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Gravity Interpretation of the Oslo Graben and Associated Igneous Rocks*

IVAR B. RAMBERG

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A gravity study of the well-exposed Lower Permian Oslo Graben, Norway, has been carried out to obtain an estimate of mass and geometry of the associated plutonic bodies, to determine the gross crustal structure along the rift zone, and to investigate how geophysical and geological data from a paleorift fit with modern rifts and plate tectonics theory. Data from about 5300 gravity stations have been collected from the approximately 8500 km^2 graben area and neighboring districts. The geology of the area is presented with emphasis on items that affect the interpretation most significantly, such as structural trends, rock-types involved, and theories of petrogenesis. More than 1900 rock density measurements have been made from the graben area and adjacent Precambrian terrain. The weighted mean density of the Permian plutonic rocks is about 2.66 g/cm^3, resulting in a negative density contrast of about 0.08 g/cm^3 with respect to the Precambrian gneisses (2.74 g/cm^3). Concentric lateral density variations occur within single rock units in general harmony with the common occurrence of ring complexes. A Bouguer map prepared for Southern Norway has been used to construct a contour map of the thickness of the gravity normal crust. This map indicates crustal thinning toward the coastlines and below the graben axis. The gravity field of the Oslo Region has been separated into a 'regional' and a 'residual' field, respectively. The regional map reveals a broad gravity high of about 45 mgal along the entire rift zone. The residual field includes a number of local highs and lows (with magnitudes up to 30 mgal) which generally closely follow the outlines of the exposed complexes.

Two- and three-dimensional modelling of the 5 major felsic bodies has revealed that the density contrasts extend down to different depths from about 10–12 km or more for the huge Nordmarka-Hurdalen syenite batholith to about 3–5 km for the funnel-shaped Drammen granite complex. The favored geological interpretation is that the various felsic bodies are not floored by solid gneisses but that they grade into a mixture of more basic intrusives, stoped blocks and accumulates. The basic 'Oslo-essexitic' plugs all show remarkably small anomalies (mostly around 2–3 mgal) and hence constitute only minor masses and volumes. They may represent sub-volcanic storage chambers fed by basaltic magma through narrow vents or feeders. Of the several cauldrons, two of the complexes are associated with felsic central plutons whereas three or possibly four seem to ride on central intrusions of an overall intermediate composition, with only minor volumes of felsic rocks present. Several local gravity highs have been tentatively ascribed to the presence of shallow, mafic, Permian intrusives, while certain Precambrian mafic bodies (Kongsberg, Tyrifjord) mark the western border of the graben area.

* *Editors note.*

The final manuscript of this paper was received in April 1975 and accepted after minor revision in August 1975. Because of the general interest of the subject-matter and in view of the fact that a generous financial contribution to the printing had been made by the Norwegian Research Council for Science and the Humanities, it was decided to accept the manuscript without further abridgement. We wish to state, however, for the guidance of future contributors that, primarily for fincancial reasons, excessively long manuscripts will not normally be considered for publication in the NGU Bulletin series.

The regional gravity high demonstrates that the predominantly felsic intrusives of the upper crust are associated with significant amounts of dense rocks at depth. Analysis suggests that this high is perhaps the combined effect of a marked crustal thinning of about 6–10 km along the graben axis and of a dense block of mafic rocks located in the deep to intermediate crust. Mass estimates of the various intrusives and extrusives show a frequency distribution with a striking preponderance of mafic/ultramafic rocks (anomalous mass $\Delta M = 1.3 \cdot 10^{13}$ tons, or more), relative to the felsic rocks ($\Delta M = 1.9 \cdot 10^{12}$ tons). The calculated mass relations put severe constraints on earlier petrological models for the Oslo igneous rock suite. Gravity data imply that there is no volumetric objection to the idea that the felsic rocks of the Oslo Region originated from a basaltic parent magma ultimately of upper mantle provenance. It is suggested that the rise of a mantle diapir which was accompanied by plastic deformation and necking in the deeper crust and tensile faulting and fracturing of the brittle upper crust, led to intrusions of large masses of hot material into deep and intermediate crustal levels. Assuming paleo-grabens and modern rifts to be products of similar deep-seated processes, the thermal régime associated with rifting no longer applies to the Oslo Graben but the physical consequences (i.e., the dense axial intrusion and differentiated igneous suite) remain. The Oslo Graben would seem to be part of a regional, braided fracture system, the Graben representing a failed arm in Permian time. The graben and related features indicate that Mesozoic to Cenozoic spreading of the North Atlantic followed periods of tension and crustal break-up in the neighboring continental masses.

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CONTENTS

1. Introduction

1.1 GENERAL REMARKS

Trains of elongate fault troughs which constitute grabens or rift valleys are major tectonic features of the earth upon which attention has recently been focused through the revolution brought about by plate tectonics. In spite of various apparent ages and pre-existing structures, contrasted geological histories and different lithologic compositions, most of these grabens show great similarities in their physiographic appearance and morpho-tectonic development. The Paleozoic Oslo Graben has for years been thought to form the northern end of a more or less continuous trans-European rift system from the Mediterranean to Lake Mjøsa, Norway (Stille 1952, Illies 1970, McConnel 1972, and others). Because of the relatively deep erosional level and variable resistance to erosion, the Oslo Region generally shows a positive physiographic relief, and offers the opportunity of direct observations of the deeper structures of a graben. It also represents one of the best exposed and geologically most studied segments of the postulated rift system.

The Oslo Region, which largely consists of Cambro-Silurian sedimentary rocks and Permian plutonic and volcanic rocks, has subsided into the surrounding Precambrian gneisses of the Fennoscandian Shield. The present work consists of a gravimetric and geologic study of the Oslo Region, designed and directed towards a three-dimensional structural interpretation of the graben and associated plutonic rocks.

The general geological and geographical situation is depicted in Fig. 1. As seen from the figure the Oslo Graben has a NNE–SSW trend. It extends about 200 km from the coast of Skagerrak to Lake Mjøsa (from about 59°00′ to 60°40′ N). The width is that of a typical graben (Table 1), the average being about 40 km. In common with many other zones of rifting, it is not a simple graben but is characterized by irregular block faulting and a rather asymmetric outline. Detectable major faults only occur on the east flank in the southern part and on the west side in the northern part of the region. Of the present-day surface 75–80% consists of plutonic and volcanic rocks of the alkaline kindred typical of continental rifts. In a very generalized way the Oslo Region igneous rocks can be divided into three major, composite batholiths (Fig. 1), all three mutually separated by belts of Cambro-Silurian and/or Precambrian rocks which are continuous across the graben 'floor'.

Note on terminology: The term 'rift valley' as applied by Gregory (1896) refers to elongate crustal blocks that have subsided between faults. The term always has a morphological implication, as has the expression 'graben'. The widely used term 'rift' suggests that these structures have formed as a result of extension. Structurally a rift is always a complex graben (Beloussov 1969).

The Oslo Graben only partly fulfills the structural and morphological requirements. The term has, however, been retained for the subsided area because of its common use in earlier literature and the many similarities between the 'Oslo Graben' and other graben regions. The 'Oslo Rift' or 'rift system' is applied in a broader sense, referring to the more extensive fracture zone of which the graben is but a part. The petrographic notations 'Oslo Igneous Province' or 'Igneous Rock Complex' have, according to previous usage, been reserved for the alkaline rock series associated with the graben. 'Oslo Region' is a general geographical term, loosely referring to the area characterized by the structural features and lithological units restricted to or associated with the area of subsidence.

