently summarized by Reynolds (1958), the transformist school, on the other hand, does not believe the intervention of magma is necessary for granite diapirism to occur. They would invoke synkinematic migmatization which increases the plasticity of the rocks as they become more granitic at certain structural levels. Then under orogenic stress, the migmatites flow into structurally determined arches and break through the suprajacent strata. During their ascent, the migmatites are homogenized by differential movements until the final product is a discordant rather uniform granite emplaced in a higher structural level than that in which it originated. The basic principles of this process have been described by Wegmann (1930, 1935). According to these views ,discordant granites can be intruded without the intervention of a magma.

Balk (1953, p. 2472—73) has even suggested the possibility that granite may move in a similar fashion as a salt aggregate. At deep levels, quartz may deform and recrystallize, carrying along the feldspar in suspension much as halite flows in a solid state while the anhydrite crystals are carried along passively. The analogies with salt tectonics demonstrate that such typically "igneous" features as apophyses and breccias may occur in rocks in the solid state when one rock unit is considerably more plastic than the surrounding ones; the concept of complete or partial fluidity for an intrusive granite is not always necessary.

Whether the granite rises as a true melt or plastic mass, the location of a granite diapir is certainly conditioned by the regional tectonics. Wegmann (1930) described the rise of granite diapirs at the intersection of two fold directions and Stille (1925) has done the same with salt diapirs. The Finnish gneiss domes (Eskola, 1948)provide an excellent example of this phenomenom. In addition, the results of Parker and McDowell (1955), which show that diapirs are always initiated in anticlines and never in synclines, provide experimental confirmation for the natural occurrences. Diapirs should be localized along anticlinal axes and particularly at axial culminations. The lower density of granite creates a universal tendency for the granite to try to rise, but the orogenic stress and structural geometry are the trigger and localizer for granite diapirism.

The determination of the physical state of a granite during its emplacement is the problem of the structural geologist and petrologist; this problem lies outside the scope of gravimetry. *Hypotheses regarding the origin of the granite, however, imply certain density distributions, and these density distributions are very much within the field of gravimetry.*

Gravity Anomalies and the Granite Problem

The emplacement of a granite results in a local mass deficiency because granites are commonly less dense than their surroundings; therefore, negative gravity anomalies are found over granites. From the gravimetric standpoint, granite emplacement presents a mass problem rather than a room problem. Where is the missing mass? The disposition of the removed mass is reflected in the various hypotheses of emplacement; *i.e.*, stoping, forceful intrusion (including diapirism), replacement with downward migration of the heavy elements, replacement with formation of a "basic front". Bott (1956), who has treated the granite problem from a gravimetric viewpoint, has made a contribution of great significance to the granite problem.

For a complete discussion of the subject, the reader is referred to Bott's *(ibid.)* interesting paper; however, a few pertinent facts will be considered here. Negative gravity anomalies occur over granites, and the shape of the anomaly profile is partly a function of the disposition of the compensating surplus mass, the mass removed when the granite was emplaced. Using various assumptions about the distribution of this removed mass surplus, Bott *(ibid.)* has calculated a number of gravity profiles for different theoretical mass configurations. Bott obtains two mass configurations which satisfy the typical observed profiles of granites. One model (Fig. 14) pictures the mass deficiency of the granite body without any compensating mass surplus. The other model (Fig. 15) represents the granite with its compensating mass surplus at depth well below the lower limit of the granite. This model is supposed to simulate a granite that has been emplaced by stoping, but the stoped country rock has sunk to so great a depth that its gravity effect is minimized. Bott has documented the similarity between the calculated profiles and measured profiles over granites with numerous examples, and these are in complete conformity with the writer's experience excepting the small deviations that one would expect in nature.

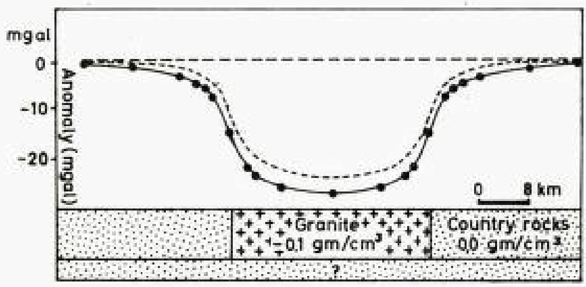


Fig. 14. Gravity profile of a prismatic mass deficiency (2-dimensional) without a compensating surplus mass. Broken line shows gravity profile of a vertical cylinder. The curves are similar in shape and are typical of gravity profiles found to occur over granites. (Adapted from Bott, 1956.)

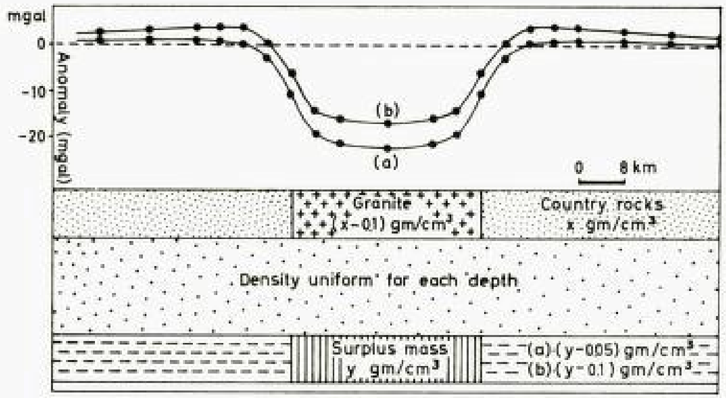


Fig. 15. Gravity profile of a prismatic mass deficiency (2-dimensional) with an underlying mass surplus at a much greater depth. Curve (a) is drawn for a mass surplus equal to one half the deficiency and curve (b) for an equivalent surplus. Although curve (b) deviates somewhat from the ideal profile over granites, the deviation is probably within the resolution power of many gravity surveys. (Adapted from Bott, 1956.)

Bott concludes that stoping is a means of emplacement that is fully consistent with the gravity anomalies. He also qualifies that this analysis is not meant to apply to syntectonic intrusions, but to those with undeformed borders and sharp crosscutting contacts. This type of pluton would be classified as epizonal.

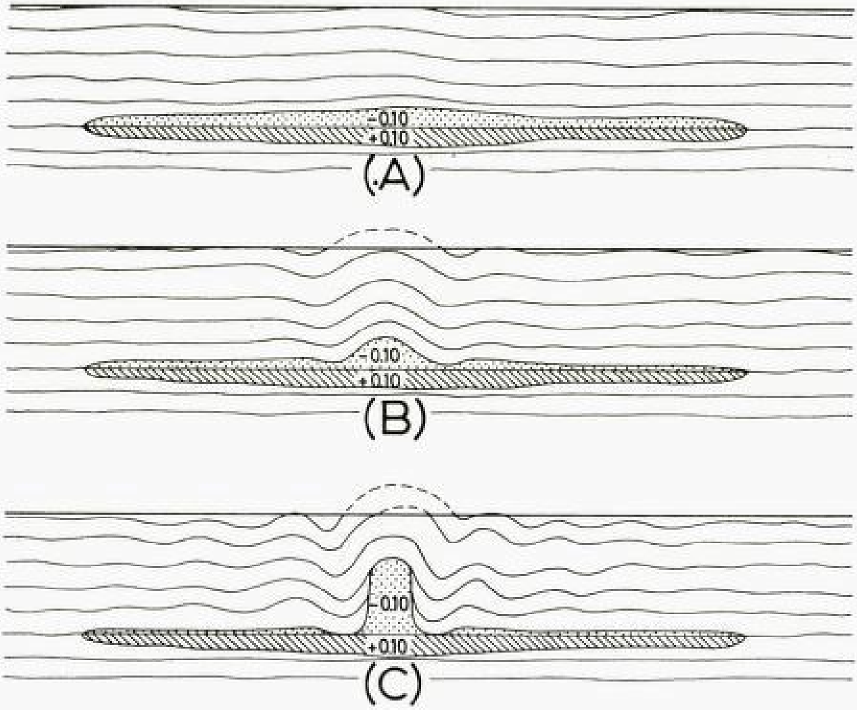
The granites discussed in this paper belong to the deeper levels of the meso- and catazone and would be called late-to postkinematic except for the synkinematic Oddersjå granite. Since these Precambrian granites must represent a deep level of exposure and since modern theory (Wyllie and Tuttle, 1960; Barth, 1961) places the source of the granite near its base within the geosyncline rather than at the base of the crust, these granites form an important link in the overall picture. The anomalies over Precambrian granites resemble the younger granites and Bott's two theoretical models. We should remember that the amount of excess mass representing either mafic differentiates or the missing country rock may be much larger than the mass deficiency of the partially eroded granites that we commonly study. The problem becomes even more acute if we picture the present Precambrian shields to be the eroded roots of mountain chains that contained hugh Pacific-coast-type batholiths. The important result of gravity studies of granites is that *no gravity survey has, as yet, detected either the compen-*

sating mass surplus that represents the missing country rock or the mafic residuum that would be expected from magmatic differentiation.

The conclusion is inescapable that the displaced country rock and the mafic differentiates must either be very deep or be dispersed. The gravity effect of a mass can be minimized by placing the mass as deep as possible or by spreading the mass over such a large area that its effect becomes part of the background anomaly.

The first task is to dispose of the country rock that is now occupied by the granite; *i.e.*, the room problem. Diapiric intrusion eliminates the room problem by analogy with salt diapirs, but the accompanying deformation must be demonstrable. Replacement also solves the problem if the heavy ions are dispersed so that a majority of them do not remain just outside the granite as a "basic front" or "basic behind". Stoping in which the stoped mass sank relatively deep would normally be permissible gravimetrically; however, the deep level of exposure for Precambrian granite effectively shortens the distance from this mass to the gravity meter and increases detection possibilities. Stoping may be only a minor factor in the emplacement of deep granites but, from a gravimetric viewpoint, cannot be completely excluded if the stoped mass excess were eventually dispersed by metasomatic processes. Forceful intrusion with uplift of the country rock along fractures and subsequent dispersion by erosion (Noble, 1952) is another possibility. The gravity data allow the above-mentioned possibilities, but geologic mapping is the ultimate determinant.