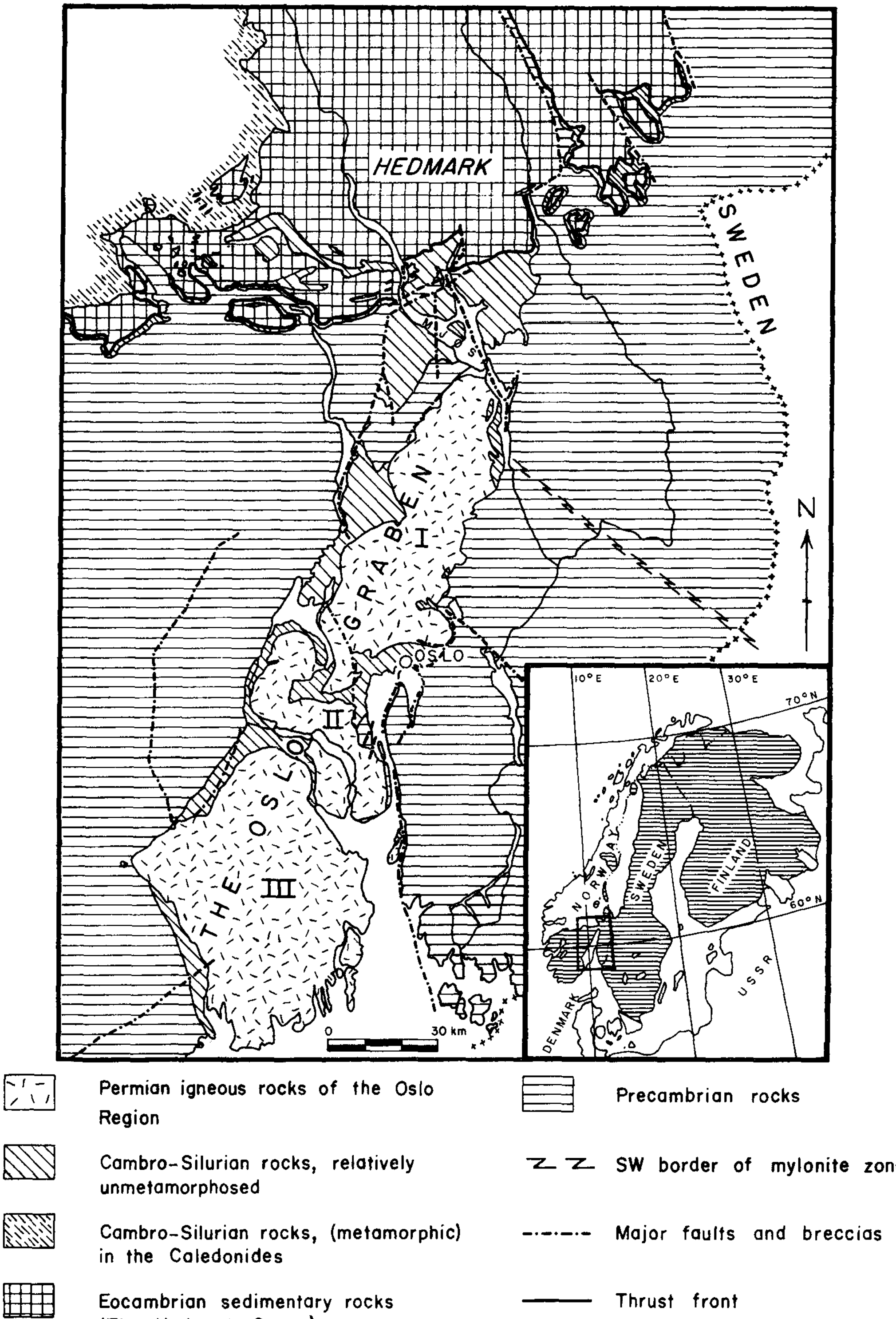


Fig. 1. Location of the Oslo Region.

Table 1: Mean widths of continental and oceanic rifts (Girdler 1964, Beloussov 1969)

Rift	~ Width, km
Oslo	40
Rhine	40
Baikal	50
Midland Valley, Scotland	75
East African rifts	50
Dead Sea	35
Gulf of Suez	35
Gulf of Aquaba	50
Red Sea (inner graben)	60
Mid-Indian	25
Mid-Atlantic	25–40

1.2 PURPOSE AND METHODS OF STUDY

The object of the investigation is threefold: 1) to obtain an estimate of mass and geometry of the various plutonic bodies, 2) to determine the gross crustal structure along the rift zone, and 3) to compare the data and interpretations with recent interpretations of continental and oceanic rifts.

Points (1) and (2) will provide a three-dimensional interpretation of the Oslo Region that might serve as a basis for models for *a*) the tectonic development of the Oslo rift, and *b*) the petrogenesis of the Oslo igneous province. Many of the present and past tectonic and volcanic events may be better understood in the light of (intra-)plate tectonics theory, and point (3) is an effort to clarify similarities and differences that might exist between recent rifts and possible equivalent features of an older age, like the deeply eroded Paleozoic Oslo Graben.

Several contrasted ideas have been suggested in the literature about graben and rift formation (Chapter 3), many of which imply large-scale mass differences. The many rock types exposed in the Oslo Region also reveal significant density contrasts (Chapter 4). These facts suggest the usefulness of gravity studies since, according to the law of gravitational attraction, any lateral variation in the subsurface mass distribution will make a corresponding variation in the gravity, g, at the surface. To have the variations in the gravity field transformed into meaningful geological models, the observed data have to be corrected for all sources of variation other than geological features. The corrections applied are briefly outlined in Chapter 5.

Interpretation of gravity anomalies involves solutions to the 'inverse' potential problem, that is, deducing the shape of the anomalous rock body from its potential field. These solutions are never unique. Nor are they wholly ambiguous, since they also rely upon factors other than gravity potentials. Such limiting factors in the Oslo Region are: a) geologic maps and structural observations, b) known and inferred rock densities, c) deductions about the geological history and petrogenesis of exposed rocks, d) geochemical data and considerations, and e) geophysical studies other than gravity (especially seismic