Although the granite may have been emplaced in a state that all petrologists would call magmatic, the nondetection of the gravity effect of a mafic residuum implies an initial metasomatic origin. Fortunately for magmatists, this is not necessarily true. Although the presence of the large amount of mafic material necessary for a descent by differentiation from a basaltic magma can almost certainly be excluded for orogenic granites, granitic magma could originate by differential melting within a horizon. Unless one postulates a granitic layer as the source, some sort of differentiation is necessary to give rise to granitic melt. As described previously, differential melting might be expected to occur within a horizon of great extent that is predetermined by its chemical composition. Fig. 16 illustrates that the granitic pore melt (sp.gr. = —0.10) could separate and then intrude as a diapir due to its high mobility and low density while it left behind a mafic residuum with a density surplus (+ 0.10). No room problem results,



Granitic melt (mass deficiency)

Mafic residuum (mass excess)

Fig. 16. Schematic illustration of how a granite diapir might form and be emplaced. (A) Pore melt (-0.10) forms and separates from crystalline mafic residuum (+ 0.10). (B) Pore melt begins to flow into dome analogous to a salt diapir. (C) Granite diapir grows until the flow is cut of by the development of rim syncline. No space problem is involved. Part of the country rock is domed and removed by erosion and part is thrust aside and sinks into the rim syncline. The diapir represents a concentrated mass deficiency and the plate of mafic residuum represents a dispersed mass surplus. The total mass is unchanged except for the eroded material.

but a very definite mass problem results. Figure 17 shows how the mass surplus could, in effect, go undetected if the residuum were in a thin, widespread plate. The small gravity increase over the edge of the thinner plate (Fig. 17) could go undetected or be misinterpreted at a distance from the granite, but, as the plate becomes thicker and narower, the gravity field deviates from the recognized shape. According

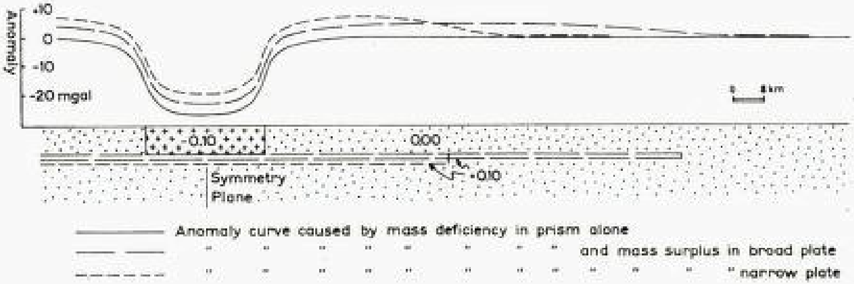


Fig. 17. Gravity profiles caused by the density configuration that resulted in the schematic drawing in Fig. 17. Mass deficiency in the prism is underlain by an equal mass excess in either plate. Mass excess in either plate exactly compensates mass deficiency of prism. As the plate becomes narrower and thicker, the theoretical gravity profile deviates from the gravity profile (solid line) commonly observed over granites. (Adapted with additions from Bott, 1956.)

to the shape of the gravity field, the residuum of the granitic melt cannot be concentrated; there is no advantage in forming a granite melt magmatically and then dispersing the residuum metasomatically.

The distinction between the granites of orogenic regions, the plutonic associations, and the granites of tensional regions, the volcanic associations, has been made by Kennedy and Anderson (1938). The great plutonic masses of granite occur together with volcanics in orogenic belts, and the large volumes of effusives occur together with intrusives of considerable size in fracture zones. Plutons and effusives may be found in both regions but the petrology differs as well as the tectonic setting. Although there is by no means agreement on this point (Buddington, 1959, p. 739—40), Raguin (1957, p. 186) regards the granites of volcanic provinces as special and also notes the distinction between orogenic granites and subvolcanic granites.

Gravity anomalies are significant in this connection; the gravity anomalies over orogenic granites do not appear to be influenced by mass surpluses. On the other hand, although the gravity profiles over granites of the Oslo province are similar in detail to those of orogenic regions, the whole region is the site of high positive anomalies. Barth (1952, p. 208) has remarked that the subordinate amount of known gabbroic rocks excludes a gabbroic mother magma for the Oslo igneous series unless there are colossal amounts of gabbro hidden under the present surface. Schwinner (1928) first noted that positive isostatic gravity anomalies occurred in the Oslo area. A Bouguer gravity map of the Oslo region (Norges geografiske oppmåling, Oslo) reveals positive gravity anomalies upon which are superposed the smaller "lows" of the granites. These positive anomalies cover much of the Oslo graben, which is a subvolcanic province occurring along a major fracture zone, and acquire additional significance because they occur over a subsided region where one would normally expect negative anomalies over a downfaulted block. Likewise, high positive anomalies have been reported by Cook and Murphy (1952) from the Slieve Gullion volcanic province. The conclusion that these positive anomalies are caused by a large mass surplus that is the residuum from the basaltic parent that gave rise to the igneous series is tempting.

Gravity profiles over orogenic granites do not show the effect of a compensating mass surplus. The gravity effect of the mass surplus must be diminished either by depth of burial or by dispersion which could either be primary and magmatic or be metasomatic. Granites of the volcanic association seem to be associated with large positive anomalies that suggest the presence of large amounts of dense material and a possible origin by magmatic differentiation.

Suggestions for Further Work

The following six projects would be worthwhile subjects for further study:

(1) The breccia zone and Bamble anomalies constitute the outstanding project for further study by means of more detailed gravity measurements and geology, numerous density profiles (Nettleton, 1939), and possibly seismic profiles.

(2) The Levang granite, a granite in which the regional fold pattern is preserved, has a small negative anomaly whose elucidation would be of great interest.

(3) The many small hyperite bodies would make an interesting study, but gravity work is complicated by their large terrain effects.

(4) Anomalies that occur along the NGO's profiles in Setesdal north of Evje and in Tørdal represent fluctuations of about 10 mgal.

(5) Detailed measurements around the Oddersjå granite (difficult terrain)and Evje amphibolite would be useful.

(6) A most important world-wide study of gravity anomalies and the volcanic association should receive attention.

Conclusions.

The breccia zone and Bamble anomalies.

Although concomitant deep causes are possible, the Bamble gravity gradient, which has its hinge at the breccia, is a shallow phenonemon, attributed to the denser Bamble rocks lying on a migmatitic substratum. The calculated throw of 500 m along the breccia at Hynnekleiv must be regarded as a minimum figure because the partially eroded surface rocks are the cause of the anomaly. Thickening of the Bamble supracrustal rocks over the migmatite front is postulated for the cause of the Bamble gravity gradient, but a mass surplus at moderate depth could contribute to part of the gradient. The proposed model is plausible in the vicinity of the breccia zone, but it becomes necessary to postulate either a greater density contrast or a subjacent mass of greater density near the coast.

Grimstad granite.

The Grimstad granite has an approximately cylindrical form and a northern contact that dips outward moderately to steeply. The "thickness" may range from 2.6 to 4 km and depends on the possible presence of lighter rocks, migmatites, at depth. The gravity effects of both models agree well with the measured profile. The granite appears to be strikingly uniform in density; a differentition pattern has been demonstrated by Heier and Taylor (1959). The form of the granite is compatible with a diapiric mode of intrusion and is particularly similar to diapir models intruded into a relatively strong overburden which may find its couterpart in the banded gneisses that surround the Grimstad granite. The structural relations do not seem to substantiate this. The granite is mainly crosscutting and does not appear to have greatly deformed the surrounding rocks so that stoping may have played an important role. Neither the displaced country rock nor mafic resistates of a granite magma were detected by the gravity survey. A very real room problem exists for the emplacement of this granite. The granite presents characteristics that are strongly magmatic; however, magmatic emplacement encounters difficulties regarding the missing country rock that are not easily circumvented.

Herefoss granite.

The Herefoss granite can be regarded as a funnel shaped intrusion whose greatest thickness occurs on the east side, and the gravity model agrees quite well with the field observations. The "thickness" of the granite depends on the density contrast assumed and may range from 2 to 5 km. The thickness of 5 km represents the maximum probable "thickness" of the granite based on a minimum expected density contrast. The density contrast between the west side of the granite and the granitic gneisses immediately adjacent is very small. These granitic Telemark gneisses are presumed to underlie the Bamble rocks around the granite east of the breccia. The true form and "thickness" of the granite probably lie somewhere between the two models presented.

The small negative anomalies over the northern part of the granite are attributed to the presence of large amounts of foreign material within the granite. This material occurs both as inclusions and as grey granite of higher than average density. A large pegmatite-containing inclusion was reflected in the gravity anomalies. An amphibolite layer may continue into the granite as a coherent body or as a series of large inclusions that are nearly in place. The north-central sector of the granite is calculated to contain about 77 per cent inclusions of average density (2.81).

The granite has strongly deformed the country rock and brecciated the contacts. The form, density contrast, and tectonic relations are compatible with a diapiric mode of emplacement accompanied by widespread chemical effects. The dominant process was probably diapiric intrusion concomitant with permissive emplacement and assimilation in the northern sector. Diapiric emplacement does not necessarily exclude the retention in place of some rock units. DeSitter (1956, p. 369) has described a gypsiferous marl diapir that broke through the weaker formations and left a competent limestone bed either partially interrupted or undisturbed. Much or all of the room problem for the Herefoss granite can be explained by combined processes of diapiric intrusion and replacement.

The density determinations provide important petrologic information. The two different levels exposed in the granite differ in two respects, mean density and heterogeniety. In the upper level on the east side, the effects of assimilation have not proceeded so far, and both inclusions and grey granite are common; however, on the west side, the inclusions have been almost completely assimilated, and the resulting denser granite has been homogenized.

None of the above-mentioned properties of the granite requires an igneous orgin; nevertheless, Heier and Taylor (1959) have found a differentiation process within the Herefoss granite. The development of the Herefoss granite seems to have been more complicated than that of the Grimstad granite, but the difference may only be apparent.

Comparison of Herefoss and Grimstad granites.

The relations of these two granites become more easily understood in light of the gravity model for the Bamble rocks (Fig. 9). The postulated wedge of Bamble rocks is thicker near the Grimstad granite (Fig. 18) The mean density of the granites which varies as Grimstad < east side of Herefoss granite < west side of Herefoss granite, is a function of level of exposure. The least dense Grimstad granite represents the highest level of exposure; the lightest fractions of the granites may have risen the most while inclusions have sunk to an intermediate level (E. side Herefoss) and have finally disappeared (W. side Herefoss). The west side of the Herefoss granite, however, may have been homogenized mechanically by "kneading" as it travelled a longer path into the western extension of the granite. These two granites provide an example of small vertical extent of Read's (1955) granite series.

The emplacement of the granites was conditioned by the strength properties of the surrounding rocks. The present section of the Herefoss

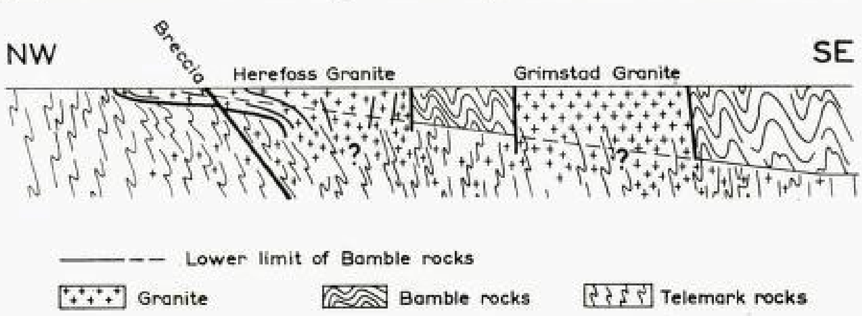


Fig. 18. Schematic cross section of the relations between the Herefoss and Grimstad granites, the «great friction breccia», the Bamble supracrustal rocks, and the Telemark granite gneisses. The difference in thickness between the two granites is related to the increasing thickness of the Bamble rocks toward the coast. The granites are conceived to be mobilized fractions of the migmatic Telemark substratum projecting through the overlying Bamble rocks. The granites represent two different levels of exposure.

granite is a deeper level surrounded by more plastic heterogeneous rock units, but the Grimstad granite is enclosed by rather homogeneous rigid banded gneisses.

At the breccia, the gravity effect of the Herefoss granite closely resembles that of the Telemark gneiss. The gravity anomalies picture the two granites as projections of the infrastructure (Wegmann, 1935), the Telemark granite gneiss and migmatites, through the transition zone, the Bamble rocks (Fig. 18). The "thickness" of the two granites would then be a function of the thickness of the wedge of Bamble rocks.

Evje amphibolite.

The Evje amphibolite has a positive anomaly of 12 mgal and a thickness of 1.25 km calculated from two profiles. Deviations between the calculated values and the measured curve are caused by the migmatitic, gradational nature of the contacts.

Oddersjå granite.

This syntectonic granite causes a 2-mgal negative anomaly. The calculated "thickness" is 1.4 km.

Density measurements.

Although density measurements are a necessary accompaniment to gravity measurements, they can also hold considerable petrologic significance. The demonstration of significant differences between the granites was achieved at a small expense of both time and money. The granites would have been studied with great difficulty by petrographic modal analysis. Density determinations are suggested as a means to obtain a quick survey of rock variability before chemical analyses are performed.

Diapirism.

A comparison between salt and granite tectonics provides useful information concerning granite emplacement. A diapiric mechanism of intrusion with concomitant deformation of the country rocks eliminates the room problem. Diapirs form in anticlines and particularly in axial culminations due to compressive stress and the high plasticity and low density of the diapiric material. By analogy with the behavior of salt beds, intrusion breccias and apophyses do not require the intervention of magma but only of a more plastic medium.

Gravity anomalies and the granite problem.

The characteristic feature of gravity profiles over granites is that neither the displaced country rock nor mafic differentiates exert any detectable influence on the profiles. While density configurations that compensate the displaced mass are possible, they are regarded as a special case. The more general case is that these displaced masses or mafic differentiates are dispersed so that they become part of the regional gravity pattern.

Metasomatic processes will explain the characteristic gravity profile, but the transfer must have taken place over rather large distances (kilometers). Uplift of the displaced country rock and subsequent dispersal by erosion are compatible with the gravity anomalies. Differential melting will satisfy the gravity anomalies if the melting took place over a thin widespread horizon. The pore melt must then separate from the residuum and flow to the point of intrusion, leaving the mafic residuum behind as a dispersed layer. A granitic or migmatitic basement (Wegmann, 1935) projecting through the overlying supracrustal rocks agrees with gravity anomalies.

Melting in a magma "chamber" and intrusion that leaves the dense residuum behind concentrated below the granite is incompatible with observed anomalies. Granites of the Oslo subvolcanic province, however, are surrounded by a positive regional gravity anomaly that parallels the Oslo graben. This may be the trace of an origin by fractional crystallization.

Little has been said concerning the metasomatic origin of granites in light of their gravity anomalies. As Bott (1956) has noted, there is nothing against a metasomatic origin in the typical gravity profile of a granite. Although complete dispersion suggests metasomatism, quantitative experimental confirmation of the exact process is generally lacking. The experimental work cited previously that simulates differential fusion, however, utilizes such powerful solvents as H_2O, HF, and HCl.In light of Orville's experiments (1960, 1961) with alkali metasomatism at 300° to 700° C, the inquiring reader may well wonder how these solvents remain passive in an orogenic environment until the rock can melt. Metasomatic processes cannot, at present, be ruled out, and they are in complete agreement with measured gravity profiles over granites if transfer has operated over rather long distances.

There is no "cure-all" for the granite problem. Gravimetric data can, nevertheless, set limits on the kind of a mass distribution that can exist and are just as necessary as petrographic data. The origin of each granite body must be judged on the basis of its field relations, petrochemistry, and gravity anomalies.

Appendix A.

Field procedure.

The Station Net

A pre-existing net, measured on precision-levelled elevations (Trovaag, 1956) by the Geographical Survey of Norway, consisted of 370 stations and was supplemented in places of special interest, particularly in and around the Herefoss and Grimstad granites. An additional 210 stations were measured by the Geographical Survey of Norway, and 207 stations were measured by the writer. Data concerning the Survey's stations can be obtained from its office in Oslo; the writer's data appear in Appendix B.

Originally, only the immediate area of the two granites was intended to be covered; but, as several features of interest came to light, the station net was extended to give a minimal coverage of the Bamble rocks, the Evje amphibolite, and the Oddersjå granite. As a result, the gravity stations established by Collette (1960) were added to the map in order to bring out the general trend of the isoanomaly lines off the coast.

Several of the gravity stations of the Geographical Survey of Norway were established as base stations in the vicinity of the two granites and connected directly by looping (Nettelton, 1940). These stations, in turn, are referred to the first order geodetic station in the Geologisk Museum, Oslo. These base stations were then occupied two times a day in order to correct for drift. For the regional work, however, it was usually necessary to go out from one of the Geographical Survey's stations and tie into another one. The maximum closure error in this case was 0.6 mgal, a value that is comparatively unimportant for the regional work.

Almost all of the gravity stations are located along roads since speed is a major consideration in establishing stations based on barometric elevations. The only exceptions were five stations, three of which were placed on islands and two stations on trigonometric points to which the party proceeded on foot. The stations are closer together over features of interest; *e.g.*, contacts, a fault, a large inclusion.

Elevations

Ordinarily the accuracy of the Bouguer anomalies is controlled by accuracy of the elevation determinations. In Norway, the accuracy is a function of both the elevation and the terrain effect.

Elevations were barometrically determined in most cases. The accuracy of a barometric elevation depends not only on the sensitivity and mechanical constancy of the barometer but also on humidity, temperature, weather conditions, and local variations in pressure. One Paulin and one Short & Mason barometer were used simultaneously; however, all elevations used were based on the Short & Mason barometer while the Paulin barometer was used as a check against "jumps". The humidity was disregarded as elevation differences were too small for this factor to cause significant errors. In order to correct for the effects of weather variations, the barometers were returned to a known elevation every hour, and the same elevation was used when practical. In some cases, the barometer traverses were as long as two hours; occasionally, the barometers had to be checked between two different known elevations.

Figure 19 is a histogram showing the distribution of closure errors (errors between 2 measurements on a known elevation) actually encountered in the field. The assumption was made that all closure errors greater than 1 m were caused by changes in barometric pressure and that this change was linear and unidirectional. If these assumptions are valid, the error due to changes in air pressure will be removed.

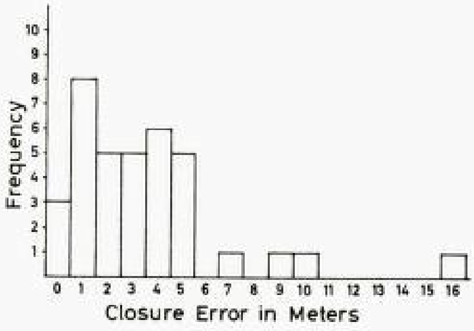


Fig. 19. Histogram showing the distribution of closure errors of barometric traverses for the elevation determinations.

The elevations of 26 stations have been checked against the levelled values for these stations. Although all the stations were not in the area studied in this paper, both the barometer and the technique employed were the same. If the levelled elevation is presumed to be the true value, the difference between the levelled elevation and the barometric elevation is the error. The mean of the error for 26 stations is + 0.6 m and the standard deviation is 1.96 m; therefore the barometric elevations are somewhat biased. Slightly over two thirds of the barometric station elevations should be within 2.6 m of the true value; a 2.6-m error in the station elevation corresponds to 0.5 mgal error in the Bouguer anomaly.

Terrain

The terrain is major factor in any gravity survey in Norway. Compared with most of Norway, the area studied has minor relief; abrupt differences in elevation of 200 m are found but those of 30 to 100 m are much more common. Unfortunately, most of the roads are in steep-walled valleys that follow structural lines of weakness. The gravity stations were located as far as possible away from topographic features. In some cases, however, undesirable locations adjacent to valley walls could not be avoided.

The Herefoss granite forms the highest topographic feature in the area and is so distinctive, in fact, that its eastern contact could be drawn on a topographic map. The Grimstad granite, on the contrary is characterized by nearly flat terrain.

Computations.

General Remarks

The observed gravity that is measured at each station cannot be used as such, but must be compared with the normal (theoretical) value of gravity. To accomplish this, certain reductions must be applied to the observed gravity in order to refer this value to a common reference plane (usually sea level). These reductions are the free-air correction; the Bouguer correction; the terrain correction, which is a special case of the Bouguer correction; and the isostatic correction. The difference between the reduced value of gravity and the theoretical value is the gravity anomaly. The free-air, Bouguer, and isostatic anomalies are obtained as the difference between observed and theoretical gravity after the free-air, Bouguer ,and isostatic corrections are applied respectively. The Bouguer anomaly is the anomaly that is used in this study because this anomaly best reflects the effects of near-surface structures.

Isostatic anomalies are used for studying deep-seated crustal structure. An isostatic correction, which is a function of the terrain elevation, is applied to the observed gravity at each station. Isostatic anomalies reveal deviations from isostatic (hydrostatic) equilibrium in the outer layers of the earth and are generally interpreted as regional flexures of the crust of the earth, which rests on a semiviscous medium. Shallow geologic features can cause isostatic anomalies, but Bouguer anomalies are better suited to the study of these features.

Theoretical Gravity

If the earth were a perfect sphere, the theoretical value of gravity would be everywhere the same at sea level. Because the earth is flattened at the poles, the theoretical value of gravity is about 5,000 mgal higher at the poles than at the equator. For this reason, the value of theoretical gravity is a function of latitude. Theoretical gravity, as computed from the International Gravity formula, gives the theoretical value of gravity for a given latitude on the normal earth spheroid. Each gravity station must be located accurately because, at the latitude of this survey, the value of theoretical gravity changes about 0.13 mgal for every 0.1 minute *(ca.* 180 m) of latitude. Since most of the stations are located by distinctive topographic or cultural features, they should be quite accurately located on the map, and any possible error depends on the map itself.

Observed Gravity

The value of observed gravity read in the field is only a dial reading. This reading is referred to the dial reading at a base station that has a known value of observed gravity; the dial constant times the difference of the readings gives the difference in observed gravity between the two stations. Worden gravity meters no. 178 and no. 135 with dial constants of 0.1086 and 0.0902 mgal per division were used by the Geographical Survey and the writer respectively. The gravity meter readings were corrected for drift under the assumption that drift was linear. Since the observed values of gravity vary not only with geologic structures but also with elevation, they must be corrected to sea level by applying the free-air and Bouguer corrections.

Free-air and Bouguer Corrections

Newton's formula for gravitational attraction shows that the force of attraction between two bodies (here the earth and a small mass in the gravity meter) varies inversely as the square of the distance between them. Unless gravity was observed at sea level, and it seldom is, its value will be either increased or diminished depending on whether the station was located above or below sea level. Because this correction is only a function of the distance between the station and sea level (it neglects the attraction of the intermediate material), it is called the free-air correction. Since the observed value will be either too small or too large if measured above or below sea level, the free-air correction can be positive (the usual case) or negative (commonly only at sea). The correction can be expressed as a simple constant, 0.3086 mgal/m.

In making the free-air correction, no account was taken of the attraction caused by the material between sea level and the station elevation. This material is considered to be an infinite slab of appropriate density to similate the actual rock density between station elevation and sea level. This correction, called the Bouguer correction after the French geodeticist who first applied it, is expressed by the constant 0.1118 mgal/m for a Bouguer density (density of the abovementioned infinite slab) of 2.67 gm/cm^3. This density is the one that is normally used for areas composed of crystalline rocks and allows maps from adjacent areas to be compared. Any large deviation from this assumed density could cause significant errors in the Bouguer anomalies, particularly if the station was located high above sea level.

The free-air and Bouguer anomalies are both simple constants, multiplied by the elevation; therefore, they can be combined into one constant. The free-air and Bouguer corrections are always opposite in sign because the free-air reduction corrects for decreased (for elevations above sea level) gravity due to greater distance between the center of the earth and gravity meter, and the Bouguer correction corrects for increased gravity due to the attraction of the slab of material between sea level and the gravity meter. A combined free-air-Bouguer constant of $+$ 0.1968 mgal/m was used.