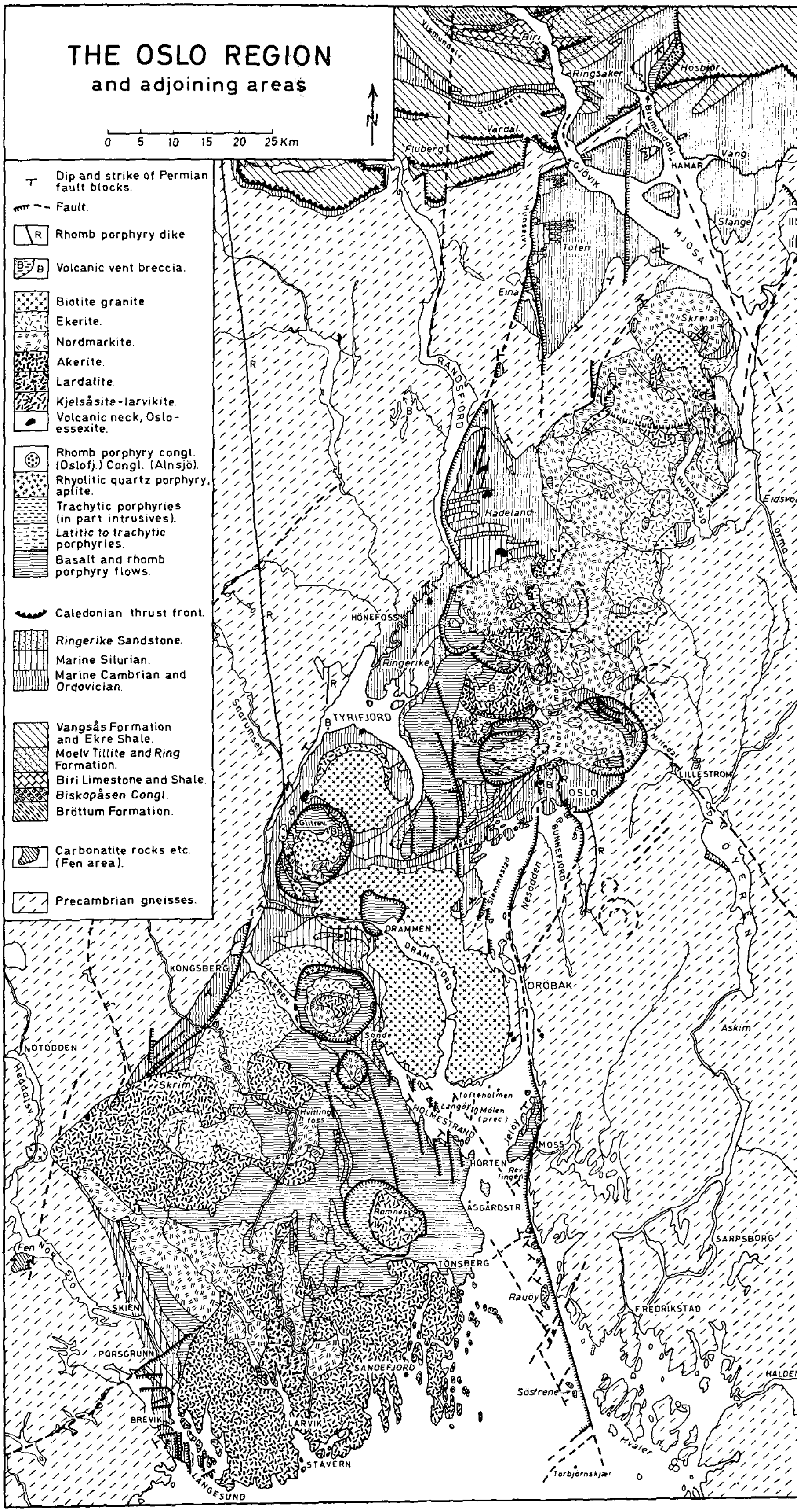


Fig. 2. Geological map of the Oslo Region (slightly modified *after* Oftedahl, 1960).

results) providing information about physical properties of rocks at depth. The rather critical rôle of various items of geological information is reflected in a fairly detailed review of the geology of the Oslo Region as presented in Chapter 2.

The field work was conducted during the period 1966–1972. Detailed geological studies were carried out in selected areas. The main emphasis was on gravity field work, which was first concentrated in the northern part of the region (largely occupied by syenitic intrusives). Late in this work, however, it was decided to extend the gravity survey to the whole Oslo Region, since the solution of the problems confronted both in the southern and in the northern halves would greatly benefit from a more general approach. Detailed descriptions of field methods and gravity data from the various subregions are presented separately (Ramberg 1972a).

1.3 PREVIOUS LITERATURE

The Oslo Region has been intensely studied by geologists since the middle of the last century, while geophysical studies have been undertaken just in the last few years. A comprehensive literature describing various aspects of the igneous and Cambro-Silurian metasedimentary rocks exists. Here, only the main sources of information will be pointed out.

The Oslo Region became known first of all through the works of W. C. Brøgger, particularly his monograph of 1890 on the geology and mineralogy of the Langesundsfjorden nepheline-syenite pegmatites, and later the series 'Die Eruptivgesteine des Kristiania-(Oslo)-Gebietes'. Complete references to the publications of Brøgger have been given by Barth (1945), while Holtedahl (1934) has surveyed the older literature. A new series of monographs, 'Studies of the Igneous Rock Complex of the Oslo Region,' was initiated by the papers by Holtedahl (1943) and Barth (1945), and now totals some 26 volumes.

General geological surveys of the Oslo Region have been presented by Holtedahl (1934, 1953) and more recently by Oftedahl (1960). A general map of the Oslo Region was prepared by Brøgger & Schetelig (1923). A revised version was published by Oftedahl (1960) (Fig. 2).

The relatively thin (max. ca. 1400 m) but almost complete sequence of fossiliferous, marine Cambro-Silurian deposits, and the succeeding Ringerike sandstone (Old Red facies) have been surveyed by Henningsmoen (1960). Bjørlykke (1974) has studied the geochemistry and sedimentary petrology of the strata, and also presents an up-to-date summary of the relevant literature. The termal effects of the igneous intrusions on the Cambro-Silurian sediments have become 'classic' through the work of Goldschmidt (1911).

Aeromagnetic maps of the whole region have been prepared by the Geological Survey of Norway and interpretation is in progress (Åm & Oftedahl, in prep.). Results from a seismic refraction profile that crosses the Oslo Region have been presented by Kanestrøm & Haugland (1971). Detailed crustal studies around the Norwegian Seismic Array (NORSAR) situated at the northern end of the graben are either recently published (Kanestrøm 1973, Berteussen 1975) or in progress, while preliminary results from seismic refractions studies in the central part of the Oslo Region have been reported by Sellevoll (1972). A Bouguer gravity map of the Oslo Region was presented in 1960 by the Norwegian Geographical Survey (NGO). Interpretation and preliminary reports of various parts of the gravity anomalies in the Oslo Region have been given by Smithson (1961), Ramberg & Smithson (1971), Grønlie (1971) and Ramberg (1971, 1972b and 1973).

2. Geology

A fairly detailed summary of the geology is regarded as a necessary background for a gravity interpretation and discussion of such a complex region as that of the Oslo Graben and its associated igneous rocks. In the following description the main emphasis is on Permian rocks and structures, while only a brief outline is given for the Cambro-Silurian and surrounding Precambrian formations. The description largely relies upon the voluminous earlier literature, but is supplemented by the author's own observations obtained through the field seasons 1966–1972.

2.1 THE PRECAMBRIAN

The Precambrian of Norway is part of the Fennoscandian Shield. Based on radiometric datings and geologic structure the shield has been subdivided into three major zones: the Saamo-Karelian zone (3600–1900 m. y.), the Sveco-Fennian zone (2300–1650 m. y.) and the Sveco-Norwegian zone (1200–900 m. y.) (Polkanov & Gerling 1961, Kratz et al. 1968); see Fig. 3. The Sveco-Norwegian zone, which occupies the southwestern part of the shield, has isotopic ages generally around 1000 m. y. and younger, but relict ages of older rocks dating at 1600–1700 m. y. commonly occur (Neumann 1960, Broch 1964, O'Nions et al. 1969, Priem et al. 1970, O'Nions & Heier 1972).

The Sveco-Norwegian zone is dissected by the Oslo Region rift zone: a) the southeastern part contiguous to the Precambrian of Sweden, and b) the main southwestern part. Both parts are characterized by a variety of gneisses and migmatites together with quartzites, amphibolites, syn- and post-orogenic granites and gabbroic rocks. While the southwestern region is composed predominantly of granite and granodioritic gneisses (Barth & Reitan 1963, Table 1), the Kongsberg-Bamble portion of the Precambrian, situated just west and southwest of the Oslo Region (Bugge 1943), contains abundant meta-volcanic and plutonic basic rocks.

No obvious parallelism exists between the Precambrian metamorphic fabric such as foliation, fold axial traces, etc., and the borders of the Oslo Region as defined by faults, intrusive or erosional contacts. Detailed observations at

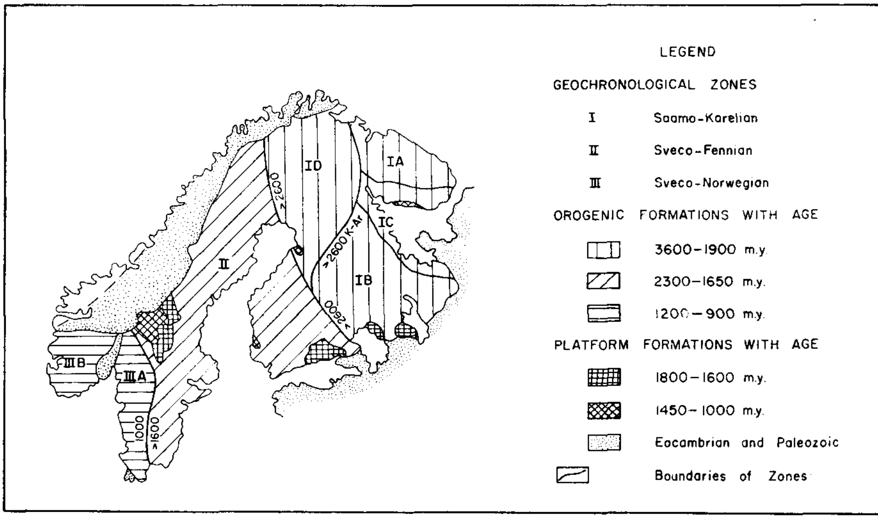


Fig. 3. Subdivision of the Precambrian terrain in Fennoscandia in geochronological zones (*after* Kratz et al. 1968).