Terrain Correction

The terrain correction is really a special case of the Bouguer correction that is usually not applied unless absolutely necessary. In making the Bouguer correction, the material between the station elevation and sea level was assumed to be an infinite plate; *i.e.*, the station was located on a high plain or plateau. In practice this is seldom the case; hills project above the station and valleys descend below it. Both the hills and valleys represent deviations from the flat plate and cause the Bouguer gravity anomaly to be too low; therefore, the terrain correction is always positive and is added to the Bouguer anomaly.

The terrain correction can be insignificant in lowlands, but, in Norway, the terrain effect can be much greater than the anomalies due to geological features. A Bouguer anomaly map of Norway would show gravity "troughs" or "lows" over all the large valleys and fjords, along which the major roads run, while the intervening divides and plateaus would be the site of gravity "highs". The terrain effect can be greatly diminished by locating the gravity stations as far as possible from abrupt relief. A glance at the charts of Hubbert (1948a) will show how rapidly the terrain effect decreases as the station is moved away from a topographic feature.

The usual practice in calculating the terrain correction is to use the method of Hammer (1939) which utilizes a circular template subdivided into compartments. This template is placed over the station on a topographic map and the elevation difference between the station and each compartment are estimated. The terrain correction for each compartment in hundredths of a mgal can then be looked up in a table. This is a laborious process when a large number of stations require terrain corrections and is poorly suited to abrupt changes in relief.

Hubbert (1948 a) has published diagrams for estimating the corrections of 2-dimensional topographic features; *i.e.*, features which are considered to extend infinitely in a direction perpendicular to their two smallest dimensions. In practice, this condition is approximately fulfilled when the length of a topographic feature is 5 times its width and the gravity station is located near the middle of the feature. These diagrams are admirably suited to Norwegian topography which is typified by steep, narrow valleys dissecting a rather flat upland. Naturally, they can only approximate the true terrain, but they reduce the error caused by topography to a value that can be tolerated within this survey.

Terrain corrections were applied to most of the stations using the method of Hubbert *(ibid.)*. As a check, the terrain corrections of five stations located near varying types of topography were computed using the methods of both Hammer and Hubbert. A comparison of the determined values appears in Table 1. These values do not differ greatly so that the topographically corrected Bouguer anomalies should not include any important errors from the terrain effect.

Table I.

Comparisons of terrain corrections.

Station	Terrain correction by 3-dimensional (Hammer's method)	Terrain correction by 2-dimensional (Hubbert's method)
E39T53	2.88 mgal	2.4 mgal
E39N46	0.72	0.3
S34	0.61	0.4
S68	0.88	0.8
S79	0.28	0.3

Accuracy

The accuracy of the Bouguer anomaly values depends on additive error from many sources. Although a quantitative estimate of the total error is desirable, the data are statistically inadequate and the final evaluation of error will be rather subjective. The largest single source of error is probably the terrain effect, but this should not exceed 0.5 mgal at its greatest. Most of the elevations are probably within 2.6 m of the true value, but some are in greater error than this. An error of 2.6 m in elevation gives an error of 0.5 mgal (combined free-air-Bouguer correction is 0.1968 mgal/m). The expected error in the theoretical gravity and the observed gravity should not exceed 0.1 mgal for each. The maximum total error of the Bouguer anomalies from all sources is slightly over 1 mgal; therefore, a contour interval of 2 mgal was used for the Bouguer and residual anomaly maps.

An error in the Bouguer anomalies may be caused by inappropriate choice of Bouguer density; a value of 2.67 gm/cm^3 was used for Bouguer density. If the true value of mean density for the surface rocks differed greatly from this and if the elevation was high, an appreciable error would be introduced into the Bouguer anomalies. For every difference of 0.10 gm/cm^3 from 2.67 gm/cm^3 in the Bouguer density together with an elevation difference of 250 m, the Bouguer anomaly will be incorrect by about 1 mgal. A density of 2.77 gm/cm^3 more closely approximates the mean density of the gneisses; but, since the highest elevations generally occur in the Herefoss granite and the mean density of the granite is about 2.67 gm/cm^3, a value of 2.67 gm/cm^3 is the most realistic one to use. The relative error between stations caused by deviations from the Bouguer density used is inconsiderable compared with the error introduced by the elevation determination and the terrain effect.

Densities

A most important factor for gravity interpretation is the mean density of the various rock units. For this study, rock samples were taken at or near the gravity stations. The extremely heterogeneous nature of the gneisses should be emphasized so that sampling is a major problem. The mean densities were computed for the Herefoss granite on both sides of the "great friction breccia" because the granite on different sides is somewhat different petrographically (Elders, oral communication).

The density of the samples, which were of hand-specimen size, was determined by the water displacement method. The precision of the method had a standard deviation of 0.014 gm/cm^3. The rocks were assumed to be very nearly impermeable since specimens weighed after being submerged in water under a vacuum exhibited a density only

Table 2.

Rock Densities.

Rock Unit	No. of Samples	Mean Density	Standard Deviation	Range of Values
Bamble rocks	54	2.81_4	0.140	2.61—3.08
Telemark rocks	36	2.76_9	0.110*	2.59—3.06
Grimstad Granite	20	2.64_5	0.018*	2.61—2.68
Herefoss Granite east of breccia	40	2.68_0	0.054*	2.58—2.80
Herefoss Granite west of breccia	18	2.70_8	0.033*	2.64—2.78
Evje Amphibolite	5	3.03	0.131*	2.86—3.17
Oddersjå Granite	7	2.73	0.050*	2.67—2.80
Gneiss surrounding the Oddersjå Granite	6	2.83	0.155*	2.69—3.13

* corrected standard deviation

about 0.01 gm/cm^3 greater. The mean densities for the different rock units appear in Table 2.

The density determinations reveal information of petrological significance. The mean density of the Bamble rocks is greater than that of the Telemark rocks at the 0.05 level of significance. The Grimstad granite, the east side of the Herefoss granite, and the west side of the Herefoss granite differ in mean densities and dispersion at the 0.05 level of significance. We can, thus, state that the western part of the Herefoss granite is denser than the eastern part, and the eastern part of the Herefoss granite is denser than the Grimstad granite. In addition, the eastern part has a greater dispersion (standard deviation) that either the western part or the Grimstad granite. The differences in both dispersion and in mean density must be related to the genesis of the granites.

Interpretation of gravity data.

The interpretation of gravity data can be separated into four phases: (1) The isolation of the anomalies. (2) Consideration of the most likely cause (or causes) of the anomalies in terms of the known geological and/or geophysical data. (3) Computation of the model (or models) that are stipulated by the above data. (4) Integration of the final 3-dimensional interpretation that takes account of the gravity model, the geology, and any other possible geophysical data.

Gravity anomalies are caused by horizontal variations in density. In other words, a non-horizontal density discontinuity, which could be caused by folding, faulting, or plutonism, must occur in order to cause an anomaly; a succession of horizontal layers, all of different density, will not cause a gravity anomaly. Likewise, movements, and plutonism which result in no lateral density contrast will not cause an anomaly. Gravity anomalies will be produced by folding, faulting, and plutonism involving rocks of different density, but may be too small to be detected or separated from the "background".

The gravity anomalies of any area are composed of the interaction of the attractions from many different sources. Generally speaking, the broad regional trends are caused by deep-seated structure; *e.g.*, inclined crustal layers or warped crustal layers, and the local, more restricted anomalies are caused by surface or near-surface features; *e.g.*, faults, folds, and plutonic bodies. The terms "local" and "regional", however, are relative so that these two generalizations must not be taken too literally. The "regional" anomaly could be caused by a continent and the "local" anomaly by a basin, or the "regional" anomaly could be caused by a basin and the "local" anomaly by an anticline. In order to determine the nature of the geologic features, the gravity anomalies must be separated into their component parts.

The simplest method is to construct a typical profile through an area of interest and draw the smoothed-out line that represents the regional gradient (see Dobrin, 1960, p. 245). Then, the local anomaly at any particular point is the vertical difference between the smoothed-out line and the measured profile. This method presupposes that the regional anomalies are carried out far enough on either side of the local anomaly to give a good indication of the regional trend. This method is best applied where the trends are simple and the number of gravity stations is limited.

A more elegant solution is to make a residual anomaly map, which is a map of the local anomalies after the regional effect is removed *(ibid.*, p. 242—45). Either graphical or computational methods are applied to draw smoothed contours which represent the "regional" effect. The "regional" Bouguer anomalies are then substracted from the measured Bouguer anomalies to reveal the residual Bouguer anomalies. Figures 3 and 4 are examples of regional and residual Bouguer anomaly maps in the vicinity of the Herefoss and Grimstad granites. If these two maps were superposed, added together, and recontoured, the resulting map would be exactly the same as the original Bouguer anomaly map, Plate 1.

The sign of local anomalies is only a relative property. If the anomaly over a granite is less than that of its surroundings, the anomaly is considered negative even though it may have a positive sign.

Once the local anomalies are resolved, a model whose gravity field satisfies the given anomaly is calculated. Unfortunately, there are an infinite number of solutions that will satisfy the given anomaly (Nettleton, 1940, p. 120). In order to calculate the "most likely" solution, something must be known about the density distribution; in other words, supplementary geological and/or geophysical information is necessary. Obviously, the situation is considerably simplified if features that crop out are studied. Then the positions and possibly the attitudes of geological contacts are usually known, and the densities of the surface rocks can be estimated. Because of the inherent lack of an unique solution to gravity problems, the geological data are used to set certain limits for the parameters of the solution so that a "most probable" model or models can be computed.

The actual computations are rather simple, but, as they involve "cut and try" methods, they can become time-consuming. Many geologic features can be approximated by simple models, such as a horizontal cylinder for an anticline, a vertical cylinder for a salt plug or granite stock, and a horizontal step for a fault (Nettleton, 1942). The attraction of these models at different points can be calculated and compared with the respective gravity anomaly. For elongate bodies of irregular profile, a graticule (Jung, 1927; Hubbert, 1948b) can be applied and gives a quick solution graphically; the simplifying assumption is made that the bodies are 2-dimensional; *i.e.*, that they are infinitely long in a direction perpendicular to the two shorter dimensions. If the length of a form is four to five times its width or greater, the assumption of infinite length introduces only a relatively small error but simplifies computations considerably. By one of the above-mentioned methods, a model whose gravity field satisfies both the given anomaly and the restrictions imposed by geological data is computed, and this model represents the best information that can be obtained from the gravity survey.

Until now, the geologist has been restricted to the extrapolation of subjective profiles from his 2-dimensional map. By comparing surface geology with the gravimetrically determined model and re-evaluating the one in the light of the other, an actual 3-dimensional interpretation becomes feasible. To be sure, some factors will always be in doubt, but more information will have been gained than would otherwise be possible.

Interpretation of regional anomalies.

General Remarks

Regional anomalies of great extent may be caused by near surface geological features, but are usually attributed to deep structure. A large negative anomaly in India was interpreted by Glennie (1932) as a basin-like downwarp in the "granitic" and "basaltic" layers of the crust. Ansel (1937) argued that the same anomaly could be caused by a change in thickness of the "granitic" layer; i.e., a downwarp of the interface between the two layers without a corresponding flexure at the base of the "basaltic" layer. More recently, Garland (1950) proposed that gravity anomalies in the Canadian Shield are principally caused by thickness variations in the "granitic" layer, and Hinze (1959) explains the negative anomaly over the Baraboo syncline as a thickening of the granitic layer. Undulations in the crustal layers have become a widely applied concept for interpreting large-scale anomalies.

In the process of interpretation, a gravity anomaly is always compared with the known geology. If an anomaly of large extent cuts across all geologic contacts or cannot be correlated with the density distribution of the surface rocks, it is customary to attribute the anomaly to "deep" structure. In the absence of a likely feature such as a concealed granite batholith or mafic intrusion of large extent, undulations or thickening and thinning in the "granitic" and "basaltic" layers are invoked to explain these anomalies. Since a given gravity anomaly can be explained by an infinite number of density distributions, the only restriction is that a given gradient of the gravity field limits the depth and density contrast of the structure causing it. With only this one restriction, the conclusion is obvious that regional anomalies find a ready explanation as the effect of undulations in crustal layering. Seismic data are, therefore, necassary to obtain additional information that makes the interpretation less speculative.

Nature of Crustal Layering

The concept of crustal layering is the direct result of seismic studies. By analyzing earthquake seismograms, a P-wave velocity discontinuity of 8 km/sec was discovered by Mohorovičic in 1910. This discontinuity, which bears the discoverer's name and is commonly shortened to Moho, has more recently been identified as the transition between the earth's crust and the mantle. In 1925, Conrad discovered a velocity discontinuity of 5.6 km/sec within the crust itself. From these data, has evolved the concept of crustal layering in which a layer with a velocity of 5.6 km/sec overlies a layer with a velocity of 6.5 km/sec. They are underlain by a medium with a velocity of 8.0 km/sec.

These layers have become identified with rocks of specific compositions because these rocks have elastic properties that satisfy the seismically determined velocities. The designations of granitic and basaltic have been assigned to the layers with velocities of 5.6 and 6.5 km/sec respectively while peridotic or eclogitic material was proposed for the upper mantle with a velocity of 8.0 km/sec. The velocities in crustal layers were later revised upward to agree with observed velocities from explosions (Gutenberg, 1955, p. 211) to about 6.0 km/sec for P waves near the surface. The choice of a rock composition, thus, depends on our knowledge of the propagation velocities of rocks at various pressures.

The velocities of compressional waves (P) in a large number of rocks have recently been determined experimentally by Birch (1960). The velocities of such rocks as gneiss (felsic), quartzite, serpentinite, greywacke, charnockite, slate, and granodiorite gneiss may all lie within the range of granite velocities at moderate pressures. The densities of these same rocks range from 2.601 to 2.758 gm/cm^3. The velocities of the various granites measured range from 5.97 to 6.46 km/sec at 2000 bars pressure. On the other hand, more dense, mafic rocks such as anorthosite, chlorite schist, norite, gabbro, and amphibolite have velocities that are distinctly higher, 6.82 km/sec and above at 2000 bars pressure. Since seismic crustal velocities are measured at distances from 10 to over several hundred km, the velocity obtained is, of course, the average velocity over a rather long path. Birch (1955, p. 103) has commented on the probable heterogeneous nature of the upper crust and Tatel and Tuve (1955) emphasize regional variations in crustal density. In light of these data, the "granitic layer" can be pictured as a mosaic of different rocks similar to the Precambrian shields in which the role of felsic (sialic) rocks greatly predominates over that of mafic (simatic) rocks.

Seismic data indicate the preponderance of sialic material in the upper part of the crust. Although there is little in the seismic information to suggest that this is a pristine granitic layer that forms a universal source of granitic magma, the source of granite can certainly be within this sialic material.

While geophysicists have been clear over the heterogeneous nature of the "granitic layer" (Gutenberg, 1951, p. 410), the problem of crustal layering is considerably more confused. In recent years, considerable doubt concerning the validity of the "granitic" and "basaltic" subdivisons in the crust has arisen. This crustal model was determined from earthquake seismograms. The study of seismic waves from explosions, which allows better control of conditions and eliminates the unknowns of hypocenter and origin time, has suggested an alternate interpretation of crustal structure. Willmore (1949) failed to definitely detect the "basaltic layer" in northern Europe. Tatel and Tuve (1955) did not detect this layer in a number of locations in the U.S. Worzel and Shurbet (1955) chose to eliminate a two-layed crustal model from their standard crustal sections. Wollard (1959, p. 1540) suggests that the top of the "basaltic layer" may lack a sharp acoustic boundary and cannot be resolved even though gravity data imply its existence.

Mean densities of 2.84 to 2.88 gm/cm^3, based on gravity considerations, are proposed for the standard crustal section (Worzel and Shurbet, 1955; Woollard, 1959). This mean density implies the presence of rocks more dense than the sialic rocks of the upper crust; *i.e.*, the basaltic rocks whose measured velocities correspond to that measured near the base of the crust. The paradox which occurs in trying to reconcile gravity data and recent seismic investigations is further complicated because the velocity of sialic rocks at high pressure approaches 7.0 km/sec, the velocity near the base of the crust. Bullard (1954, p. 61—71)has written a particularly lucid discussion about the problems involved in determining crustal structure seismically and prefers to omit thicknesses of the "granitic" and "basaltic layers" in his crustal sections. The existence and nature of crustal layering are questions that remain uncertain and require considerable clarification in the future.

The base of the crust or "Moho" is, in contrast to internal crustal structure, marked by a sharp acoustic discontinuity that has been universally detected. It occurs at a depth of about 30 km under the continents and may be over 60 km deep under young mountain ranges. Crustal thicknesses are affected only slightly by the addition or omission of the "basaltic layer". The Moho is only about 10 km below sea level in the ocean basins. Large-scale undulations are found in the base of the crust.

What then, is the role of crustal and subcrustal structure in gravity interpretation? The literal interpretation of widespread gravity anomalies that are unrelated to local geology as flexures in the "basaltic layer" has little to offer because the crustal layering concept, itself, appears to be in a state of flux. A better solution might be to calculate a geologically possible model or series of models that satisfy the given conditions without attempting to fit the solution into a "basaltic layer" concept. Undulations of the base of the crust are necessary to explain many anomalies of large extent. In this case, a density difference of *ca.* 0.43 gm/cm^3 is used. Until more data become available, attributing gravity anomalies to flexures in the "basaltic layer" is not particularly useful; therefore, another solution could well be sought.

Appendix B
Station Data.

Station Number	Latitude N. ° ′	Longitude W. of Oslo ° ′	Elevation Meters	Observed Gravity Milligals	Combined Free Air & Bouguer Corrections Milligals	Bouguer Gravity Anomaly Milligals
S 1	58 23.3	2 02.1	9.23	981 803.7	+ 1.8	+ 9.7
2	23.7	01.6	7	809.9	1.4	+ 15.1
3	22.8	02.4	37	794.3	7.3	+ 6.5
4	22.2	05.2	57	785.8	11.2	+ 2.9
5	22.1	06.2	36.96	788.7	7.3	+ 1.9
6	22.7	04.8	29	793.1	5.9	+ 4.1
7	23.5	06.0	49	790.4	9.8	+ 4.4
8	23.2	04.2	18.2	795.5	3.8	+ 3.7
9	23.8	04.0	40	796.3	8.1	+ 8.0
10	24.4	04.7	35.70	800.9	7.0	+ 10.8
11	25.3	05.9	55.3	796.8	10.3	+ 9.8
12	24.3	05.9	52	794.1	10.8	+ 8.0
13	23.9	07.4	50	792.2	10.4	+ 6.2
14	22.7	07.2	85	779.1	16.9	+ 1.2
15	22.1	08.6	51	786.4	10.4	+ 2.9
16	22.2	07.4	47.6	785.3	9.7	+ 0.8
17	20.4	07.5	13.80	795.5	3.0	+ 7.2
18	14.9	20.3	21.44	778.4	4.2	— 1.7
19	15.4	21.2	39	771.8	7.7	— 5.5
20	15.9	22.8	45.69	772.0	9.0	— 4.7
21	16.2	23.5	42	771.9	8.4	— 5.7
22	16.9	25.4	49	771.4	9.9	— 5.8
23	17.2	25.8	51	771.6	10.6	— 5.3
24	17.9	26.5	60	769.1	12.6	— 6.6
25	18.2	27.0	65	769.6	13.6	— 5.6
26	18.4	27.9	69.61	770.0	14.1	— 4.8
27	19.1	28.0	71	770.0	14.4	— 5.6
28	16.2	31.3	109	762.0	22.5	— 2.6
29	15.9	30.4	137	756.5	27.2	— 2.2
30	14.8	29.9	87	765.1	17.5	— 1.0
31	13.8	28.0	47	771.9	9.3	— 0.2
32	12.8	25.8	1.1	780.5	0.3	— 0.7
33	21.2	10.2	45.79	788.4	9.0	+ 4.6
34	23.7	13.2	78	776.9	15.8	— 3.5
35	23.2	14.4	90	770.5	19.4	— 5.6
36	23.7	15.7	147	760.3	29.3	— 6.6
37	24.1	17.3	158	756.8	31.5	— 8.4

Station Number	Latitude N. ° ′	Longitude W. of Oslo ° ′	Elevation Meters	Observed Gravity Milligals	Combined Free Air & Bouguer Corrections Milligals	Bouguer Gravity Anomaly Milligals
S 38	58 24.4	2 19.0	155	981 758.9	+ 30.6	— 7.6
39	24.6	19.5	157	760.0	31.1	— 6.2
40	25.5	20.6	185	757.0	36.6	— 5.0
41	26.0	21.7	189	757.1	37.6	— 4.6
42	26.4	22.5	187	759.2	37.2	— 3.4
43	26.2	23.4	207	753.8	41.1	— 4.7
44	27.0	23.1	179	760.8	36.9	— 3.0
45	27.4	21.9	205	757.0	41.0	— 3.2
46	27.3	22.5	151	765.7	31.6	— 3.7
47	27.6	23.0	100	775.6	21.9	— 1.1
48	26.9	23.7	80	780.0	17.8	— 2.8
49	25.2	25.9	76	774.8	17.4	— 6.1
50	25.7	26.3	117	768.3	25.1	— 5.5
51	26.6	26.5	159	760.7	33.3	— 6.0
52	27.1	26.6	161	761.0	33.7	— 6.0
53	28.3	26.8	231	750.4	46.5	— 5.4
54	28.8	27.0	221	751.6	45.3	— 6.1
55	29.3	27.3	220	753.4	45.1	— 5.1
56	30.4	28.2	196	759.4	39.4	— 6.4
57	30.5	28.5	187	759.7	37.6	— 8.0
58	30.8	29.0	187	759.8	37.3	— 8.6
59	31.3	29.0	172.25	763.4	34.2	— 8.8
60	30.1	30.8	224	752.3	44.8	— 7.7
61	29.8	24.9	281	744.0	55.8	— 4.6
62	30.1	26.2	289	742.1	57.3	— 5.3
63	28.4	29.4	287	740.1	57.1	— 5.2
64	28.5	28.0	315	734.0	62.6	— 6.0
65	27.6	30.2	322	731.9	64.0	— 5.5
66	27.1	30.1	336	728.8	66.6	— 5.4
67	27.2	28.2	272	742.1	54.0	— 4.8
68	27.0	18.0	175	763.8	35.3	— 1.5
69	26.3	28.6	174	761.0	35.1	— 3.6
70	25.6	29.7	166	760.4	33.3	— 5.0
71	23.2	16.5	92	769.2	18.7	— 7.6
72	22.7	19.5	145	757.3	29.1	— 8.5
73	22.1	21.8	190	748.7	37.8	— 7.5
74	21.4	23.5	177	749.3	35.1	— 8.6
75	21.1	23.7	177	748.9	35.5	— 8.3
76	20.6	24.6	166	751.4	33.0	— 7.6
77	20.6	25.9	196	747.3	38.8	— 5.9