most contact localities reveal cross-cutting relations. Nevertheless, on a regional scale there seems to be a certain connection between the various tectonic directions of the rift and older crustal lineaments:

Two dominant trends of faults, breccias or mylonite belts occur in the surrounding Precambrian regions, that is about NNW–SSE (to N–S) and about NE–SW (Cloos 1928, Holtedahl 1943, 1953, Gleditsch 1945, 1952, Selmer-Olsen 1950, Skjeseth 1963, Morton et al. 1970). The rocks of many of these belts are blastomylonites with clear evidence of repeated fragmentation, as noted already by Brøgger (1886) from faults on both sides of the Oslofjord. Similar features have been observed for instance at the border of the Nittedal cauldron subsidence (Naterstad 1971) where a belt of Precambrian mylonites occur in amphibolite facies and with clear signs of repeated fragmentation. This crush belt is part of an extensive tectonic line with blastomylonites that can be followed toward SSE to the Swedish border (Skjernaa 1972). A multi-stage movement has also been envisaged by Bugge (1928, 1936, 1965) for the 'Great friction breccia' separating the Bamble and Kongsberg regions from the more granitic Telemark area (see Barth & Dons 1960, fig. 2). Bugge identifies at least two periods of movement in Precambrian time. Later, the broad belt of cataclastites was cut by steep fault zones with slickensides and associated quartz-carbonate dikes of presumed Permian age. This late feature has been termed the Porsgrunn-Kristiansand fault (Morton et al. 1970) along the Bamble/Telemark contact zone. To the north of the graben region, prominant Eocambrian fault or flexure zones occur along the same major directions, such as the Rendal fault (Skjeseth 1963) and the Engerdal fault (Holtedahl 1921) and seem to coincide with late Caledonian or probably Permian faults.

The two major fault directions also define the border lines of the Oslo

Graben and prevail within the graben area where they are marked by dikes and local contacts. The occurrence of Permian dikes far outside the graben region along the same major directions points to a splitting up of the Precambrian crust over a wide area in Southern Norway and again demonstrates a strong dependence of the Permian rocks upon older trends as outlined by the Precambrian faults and crush belts.

2.2 LATEST PRECAMBRIAN AND EOCAMBRIAN

To the north of the Oslo Region, in the Mjøsa district, is the type area of the so-called 'Sparagmite Group' (Münster 1901, Goldschmidt 1908, Vogt 1924, Holtedahl 1960, 1961, Skjeseth 1963, etc.), now termed the Hedmark Group (Bjørlykke et al. 1967). This is a thick sedimentary succession of mainly coarse feldspathic sandstones or arkose deposited at the Precambrian-Cambrian transition, the total thickness being about 1,500–2,000 m (Skjeseth 1963) or more (Ramberg & Englund 1969).

The sedimentary sequence is considered to have been deposited in faulted basins at an early stage in the formation of the main Caledonian geosyncline. The sedimentation is continuous into the Lower Cambrian shales which were deposited chiefly in the same troughs as the stratigraphically underlying Hedmark Group. Folding and thrusting took place mainly in the late, post-Silurian phase of the Caledonian orogeny. Later faults cut through the fold structures and the thrust planes along the same directions as the supposed marginal faults (NNW–SSE and NE–SW); these faults may be of late Caledonian (Upper Devonian) or Permian age (Skjeseth 1963, Bjørlykke 1966, Englund 1971).

2.3 CAMBRO-SILURIAN

The marine Cambro-Silurian sequence of the Oslo Region (Størmer 1953, Henningsmoen 1960) is relatively complete but rather thin (max. about 1,400 m). It consists predominantly of shales and limestones with some arenaceous and conglomeratic beds. In the central and southern parts of the region, the marine sequence is succeeded by a rather homogeneous and thick formation (1,250 m) of Old Red facies, the Ringerike Group (Turner 1974), which is supposed to be of Late Ludlovian to Downtonian age.

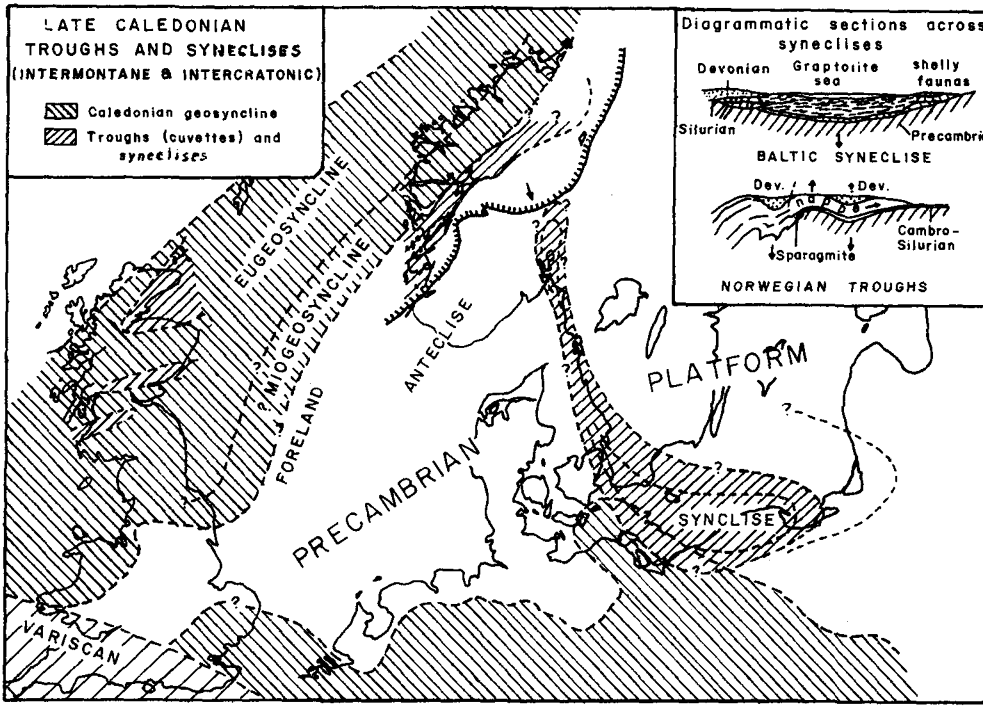
The Cambro-Silurian sediments of the Oslo Region were deposited on the foreland to the east of the main Caledonian geosynclinal belt (Bailey & Holtedahl 1938, Størmer 1967). Stratigraphical sections have revealed an increased thickness of the deposits along the axis of the Oslo Region as compared to the flanks. This basin-like structure developed in the Middle Ordovician (Skjeseth 1952) and especially through the Upper Silurian (Størmer 1967). Similar depressions occur in Scania, Poland and the Baltic, and they have been tentatively combined with the Oslo trough into one single intercratonic aulacogen or syneclise, Fig. 4a (Størmer 1967). An alternative connection is shown in Fig. 4b; the NW–SE extension of an aulacogen or trough along the Fennoscandian border zone (Poland-Scania-Jutland-Skagerrak) has been tentatively suggested on various grounds (Lindstrøm 1960, Bartenstein 1968, Spjeldnæs, pers. comm. 1971). This connection seems to fit the available borehole data in Denmark (Sorgenfrei & Buch 1964) just as well, but remains eventually to be tested by the more recent deep drillings in Denmark and off-shore yet to be published. The trend coincides with the proposed zig-zag pattern of the late Paleozoic (Permian) and Mesozoic faults (Ramberg 1971), and follows the Danish Embayment from the Polish Depression into the Skagerrak. The taphrogenic development of the Oslo Region and Scania seems to have been preceded by the formation of sedimentray troughs that may be traced back even to the Cambrian when the same areas were covered by a narrow, stagnant sea (Henningsmoen 1952, Thorslund 1960). The occurrence of a possible faulted basement topography below the basal Cambrian deposits in the Oslofjord area (Spjeldnæs 1955) also adds to the picture of an early tectonic activity within the area of the Oslo Region.