Station Number	Latitude N. ° ′	Longitude W. of Oslo ° ′	Elevation Meters	Observed Gravity Milligals	Combined Free Air & Bouguer Corrections Milligals	Bouguer Gravity Anomaly Milligals
S 78	58 20.8	2 27.5	135	981 758.9	+ 26.9	— 6.5
79	22.6	24.6	214	744.8	42.5	— 7.5
80	21.7	25.9	214	744.4	42.4	— 6.8
81	20.3	29.3	51	773.6	10.6	— 7.4
82	20.0	26.1	180	749.9	35.9	— 5.4
83	22.7	31.3	236	745.5	46.8	— 2.5
84	22.8	32.0	251	741.8	49.6	— 3.5
85	24.2	31.9	250	743.2	49.8	— 3.8
86	24.0	31.6	263	740.9	52.0	— 3.7
87	22.2	29.7	178	753.4	35.5	— 5.2
88	22.4	28.3	192	750.9	38.5	— 5.1
89	21.7	28.6	128	761.3	26.1	— 6.0
90	21.3	28.2	48	774.0	11.1	— 7.9
91	20.7	19.9	197	747.6	38.9	— 5.6
92	21.2	17.6	109	764.5	22.6	+ 5.7
93	21.0	16.3	52	774.9	12.0	+ 5.6
94	21.0	07.3	20	792.5	3.9	+ 3.9
95	21.7	06.5	1.1	795.6	0.2	+ 2.3
96	24.1	01.4	4	813.8	0.8	+ 17.9
97	23.7	01.0	0.6	813.5	0.1	+ 17.4
98	08.9	42.8	7.34	775.3	1.4	— 0.4
99	08.4	44.6	22	771.3	5.2	+ 1.0
100	08.2	45.3	6	774.5	1.7	+ 0.9
101	09.3	53.8	17	770.6	3.8	— 2.4
102	09.7	55.7	88	756.6	18.5	— 2.2
103	08.6	52.1	17	769.8	4.1	— 1.9
104	08.6	51.0	46	764.9	10.6	— 0.3
105	08.6	48.7	89	757.5	18.5	+ 0.2
106	08.5	46.7	65	764.2	14.0	+ 2.6
107	09.5	41.9	7	774.4	1.4	— 1.2
108	10.8	41.6	7	775.7	2.2	— 0.9
109	10.9	42.4	8	776.0	2.0	— 0.9
110	11.2	42.8	20	773.8	4.4	— 1.1
111	11.6	43.0	27	772.4	5.9	— 1.6
112	11.9	43.7	144	749.7	29.1	— 1.5
113	12.1	44.4	130	751.0	26.4	— 3.2
114	12.3	44.9	119	752.8	24.3	— 3.7
115	12.4	45.7	119	753.4	24.6	— 2.9
116	12.5	46.3	119	754.0	24.2	— 2.8
117	12.8	47.1	15	772.9	6.5	— 2.1

Station Number	Latitude N. ° ′	Longitude W. of Oslo ° ′	Elevation Meters	Observed Gravity Milligals	Combined Free Air & Bouguer Corrections Milligals	Bouguer Gravity Anomaly Milligals
S 118	58 12.8	2 47.6	7.4	981 774.7	+ 4.7	— 2.1
119	13.5	48.5	51	769.0	10.9	— 2.5
120	13.9	49.7	118	755.7	24.4	— 2.8
121	34.7	34.4	210.67	759.3	42.2	— 9.4
122	35.1	38.0	255	749.2	52.2	— 10.0
123	34.7	41.5	322	735.0	64.9	— 11.0
124	34.7	42.6	330	733.1	66.3	— 11.5
125	34.7	43.7	338	730.8	67.7	— 12.4
126	34.6	46.3	385	722.6	76.8	— 11.4
127	34.6	48.1	403	723.6	80.4	— 6.9
128	34.6	49.0	403	725.7	79.9	— 5.2
129	34.8	50.3	403	727.6	79.8	— 3.7
130	34.5	52.6	246	756.1	49.9	— 4.7
131	35.1	54.9	184.65	765.5	37.6	— 8.5
132	33.8	56.7	173	765.7	34.9	— 9.2
133	34.4	2 58.1	185	762.0	38.8	— 9.8
134	35.3	3 00.6	199	758.2	43.7	— 9.8
135	35.1	3 01.7	180	761.6	38.9	— 11.1
136	34.5	2 59.5	177	761.9	37.6	— 11.2
137	31.4	53.0	340	739.0	68.3	+ 0.7
138	30.5	47.8	389	729.6	77.6	+ 1.9
139	29.1	47.7	324	738.4	64.6	— 0.6
140	27.4	48.2	208	751.9	42.2	— 7.0
141	27.3	49.9	231	748.8	46.7	— 5.5
142	27.4	51.2	238	745.2	47.7	— 8.3
143	23.9	48.3	135	763.2	27.6	— 5.7
144	23.2	49.5	124.9	764.2	26.2	— 5.1
145	25.5	47.8	183	755.4	37.2	— 6.0
146	25.4	46.3	263	738.3	52.4	— 7.8
147	25.5	45.2	253	739.5	50.4	— 8.7
148	26.0	42.7	206	751.8	41.2	— 6.2
149	25.2	41.0	199	751.9	39.6	— 6.7
150	24.6	38.0	197.05	751.5	39.0	— 6.9
151	26.6	36.4	202.65	751.8	40.5	— 7.8
152	30.5	09.5	75	791.6	16.0	+ 2.2
153	31.5	16.3	243	758.8	48.9	+ 1.0
154	33.5	16.5	272	756.1	54.0	+ 0.8
155	32.7	17.0	294	751.4	57.9	+ 1.1
156	32.7	17.7	272	754.7	54.0	+ 0.4
157	31.1	17.5	306	746.7	60.6	+ 1.2

Station Number	Latitude N. ° ′	Longitude W. of Oslo ° ′	Elevation Meters	Observed Gravity Milligals	Combined Free Air & Bouguer Corrections Milligals	Bouguer Gravity Anomaly Milligals
S 158	58 31.0	2 18.9	293	981 746.8	+ 58.0	— 1.8
159	31.0	19.8	266	750.5	52.9	— 3.6
160	30.9	21.4	179	763.3	36.6	— 6.0
161	31.4	22.1	80	781.5	18.3	— 6.8
E39T.23	32.1	22.0	106.61	777.4	22.2	— 7.8
162	36.0	17.0	165	773.2	34.5	— 5.0
163	36.5	22.0	176	771.6	35.1	— 6.7
164	35.5	22.5	151	774.1	30.6	— 7.2
165	35.1	19.0	177	768.7	35.9	— 6.9
166	33.5	20.7	109	779.1	22.4	— 7.9
167	30.1	2 22.7	86	778.1	19.7	— 6.9
168	58.4	1 03.5	0.1	892.2	0.0	+ 49.5
169	57.3	05.8	0.1	891.0	0.0	+ 49.9
170	56.6	07.7	0.1	888.7	0.0	+ 48.5
171	55.8	09.4	1.6	885.8	0.3	+ 47.0
172	57.4	26.3	87	843.6	17.1	+ 19.4
173	58.1	27.4	66	844.3	13.0	+ 15.0
174	58.8	28.2	66.5	837.5	13.1	+ 7.4
175	58.7	30.0	61.5	833.2	12.1	+ 2.3
176	52.1	27.5	81.5	849.0	16.0	+ 30.8
177	50.9	27.9	77.5	849.0	15.2	+ 31.6
178	49.7	26.9	0.4	863.3	0.1	+ 32.3
179	49.7	25.1	0.1	866.5	0.0	+ 35.5
180	50.7	21.5	0.0	871.4	0.0	+ 39.0
181	51.0	18.8	1.2	875.5	0.2	+ 42.9
182	50.6	18.1	54	863.7	10.6	+ 42.1
183	49.6	19.5	42	866.7	8.3	+ 44.1
184	49.0	20.4	26	868.5	5.1	+ 43.5
185	47.8	20.8	1.4	874.6	0.3	+ 46.4
186	48.0	17.8	1.0	878.4	0.2	+ 49.8
187	47.6	18.5	0.4	877.3	0.1	+ 49.1
188	52.7	32.2	121	835.8	23.8	+ 24.6
189	48.8	38.6	105	837.2	20.7	+ 28.0
190	46.7	37.9	10	857.0	2.0	+ 31.9
191	45.8	38.4	12	857.6	2.4	+ 34.2
192	42.9	36.0	1.0	860.1	0.2	+ 38.4
193	43.0	34.9	0.3	861.6	0.1	+ 39.5
194	43.1	31.2	0.8	865.2	0.2	+ 43.2
195	43.2	29.0	1.3	867.5	0.3	+ 45.4
196	41.8	36.8	2.0	858.4	0.4	+ 38.3

Station Number	Latitude N. ° ′	Longitude W. of Oslo ° ′	Elevation Meters	Observed Gravity Milligals	Combined Free Air & Bouguer Corrections Milligals	Bouguer Gravity Anomaly Milligals
S 197	58 41.4	1 34.5	1.0	981 862.7	+ 0.2	+ 43.0
198	41.5	32.6	0.2	866.3	0.0	+ 46.2
199	41.5	30.5	2.0	868.8	0.4	+ 49.1
200	41.0	38.9	1.2	858.6	0.2	+ 39.4
201	43.1	37.5	25.5	853.7	5.0	+ 36.5
202	44.3	43.4	121	832.5	23.8	+ 32.4
203	45.1	44.4	137	816.3	27.0	+ 18.4
204	45.2	48.6	178	808.2	35.0	+ 18.2
205	46.8	51.2	190	796.9	37.4	+ 7.2
206	47.9	51.3	189	797.3	37.2	+ 5.9
207	47.8	55.7	189	795.1	37.2	+ 3.8

Appendix C.
Density Data by Rock Types.

Rock Type	No.	Mean Density gm/cm³	Corrected Standard Deviation gm/cm³	Range gm/cm³
Aplite	6	2.62	0.023	2.58—2.65
Quartzite	6	2.64	0.017	2.62—2.67
Marble	1	2.71	—	—
Garnetiferous granite gneiss	2	2.72	—	2.69—2.76
Biotite granite gneiss	16	2.66	0.038	2.59—2.71
Hornblende granite gneiss	2	2.73	—	2.70—276.
Augen gneiss	9	2.72	0.052	2.67—2.83
Biotite gneiss	22	2.78	0.049	2.67—2.88
Migmatitic biotite gneiss	6	2.74	0.024	2.70—2.76
Migmatitic hornblende gneiss	2	2.89	—	2.83—2.95
Banded hornblende gneiss	6	2.93	0.059	2.88—3.04
Biotite hornblende gneiss	10	2.90	0.043	2.82—2.97
Amphibolite	13	3.03	0.080	2.86—3.17
Gabbro	3	2.96	—	2.86—3.03

References.

Ansel, E. A. 1937. Zur Analyse von Schwereanomalien, Beitr. angew. Geophys., v. 7, p. 21—38.

Balk, Robert. 1947. Salt-dome structure, Bull. Amer. Assoc. Petrol. Geol., v. 31, p. 1296—99.

— 1949. Structure of Grand Saline Salt Dome, Van Zandt County, Texas, *ibid.*, v. 33, p. 1791—1829.

— 1953. Salt structure of Jefferson Island Salt Dome, Iberia and Vermilion Parishes, Louisiana: *ibid.*, v. 37, p. 2455—2474.

Barth, T. F. W. 1929. Die Temperatur der Anatexis des Urgebirges in südlichsten Norwegen, Centralbl. Min., Abteil. A, p. 120—27.

— 1935. The large pre-Cambrian intrusive bodies in the southern parts of Norway, Rpt. XVI Int. Geol. Congress, Washington, p. 297.

— 1947a. The nickeliferous Iveland-Evje amphibolite and its relation, Norges geol. undersøk., n. 168a, 71 p.

— 1947b. The Birkeland granite, a case of petroblastesis, Compt. Rend. Soc. géol. Finlande, n. 20, p. 173—79.

— 1952. Theoretical petrology, New York, John Wiley & Sons, 387 p.

— 1956. Studies in gneiss and granite. I Relation between temperature and the composition of the feldspars. II The feldspar-equilibrium temperature in granitic rocks of the pre-Cambrian of Southern Norway, Vid. Akad. Skr., n. 1, 35 p.

— 1960. Precambrian of southern Norway, *in* Holtedahl, Olaf, *Editor*, The geology of Norway, Norges geol. undersøk., n. 208, p. 5—57.

— 1961. Ideas on the interrelation between igneous and sedimentary rocks, Bull. Comm. géol. Finlande, n. 196, p. 321—26.

— and *Bugge, J. A. W.* 1960. Precambrian gneisses of the Skagerrak coastal area of South Norway, Guide to excursion A 8, XXI Int. Geol. Congress, 34 p.

Bean, R. J. 1953. Relation of gravity anomalies to the geology of central Vermont and New Hampshire, Geol. Soc. America Bull., v. 64, p. 509—538.

Birch, F. 1955. Physics of the crust, *in* Poldervaart, A., *Editor*, Crust of the earth, GSA Special Paper 62, p. 101—118.

— 1960. The velocity of compressional waves in rocks to 10 kilobars, Part I, Jour. Geophys. Res., v. 65, n. 4, p. 1083—1102.

Bott, M. H. P. 1953. Negative gravity anomalies over acid «intrusions» and their relation to the structure of the earth's crust, Geol. Mag., v. 91, p. 377—83.

— 1956. A geophysical study of the granite problem, Geol. Soc. London Quart. Jour., v. 112, p. 45—62.

— and *Masson-Smith, D.* 1960. A gravity survey of the Criffell granodiorite and the New Red sandstone deposits near Dumfries, Proc. Yorkshire Geol. Soc., v. 32, pt. 3, n. 13, p. 317—332.

Buddington, A. F. 1959. Granite emplacement with special reference to North America, Geol. Soc. America Bull., v. 70, p. 671—748.

Bugge, A. 1928. En forkastning i det sydnorske grunnfjell, Norges geol. undersøk., n. 130.

— 1936. Kongsberg-Bamble Formationen, *ibid.*, n. 146.

Bugge, J. A. W. 1943. Geological and petrological investigations in the Kongsberg—Bamble formation, *ibid.*, nr. 160.

Bullard, E. 1954. The interior of the earth, *in* Kuiper, G. P., *Editor*, The earth as a planet, U. of Chicago Press, p. 57—138.

Collette, B. J. 1960. The gravity field of the North Sea, Gravity Expeditions 1948—1958, v. 5, pt. 2, Netherlands Geodetic Comm., Delft, 52 p.

Cook, A. H., and *Murphy, T.* 1952. Measurements of gravity in Ireland. Gravity survey of Ireland north of the line Sligo-Dundalk, Geophys. Mem. Dublin Inst. Advanced Studies, n. 2, pt. 4, p. 1—36.

DeSitter, L. U. 1956. Structural geology, New York, McGraw-Hill, 552 p.

Dietrich, R. V. 1959. Geological reconnaissance of the area between Kristiansand and Lillesand, Norges geol. undersøk., n. 205, p. 41—78.

— 1960. Banded gneisses of the Randesund area, southeastern Norway, Norsk Geol. Tidsskr., v. 40, p. 13—63.

Dobrin, M. B. 1941. Some quantitative experiments on a fluid salt-dome model and their geologic implications, Trans. Amer. Geophys. Union, v. 22, pt. 2, p. 528—42.

— 1960. Introduction to geophysical prospecting, New York, McGraw-hill, p. 169—263.

Dons, J. A. 1960. Telemark supracrustals and associated rocks, *in* Holtedahl, Olaf, *Editor*, Geology of Norway, Norges geol. undersøk., n. 208, p. 49—57.

Elders, W. A. 1961. Geological and petrological studies in the Precambrian gneisses and granites of the southwest part of Aust-Agder, Southern Norway, University of Durham Ph. D. thesis, Durham, England.

— 1963. On the form and mode of emplacement of the Herefoss granite, Norges geol. undersøk. n. 214a p. 1—52.

Escher, B. G. and *Kuenen, Ph. H.* 1929. Experiments in connection with salt domes, Ledische Geologische Mededeelingen, Deel 3, Aflevering, 3, p. 151—82.

Eskola, P. 1948. The problem of mantled gneiss domes, Geol. Soc. London Quart. Jour., v. 104, p. 461—476.

Garland, G. D. 1950. Interpretations of gravimetric and magnetic anomalies on traverses in the Canadian Shield in northern Ontario, Pub. Dom. Obs., Ottawa, v. 16, n. 1, 57 p.

— 1953. Gravity measurements of the Maritime Provinces, *ibid.*, v. 16, n. 7, p. 187—243.

Glennie, E. A. 1932. Gravity anomalies and the structure of the earth's crust, Surv. of India Prof. Paper 27.

Goguel, J. 1950. Interprétation géologique, *in* Mesures gravimétriques en Bretagne, Publ. Bur. Rech. Géol. Géophys., n. 8, Paris.

Grosse, S. 1958. Ergebnisse gravimetrischer Regionalvermessungen im Westerzgebirge, Zeitschr. Geophys., v. 24, n. 415, p. 321—25.

Gutenberg, *B*. 1951. Internal constitution of the earth, New York, Dover.

— 1955. Wave velocities in the earth's crust, *in* Poldervaart, A., *Editor*, Crust of the Earth, Geol. Soc. Amer. Spec. Paper 62, p. 19—34.

Hammer, S. 1939. Terrain corrections for gravimeter stations, Geophysics, v. 4, p. 184—94.

Heier, *K*. S. and *Taylor*, S. *R*. 1959. Distribution of Li, Na, K, Rb, Cs, Pb, and Tl in southern Norwegian pre-Cambrian alkali feldspars, Geochem. et Cosmochim. Acta, v. 15, p. 284—304.

Hinze, *W*. *J*. 1959. A gravity investigation of the Baraboo syncline, Jour. Geol., v. 67, p. 714—46.

Holtedahl, *Olaf*. 1940. The submarine relief off the Norwegian coast, Norsk Videnskaps-Akademi, Oslo, 43 p.

Hubbert, *M*. *K*. 1937. Theory of scale models as applied to geologic structures, Geol. Soc. America Bull., v. 48, p. 1459—1519.

— 1948a. Gravitational terrain effects of two-dimensional topographic features, Geophysics, v. 13, p. 226—54.

— 1948b. Line-integral method of computing gravity, *ibid*., p. 215—225.

Jung, *K*. 1927. Diagramme zur Bestimmung der Terrainwirkung für Pendel und Drehwaage und zur Bestimmung der Wirkung «zweidimensionaler» Massenanordungen. Zeitschr. Geophys., Bd. 13, p. 201—12.

Kennedy, *W*. *Q*., and *Anderson*, *E*. *M*. 1938. Crustal layers and the origin of magmas, Bull. Vulcanol. Naples, ser. 2, t. 3, p. 23—82.

Macgregor, *A*. *M*. 1951. Some milestones in the Precambrian of southern Rhodesia, Geol. Soc. S. Africa Trans., v. 54, p. xxvii-lxxi.

Nettleton, *L*. *L*. 1934. Fluid mechanics of salt domes, Bull. Amer. Assoc. Petr. Geol., v. 18, p. 1175—1204.

— 1940. Geophysical prospecting for oil, New York, McGraw-Hill, p. 1—154.

— 1942. Gravity and magnetic calculations, Geophysics, v. 7, p. 293—310.

— 1943. Recent experimental and geophysical evidence of mechanics of salt-dome formation, Bull. Amer. Assoc. Petr. Geol., v. 27, p. 51—63.

Nilssen, *B*. 1961. Noen geologiske undersøkelser av Herefossgraniten, University of Oslo thesis.

Noble, *J*. *A*. 1952. Evaluation of criteria for the forcible intrusion of magma, Jour. Geol., v. 60, p. 34—57.

Oftedal, *I*. 1938. Om Fevik-granitten, Norsk geol. tidsskr., v. 17, p. 210—213.

— 1945. Fra Fevik-granittens grensesoner, *ibid*., v. 25, p. 295—306.

Orville, *P*. *M*. 1960. Alkali feldspar- alkali chloride hydrothermal ion exchange, Geophys. Lab. Ann. Rept. Director for 1959—1960, p. 104—108.

— 1961. Alkali metasomatism produced by alkali ion exchange within a thermal gradient (Abstract), Program 1961 Ann. Meetings Geol. Soc. America, p. 119 A.

Parker, T. J. and *McDowell, A. N.* 1955. Model studies of salt-dome tectonics, Bull. Amer. Assoc. Petr. Geol., v. 39, p. 2384—2470.

Pustowka, A. 1929. Uber rumänische Saltztonaufbruche, *in* Krejci-Graf, *Editor*, Beiträge zur Kennitnis der Tektonik Rumäniens, Neues Jahrb. für Min., Geol. Pal., 61st Beilage Bd. Ab. 13, p. 317—398.

Raguin, E. 1957. Geologie du granite, Paris, Masson et Cie, 275 p.

Read, H. H. 1955. Granite series in mobile belts, *in* Poldervaart, A., *Editor*, Crust of the earth, Geol. Soc. America Spec. Paper 62, p. 409—30.

Reich, H. 1932. Die Bedeutung der finnischen Schweremessungen fur die angewandte Geophysik, Beitr. angew. Geophys. Ergänzungshefte, Bd. 2, p. 1—13.

Reynolds, D. L. 1958. Granite, some tectonic petrological and physico-chemical aspects, Geol. Mag., v. 95, p. 378—96.

Romberg, F. and *Barnes, V. E.* 1944. Correlation of gravity observations with the geology of the Smoothing-iron granite mass, Llano Co., Texas, Geophysics, v. 9, p. 79—93.

Schwinner, R. 1928. Geologischen Bemerkungen zu den Norwegischen Schweremessungen, Gerl. Beitr. z. Geophys. Bd. 19, p. 107—33.

Selmer-Olsen, R. 1950. Om forkastninger og oppbrytningssoner i Bambleformasjonen, Norsk geol. tidsskr., v. 28, p. 171—91.

Steenland, N. C. and *Woolard, G. L.* 1952. Gravity and magnetic investigation of the Cortlandt complex, New York, Geol. Soc. America Bull, v. 63, p. 1075—93.

Stille, H. 1925. The upthrust of the salt masses of Germany, Bull. Amer. Assoc. Petr. Geol., v. 9, p. 417—41.

Sosman, R. B. 1948. Discussion, in Gilluly, J., *Editor*, Origin of granite, Geol. Soc. America Mem. 28, p. 116—17.

Tatel, H. E. and *Tuve, M. A.* 1955. Seismic exploration of a continental crust, *in* Poltervaart, A., *Editor*, Crust of the earth, Geol. Soc. Amer. Spec. Paper 62, p. 35—50.

Trovaag, O. 1956. Høyder for presisjonsnivellement i Sør-Norge, Norges geografiske oppmåling, Oslo.