The Cambro-Silurian beds of the Oslo Region were folded during the Caledonian orogeny. The folding, which was most intense in the northern part of the region, dies out south of Drammen. The axial direction of the folds curves in a very general way from about E–W in the southern and western parts to almost N–S in the northeastern part of the deformed area. The Cambro-Silurian strata are only very slightly metamorphosed, if at all, disregarding the contact metamorphic aureoles above and around the Permian intrusives (Goldschmidt 1911). The beds have been affected by Permian faulting and jointing throughout the whole region.

2.4 PERMIAN SEDIMENTARY ROCKS

The Permian sedimentary rocks occur as: 1) a basal sedimentary series overlain by a thick succession of volcanic rocks; 2) thin, semiconformable beds within the volcanic series, marking intervals of erosion in between the extrusions of individual lava flows; and 3) rather thick local occurrences, e.g., in the Alnsjø area (argillite, arkose and conglomerate of predominantly volcanic material; occurring high up in the volcanic series within the Nittedal cauldron), the Brumunddal area (sandstone), and the outer Oslofjord area (rhomb-porphyry conglomerate).

The basal series (the Asker Group) was deposited on a rather even sub-Permian peneplain cutting through the various strata of the underlying, folded Cambro-Silurian sequence (Holtedahl 1953, 1957, Oftedahl 1960, Dons & Gyøry 1967). It has been subdivided into three formations: the Kolsås, the Tanum and the Skaugum formations (Henningsmoen, pers. comm. 1973). The most complete sections (about 20–30 m) are found at Asker and Lier a.



b.

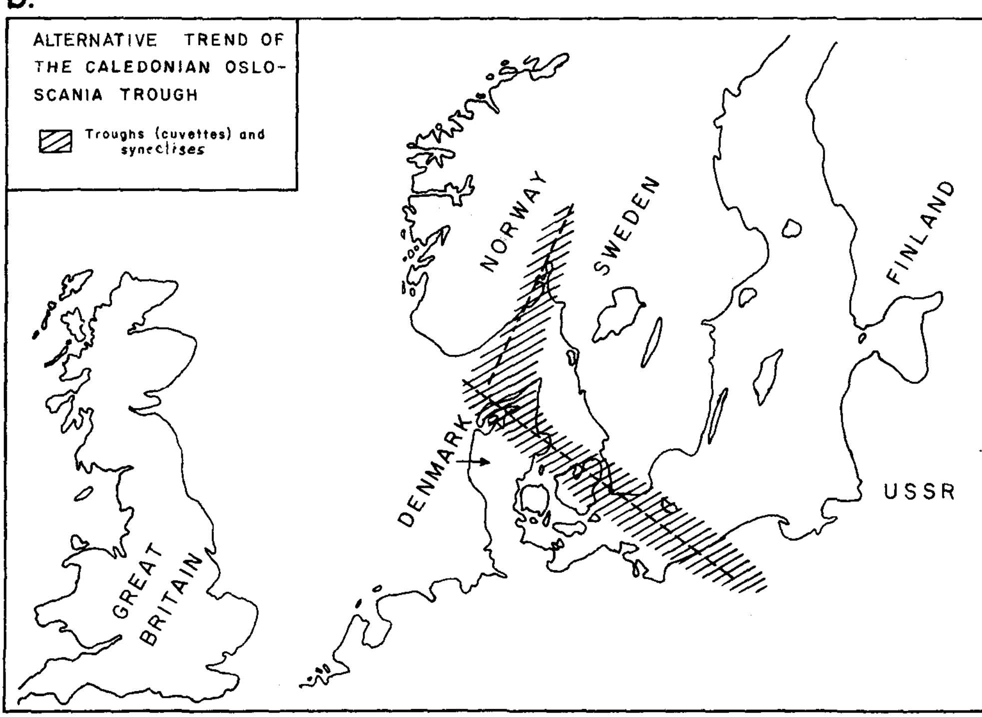


Fig. 4. Late Caledonian troughs or syneclises, (a) *after* Størmer (1967), (b) alternative trend.

southwest of Oslo, while at Brumunddal, about 110 km north of Oslo, rhomb-porphyry lavas rest directly on eroded Silurian strata. The thickest basal sections (about 50 m) occur in the Skien district. The agglomeratic and tuffitic Skaugum Formation is missing at Kolsås (near Oslo) and eastward, but constantly increases towards the south where the eruptions possibly were initiated.

There is ample evidence that the age of deposition of the Asker Group is equivalent to the time of initiation of the Oslo volcanism. Thus at Holmestrand fossiliferous shales occur above the lowermost lava flows (Holtedahl 1935). At Billingstad, thin beds of basaltic lava are present in the Tanum Formation quartz conglomerate (Brøgger 1933b). Conglomeratic or agglomeratic beds within the Skaugum Formation usually have pebbles or boulders of a type of porphyritic basalt (with plagioclase phenocrysts) that is known from the oldest basalt series in the southern part of the Oslo Region. To the north, in the Lundbergkollen and Skurvekampen supracrustals (at Skrukkelia NW of Hurdalen), the lava sequences rests conformably on dark grey, basal shale and sandstone. The sandstone contains 10–40% of dark minerals (chlorite, amphibole, etc.) and is rather similar to some of the sandy beds frequently occurring on the top of the lowermost basalt (B1) in more southern regions (Holtedahl 1935, Sæther 1946, Dons 1956a). The Skrukkelia basal sediments (0.5–3.5 m) are probably derived from already extruded basaltic flows nearby.

Fossils found in the upper part of the Tanum shales at Semsvik, Asker (Holtedahl 1931) are plants (Høeg 1935, 1937 a, b), remains of freshwater fish (Heintz 1934) and molluscs (Dix & Trueman 1935). The plant community is typical of Lower Permian to possibly middle Lower Permian, but many other fossils are related to Carboniferous forms. The fish remains indicate that the beds belong to the transition between Carboniferous and Permian.

The remarkably thick (ca. 1000 m) rhomb-porphyry (RP) conglomerates and sandstones occurring on a number of islands along the main fault zone in the southeastern part of the Oslofjord are interpreted (Størmer 1935) as fanglomeratic beds accumulated after the formation of a fault escarpment to the east. If they are situated on top of a Cambro-Silurian sequence of average thickness (and including the Ringerike sandstone), the occurrence is indicative of a possible subsidence of 2–3000 m or more in the outer Oslofjord area. The age of the RP conglomerates is unknown and may extend into Triassic time. The same age has tentatively been suggested by Spjeldnæs (1972) for the sandstone resting on RP flows in Brumunddal (Fig. 2).

2.5 PERMIAN IGNEOUS ROCKS

a. *Age*

The Lower Permian age indicated by the fossils within the basal sedimentary series corresponds to about 270–280 m.y. on Kulp's (1969) time scale. A number of radioactive datings from the igneous rocks are consistent with this (Neumann 1960, Broch 1964). Rb-Sr whole rock analysis (Heier & Compston, 1969) has indicated an age of about 276 ± 5 m.y. for the intrusion and crystallization of the plutonic series; these authors suggest that the crystallization of the plutonic rock series occurred during a remarkably short time interval. The same conclusion has previously been advocated for the volcanic series too (Oftedahl 1967).