Wegmann, E. 1930. Über Diapirismus, Bull. Comm. géol. Finlande, n. 92, p. 58—76.

— 1935. Zur Deutung der Migmatite, Geol. Rundschau, Bd. 26, p. 305—50.

— 1960. Introductory remarks on the structural relations, *in* Holtedahl, Olaf, *Editor*, Geology of Norway, Norges geol. undersøk. n. 208, p. 6—8.

Willmore, P. L. 1949. Seismic experiments on the north German explosions, 1946—47, Roy. Soc. London Phil. Trans., v. A 242, p. 123—51.

Winkler, H. G. F. and *von Platen, H.* 1958. Experimentelle Gesteinsmetamorphose II. Bildung von anatektischen granitischen Schmelzen bei der Metamorphose von NaCl-führenden kalkfreien Tonen, Geochim. et Cosmochim. Acta, v. 15, p. 91—112.

Wollard, G. P. 1959. Crustal structure from gravity and seismic measurements, Jour. Geophys. Research, v. 64, n. 10, p. 1521—1544.

Worzel, J. L. and *Shurbet, G. L.* 1955. Gravity interpretations from standard oceanic and continental crustal sections, *in* Poldervaart, A., *Editor,* Crust of the earth, Geol. Soc. Amer. Spec. Paper 62, p. 87—100.

Wyart, J. and *Sabatier, G.* 1959. Transformation des sediments pelitiques à 800 C. sous une pression d'eau de 1800 bars et granitisation, Bull. Soc. franc. Minér. Crist., Tome 82, p. 201—210.

Wyllie, P. J. and *Tuttle, O. F.* 1958 Hydrothermal experiments on the melting temperatures of shales (Abstract), Amer. Geophys. Union Trans., v. 38, p. 413—14.

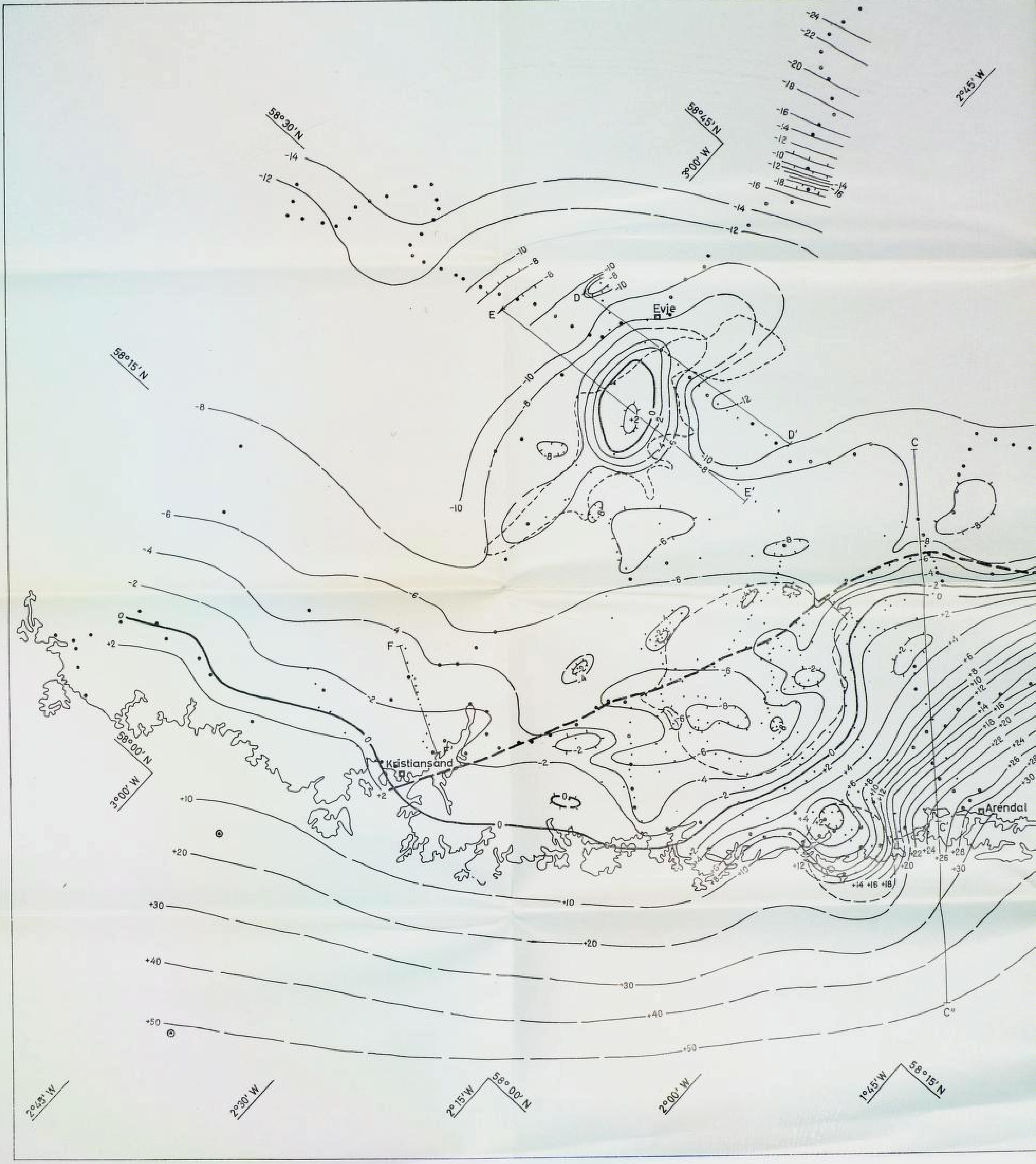
— 1960. Melting in the earth's crust, Rpt. of 21st Session, Int. Geol. Congress, pt. 18, p. 227—35.

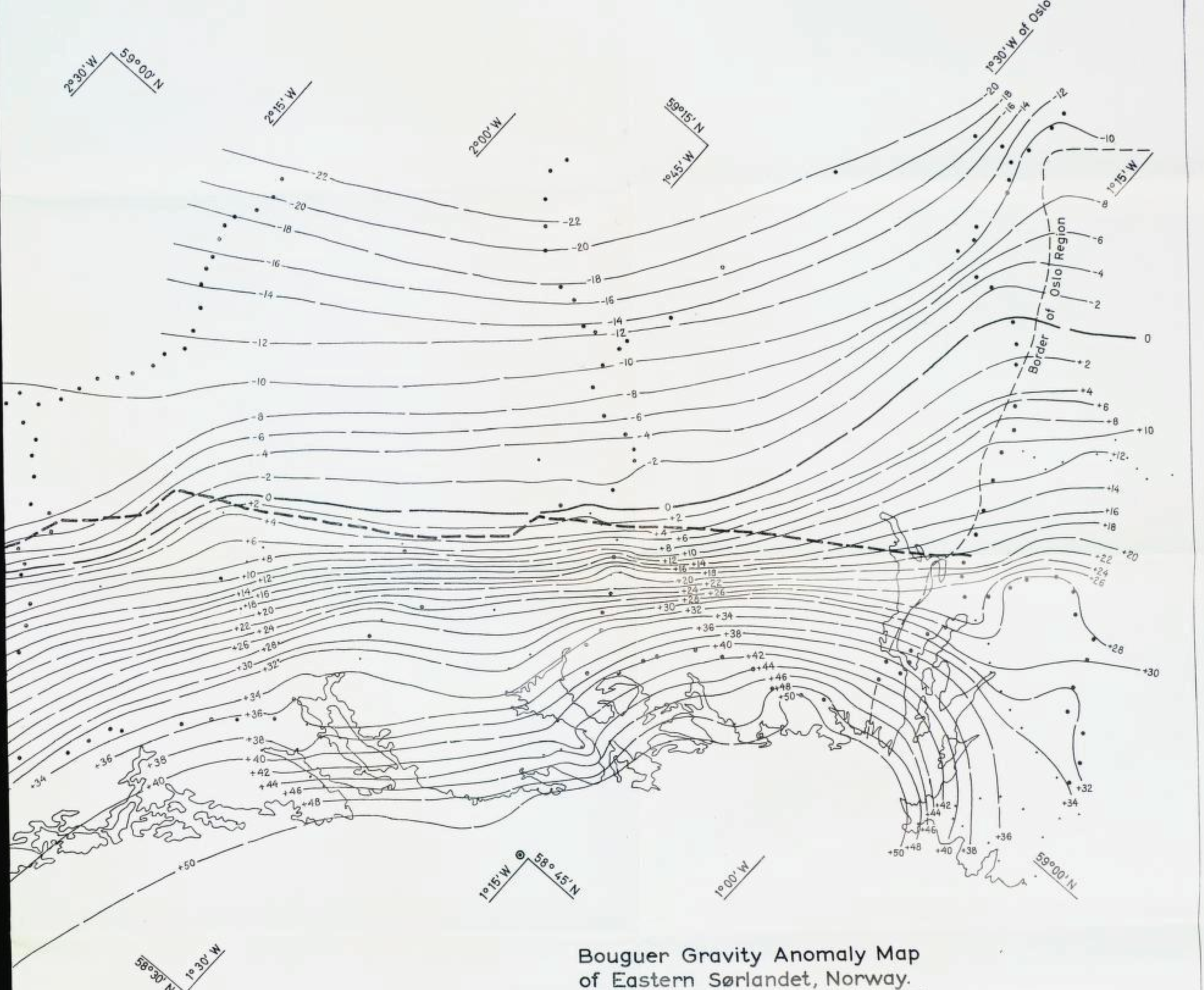
Sammendrag.

Gravimetriske undersøkelser på Sørlandet.

Tyngdeanomalier på Sørlandet blir behandlet og tolket ut fra de geologiske forhold. Tykkelsen av to granitter er beregnet på grunnlag av tyngdeanomaliene. Granittene er fra 2 til 5 km tykke. Evjeamfibolitten er beregnet til å være 1,25 km tykk. På grunnlag av tyngdeanomaliene blir det også fremsatt noen teorier for granittdannelse.


Geologic Map
of Eastern Sørlandet
Compiled by Tom. F. W. Barth
Gabbro, Hyperite
Amphibolite
Quartzite with sillimanite (S)
Limestone
Mixed gneisses
Granite gneisses
Arendalite
Augen gneiss
Gneissic granite
Coarse-grained
Fine-grained
Granite
Pegmatite dikes (in Iveland)
Dolerite dikes, basic plugs
Norite dikes
Diabases and Lamprophyres
Rhomb porphyry dikes
Tectonic lines
Scale
5 0 5 10 15
Kilometers
Evje
Vegarshei
Kragerø
Risør
Herefoss
Arendal
Oddersjå granite
Grimstad
Lillesand
Kristiansand


58°30'N
58°45'N
3°00'W
2°45'W
58°15'N
Evje
Kristiansand
Arendal
58°00'N
3°00'W
2°45'W
2°30'W
2°15'W
58°00'N
2°00'W
1°45'W
58°15'N
D
D'
E
E'
F
F'
C
C'
C"



Bouguer Gravity Anomaly Map of Eastern Sørlandet, Norway.

Bouguer Density = 2.67
Contour Interval = 2 mgal (10 mgal offshore)

- —— — Isoanomaly lines,-broken where approximately located
- – – – – Geologic contacts
- — — — Trace of the "Great Friction Breccia"
- ∘ Gravity station based on levelled elevation
- · Gravity station based on barometric elevation
- ⊛ Gravity station measured from submarine

Compiled by Scott B. Smithson

0 5 10 15 km