From the volcanic rocks only one radiometric dating is known to the author: 284 m.y. on a basalt from Vestfold (Heier & Compston 1969). Paleomagnetic studies of volcanic rocks from the southern part of the Oslo Region have revealed only reversed remanent magnetism and a Permian pole position (Everdingen 1960), while the apparent age of the metamorphism of a diabase dike is 219 ± 6 m.y. (Dons 1974). This, and other work in progress, suggests an extended or periodic volcanic activity into Mesozoic times, and would correlate with similar activity and graben formation in the North Sea area.

b. *Volcanic rocks*

The Oslo volcanic series consists of abundant rhomb-porphyry flows (ca. 60% of the present total volume), some basalt flows (ca. 25%), and to a lesser extent trachytic and rhyolitic flows (Brøgger 1882, 1890, 1931a, 1933b, Rosendahl 1929, Holtedahl 1943, Sæther 1945, 1946, 1962, Oftedahl 1946, 1952, 1953, 1957a, b, 1959, 1960, 1967, Everdingen 1960, Weigand 1975). Today only remnants of the original cover occur, mainly in the Krokskogen and Vestfold areas and within subsided ring complexes. The present volume of lavas is about 300 km^3. The average lava thickness in the Vestfold area south of Oslo is considered to be of the order of 1 km. From the Krokskogen-Øyangen area, Sæther (1962) operates with a total thickness of more than 3 km. The latter estimate may be too high since recent studies conflict with the stratigraphic correlations in the upper part of the sequence (B. T. Larsen, pers. comm. 1972). Rhomb-porphyry dikes far outside the present Oslo Region (Oftedahl 1952, fig. 17), various rhomb-porphyry inclusions in the Sevalrud explosion breccia in the Precambrian about 30 km west of the Oslo Region (Brøgger 1931b), and rhomb-porphyry fanglomerates along the eroded Oslofjord fault escarpment (Størmer 1935) imply that the area covered by volcanics originally exceeded the present Oslo Region (ca. 200 x 40 km), and may possibly have been as much as 300 x 150 km (Oftedahl 1952, 1960). In more recent and less eroded rift regions like the East African rifts, vast amounts of volcanics extend far outside the main areas of subsidence (Baker et al., 1972, McConnell 1972). With an *average* total thickness of the original volcanic pile of only 0.6 km in the Oslo Region and vicinity, the *possible* lava volume may therefore have been 25–30,000 km^3. Such high values were also discussed by Oftedahl (1952) who, however, suggested a probable original lava volume of about 6,000 km^3. Recent discoveries of thick sections of stratigraphically higher units, especially rhyolites, within the Ramnes cauldron subsidence (Oftedahl 1967) and considerable thicknesses (ca. 1,000 m) within small, subsided (cauldron?) fragments in the northern part of the region

(Hurdal-Skrukkelia), lend credence to the view that the original volume was somewhere in between the above limits, that is of the order of 10,000 km^3.

The stratigraphy of the volcanics in various parts of the Oslo Region has been summarized by Oftedahl (1960, 1967). Basaltic rocks (B), rhomb-porphyries (RP), and higher up in the series also ordinary latites (L), trachytes (T) and rhyolites (R) occur repeatedly alternating. The frequent alternation of compositionally contrasted lava flows indicate the simultaneous existence and availability of two or more independent magma types throughout the volcanic period in the Oslo Region. However, despite great lateral compositional variations within the basalt units, there is an overall tendency from bottom to top for a change to a more acidic composition of the volcanics (Oftedahl 1967).

The basaltic rocks subdivide into four main classes according to their phenocryst mineralogy: aphyric basalts, pyroxene basalts, plagioclase basalts and pyroxene-plagioclase basalts. Basalts, *sensu stricto*, are rare; based on the normative mineral content most of the basaltic rocks plot as trachy-andesites or trachy-basalts (applied according to Streckeisen 1967, p. 185). Stressing the overall similarity of the Oslo basaltic rocks to those from other alkaline provinces, a different terminology is appropriate; the Oslo basalts range from ankaramites via alkali olivine basalts and hawaiite to mugearite.

The rhomb-porphyries, well known for their characteristic, often rhomb-shaped anorthoclase-plagioclase phenocrysts, occur in a number of regional and local flows. From the rather few major-element analyses published (Brøgger 1933a, Oftedahl 1946), they represent a relatively uniform series of latitic (or trachyandesitic) composition. They appear to define a unique rock group on a worldwide scale, and recent whole rock and phenocryst analyses (Brunfelt, Finstad & Larsen, pers. comm. 1973) show that the REE rock-to-chondrite pattern is almost identical for the RP flows in the Oslo, East African and Antarctic (Ross Island) occurrences.

The various other latitic, trachytic and rhyolitic porphyries have been considered as intrusive sill-like bodies (Brøgger 1890, Holtedahl 1943, Oftedahl 1946, Sæther 1962, and others), but reinterpreted in later works by Oftedahl (1957a, 1960) and Huseby (1968) as extrusives. In the Skrukkelia area, where a number of thin porphyry beds alternate with basalt flows and sedimentary deposits, the effusive character of the porphyries is convincingly demonstrated.

The extrusion of the Oslo volcanics has been principally divided into three phases (Oftedahl 1960): two periods of plateau eruptions interrupted by a period of central eruptions. During the periods of plateau eruptions the lava flows generally derived from fissures (mainly the rhomb-porphyries) and from multicenter eruptions (basalts). The central-eruption period is characterized by local and irregular flows. The (B3) basalts erupted in great number as thin flows from many foci, building up large basalt volcanoes. Associated with the B3 basalts were trachytic and rhyolitic lava flows and ash flow tuffs. The acidic eruptions seem largely to be connected with cauldron volcanic activity (Oftedahl 1960).

The Oslo volcanics are cut by a number of faults. On the whole, or at least in the southern and central parts of the region, the outburst of the lava flows preceded the initial block faulting of the region. The regional extent and the lateral change in thickness of the bottom basalt may signify that the B1 flows erupted onto a surface that was rather even. Remnants of extrusives far outside the present graben outline are consistent with this interpretation but might also be due to eruptions from fissures outside the graben. Towards the north, however, in Skrukkelia, the basal sandstone and basalt is overlain by clearly different RP flows within a distance of less than 3 km. This indicates either a possible faulted topography at the time of the RP extrusion in this region, or that we are dealing with strictly local flows, or both. All evidence taken together, the volcanism began in the south with the eruption of great thicknesses of alkali olivine basalts in the southern districts, propagated northward, and finally reached the northern districts with thin (tholeiitic) basalt flows at a stage of more advanced tectonic deformation by rifting.

c. *Plutonic Rock series*

In his systematic petrography of the plutonic rocks Barth (1945) suggested a subdivision into two major rock suites or families:

I. The 'Oslo-essexites' (Old volcanic necks), comprising rocks generally of ultrabasic to monzonitic composition, and
II. The main Kjelsåsite-Granite series (Suite of deep-seated or subvolcanic rocks), comprising rocks of dioritic to granitic and nepheline-syenitic composition.

The volcanic necks, which exhibit a great petrographic variability and complexity, are chiefly made up of gabbroic rocks of alkaline affinities. About 13 major and some minor neck sites are known throughout the Oslo Region, six of which are double or multiple necks (Fig. 5). The necks (usually with a diameter of 0.5–1 km) are situated along major tectonic lines: one along the western border of the Oslo Region in a NNE–SSW direction from Aurenhaugen (AU) to Eiangen (EI), another in a N–S direction from Brandbukampen (BR) in the north to Torbjørnskjær (TB)* in the south. The last row is subparallel to the major boundary faults along Randsfjorden and along the east side of the Oslofjord. The necks are generally believed to represent feeder channels for the extensive basaltic flows.

The emplacement of the volcanic necks supposedly preceded that of the main plutonic series and most necks are found in border regions outside the main plutonic rocks. Exceptions are the Eiangen and the small Mylla occurrences, both of which are regarded as inclusions within the younger intrusives. At Kjelsås in Sørkedalen (SØR, Fig. 5) a gabbroic to monzonitic, coarse-

* Identified from thin-sections made from rock samples collected by Dr. S. B. Smithson.

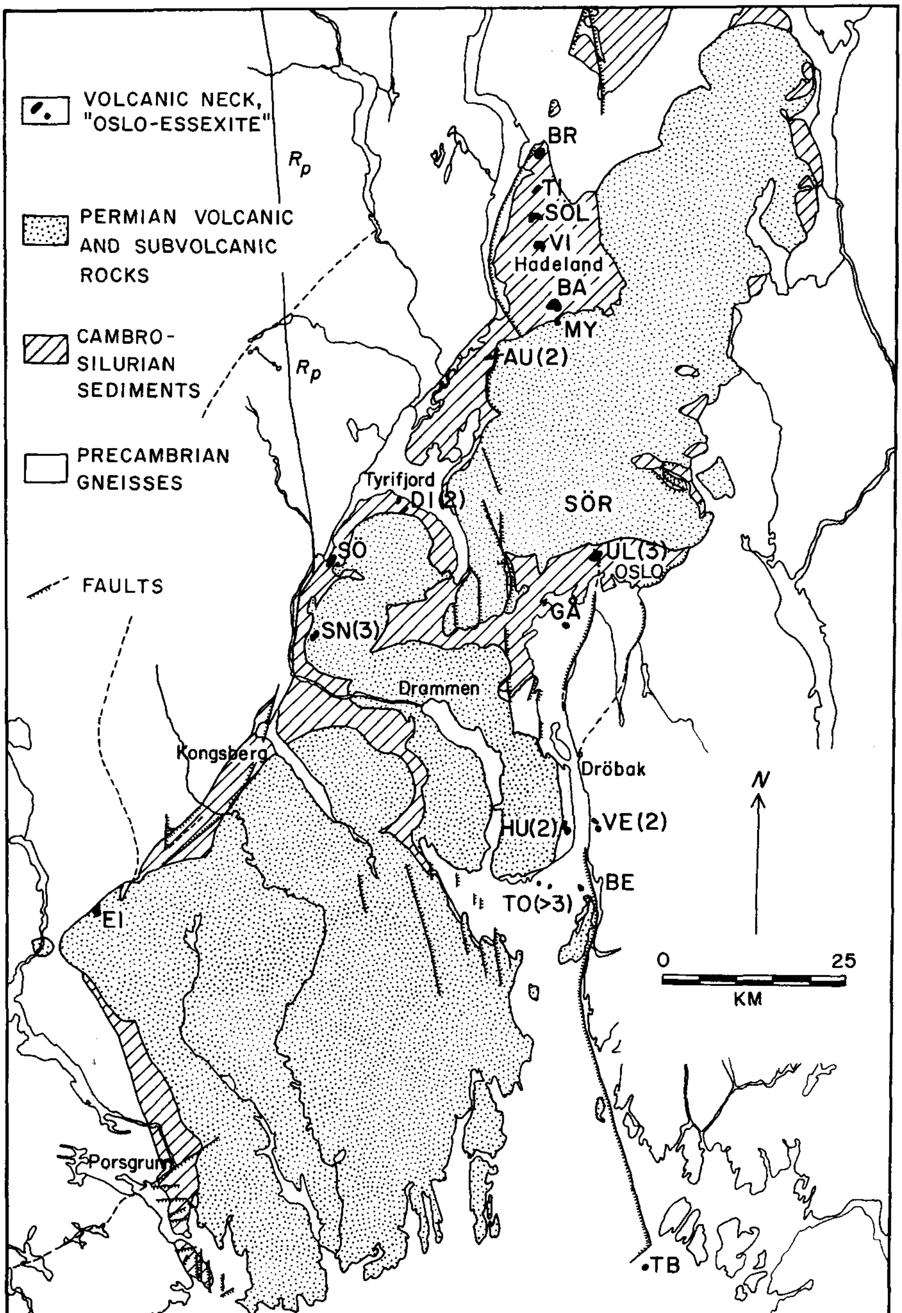


Fig. 5. Location of mafic volcanic pipes ('Oslo-essexites') in the Oslo Region. Major occurrences: BR = Brandbukampen, SØL = Sølvsberget, VI = Viksbergene, BA = Ballangrudkollen, UL =Ullernåsen, VE = Vestby, HU = Husebyåsen, TO = Tofteholmene, AU = Aurenhaugen, SØ = Sønstebyflakene, EI = Eiangen. Some minor occurrences: TI = Tingelstad, MY = Mylla (inclusion) DI = Dignes, SN = Snaukollen, GÅ = Gåsøya, BE = Bevø, TB = Torbjørnskjær (?). SØR, indicates the location of the Sørkedalite occurrence.

Table 2: Main rock-types in 'Oslo-essexitic' necks*

Rock group	Local name(s)	Characteristic mineral(s)	Streckeisen (1967) Classification
Ultramafic rocks (mafitites)	*Pyroxenite* (Earlier terms: olivine-yamaskite, yamaskite, etc.)	(Ti-) Augite and/or diopsidic pyroxene, olivine	Pyroxenite
Monzo-gabbro and Monzo-diorite	a. Olivine gabbro (Yamaskite-essexite, ol. – gabbro diabase, etc.)	(Ti-) Augite, Olivine, (py)	Monzogabbro to gabbro
	b. Bojite (Essexite-gabbro)	Alkaline hbl., (py), (bi)	
	c. *Kauaiite* (Essexite-gabbro, Biotite-essexite, Essexite, etc.)	Augite, olivine, (bi)	Diorite, Monzodiorite to Monzonite
	d. Mafraite (Essexite-gabbro, Essexite)	Alkaline hbl., (py), (bi) a–d: strongly zoned plag.	
Monzonite	Akerite	Often rectangular plag., Rel. fine-grained	Monzonite

* See Barth (1945), Brøgger (1931, 1933b), Dons (1952), Ramberg (1970).

grained plutonic body occurs engulfed in the medium-grained monzonites (akerites). A genetical relationship between the sørkedalite and a hypothetical, old, recrystallized, gabbroic neck has been suggested by Barth (1945), Sæther (1962) and others, but has been opposed by Bose (1969).

A number of rock-types occur within the volcanic necks, and a variety of names has been assigned to the different rock-types. Brøgger (1898) proposed the term essexite for the most common petrographic type. Since none of the basic rocks in question normally carries nepheline, a reclassification was offered by Barth (1945) who also suggested the name 'Oslo-essexite' as a loose term for all of the rocks found within the volcanic bosses. The main rock-types are listed in Table 2. The commonly occurring kauaiites (or 'essexites') have a wide compositional range from monzonitic to monzo-dioritic, and could generally be referred to as syenodiorites (Streckeisen 1967).

Less frequent are bojite (diorite), mafraite (monzodiorite), modumite (anorthosite of bytownitic composition) and akerite (monzonite), the last one often occurring as a border type. Rare differentiates are cumberlandite (olivine-ilmenite rock), husebyite (nepheline-syenite), hurumite and windsorite (syenites); see also Sørensen (1974, p. 569). Fine-grained, porphyritic varieties occur in places, as do pyroxenite-magnetite rocks (Ramberg 1970).

The various rock types of the 'Oslo-essexites' combine to form a complete rock series which may have formed by magmatic differentiation (Barth 1945). A preponderance of various basic rocks, the olivine-gabbro to kauaiite stem,

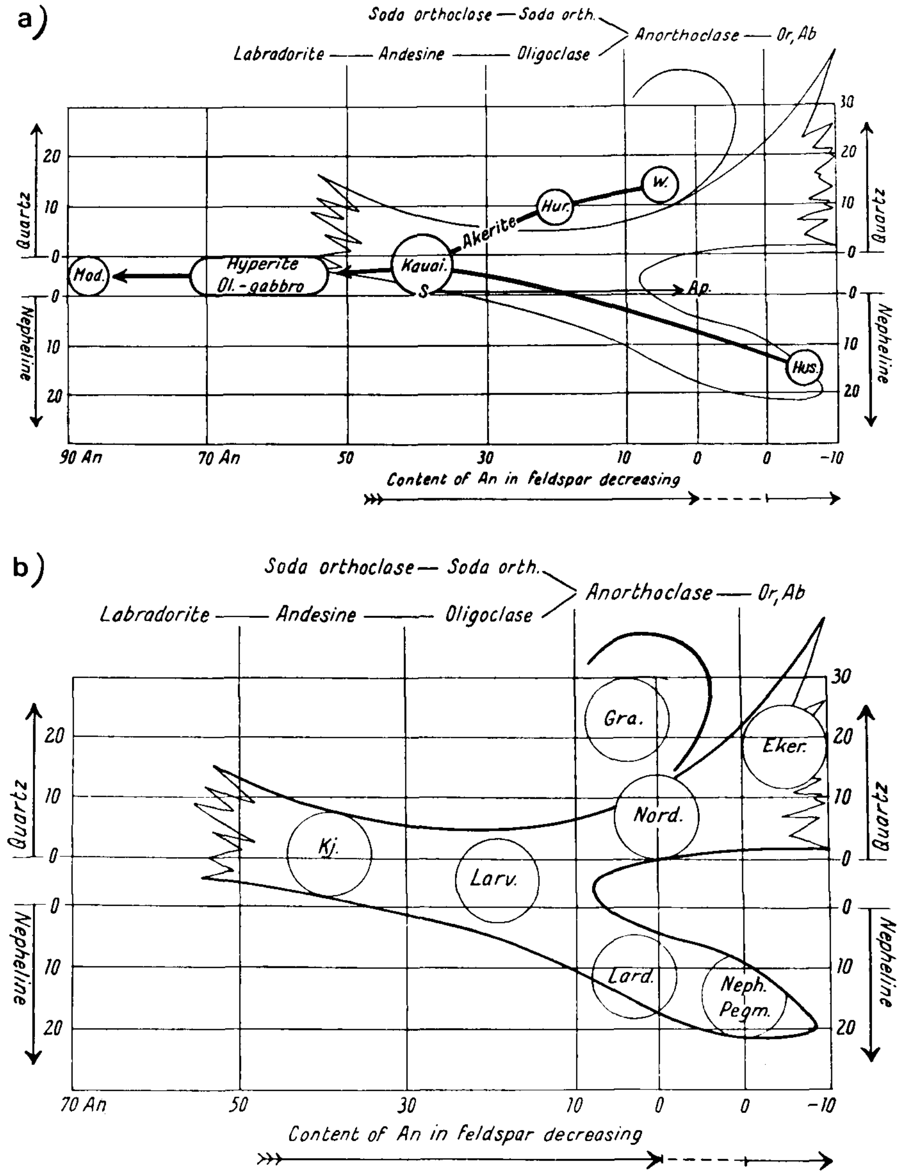


Fig. 6. Barth's (1945) 'Family Trees' showing the systematic position and mutual relationship of the principal Oslo rock types. (a) 'Oslo-essexite' series, (b) Main Kjelsåsite-granite series, indicating course of differentiation Mod = Modumite, Kauai = Kauaiite, Hur = Hurumite, W = Windsorite, Hus = Husebyite, Kj = Kjelsåsite, Larv = Larvikite, Nord = Nordmarkite, Eker = Ekerite, Gra = Granite, Lard = Lardalite; see Table 2 and 3.

branches out into the more leucocratic derivatives of minor volumetric importance (Fig. 6a).

A concentric arrangement of the rock-types commonly occur within the necks, as for instance in the Hurum and Vestby occurrences. Less regular spatial arrangements are encountered in other necks, e.g., Ullern-Husebyåsen

(Dons 1952) and Brandbukampen. Platy flow or rhythmic layering occur in some of the necks, sometimes defining a concentric, inward-dipping, cone-shaped pattern (Dons 1952). In some cases, for example in the Knalstad neck at Vestby, the layering close to the border has a vertical to steep outward dip, indicating an increased radius of the neck with depth.

The deep-seated plutonic series (main Kjelsåsite-Granite series) which occupies about 60% of the present graben surface is found within three subregions (Fig. 1): I) a northern region, the Nordmarka-Hurdalen batholith, of largely syenitic to granitic rocks, II) a central region predominantly consisting of biotite granites, the Drammen and Finnemarka complexes, and III) a southernmost region with a predominance of monzonitic rocks, the Larvik and Skrim plutonic complexes.

Each subregion contains one or more composite batholiths with a variety of rock-types. Frequently the rock-types merge into one another by gradational transitions. 'Simultaneous' contacts with no chilled margins are described between many plutonic rocks (Brøgger 1898, Oftedahl 1960). Intrusive contacts are, however, also commonly observed (e.g., Sæther 1962) and, for instance, in the Hurdal district, younger acidic intrusives follow brecciated and mylonitic border zones in large syenodioritic inclusions (Nystuen 1975b). The multitude of contact relations revealed implies a complex sequence of emplacement of the different main and transitional rock-types.

The plutonic series consists of granular, mostly coarse-grained, massive rocks. The subvolcanic character of the suite is frequently revealed by textural features such as zoned plagioclase, and a porphyric or pseudoporphyric appearance (Barth 1945). Similarly, the many inclusions and the mineralogy of the miarolitic cavities and druses in the plutonic rocks, especially in the border zones but also in more central parts (Raade 1969, 1972), indicate that the present surface is close to the original roof of Cambro-Silurian and Downtonian rocks. The thickness of the original rock cover has been estimated to about 2 km (McCulloh 1952).

The chief rock-types of the plutonic suite belong to the following four groups: 1) syenodioritic (mostly monzonitic) rocks, 2) nepheline-bearing rocks, 3) syenites, and 4) granites. A survey of the classification of the Kjelsåsite-Granite series is given in Table 3. For more exhaustive information about the main rock-types, see the summary in Appendix 1, and the papers by Barth (1945), Oftedahl (1946, 1953, 1960, 1967), Raade (1973) and Nystuen (1975a, b).

Barth (1945) suggested that the principal types of the Oslo plutonic rocks belong to a series that generated by fractional crystallization of a syenitic magma (Fig. 6b). The biotite granite has an uncertain position in the genetic scheme (Barth 1945) and has been thought not to be directly related to the principal series through a simple magmatic differentiation.

The age relations inferred from Barth's 'family tree' have in general been confirmed by field studies, and already Brøgger (1890) concluded that the

Table 3: Survey of Oslo Region deep-seated rock suite (Kjelsåsite-Granite series). Chief rock-types in cursive. *After* G. Raade (1973)

Rock group	Local name	Characteristic mafic minerals	Streckeisen classification
Diorites and syenodiorites	Sørkedalite[1]	Olivine (diopside, biotite)	Diorite
	Akerite[2]	Augite, hornblende biotite	Syenodiorite (monzodiorite and monzonite)
	Kjelsåsite	Augite, biotite (amphibole)	Monzonite to monzodiorite
	Larvikite[3]	Augite, biotite (amphibole)	Mononite to syenite
Nepheline-bearing rocks	Ditroite	Augite, barkevikite, biotite	Foyaite
	Lardalite	Pyroxene, biotite	Plagifoyaite?
	Foyaite-Hedrumite	Aegirine to diopside	Foyaite to alkali syenite
	Nepheline syenite (pegm.)	Aegirine, biotite barkevikite	
Syenites	Hedrumite ring-dike	Augite, biotite amphibole	Alkali syenite
	Pulaskite	Diopside to augite, biotite, amphibole	Alkali syenite
	Grefsen syenite	Biotite, hornblende	Syenite
	Normarkite	Alkali amphibole, aegirine (augite, biotite)	Alkali syenite
	Transitional syenites	Augite, amphibole, biotite, etc.	Syenite or alkali syenite
Granites	*Ekerite*	Alkali amphibole, aegirine	Alkali or soda granite
	Drammen granite[4]	Biotite	Granite
	Finnemarka granite	Biotite	Alkali granite
	Other biotite granites	Biotite	Granite or alkali granite
	Albite granite		Albite granite

[1] Including 'apotroctolite' (olivine syenodiorite) of Barth (1945).
[2] Including monzodiorite from Finnemarka (granodiorite of Czamanske 1965), syenodiorite from Hurdal (Nystuen 1970), and akerite-kjelsåsite transitional rocks from Nordmarka (Sæther 1962).
[3] Including tønsbergite and transitions into extrusive rhomb-porphyry.
[4] Including rapakivi granite.

rocks of the volcanic necks are the oldest and that increasing acidity seems inversely proportional to age. This simple sequence does not hold for the volcanic series, nor for the dike rocks (Oftedahl 1960, Dons 1952), and recent

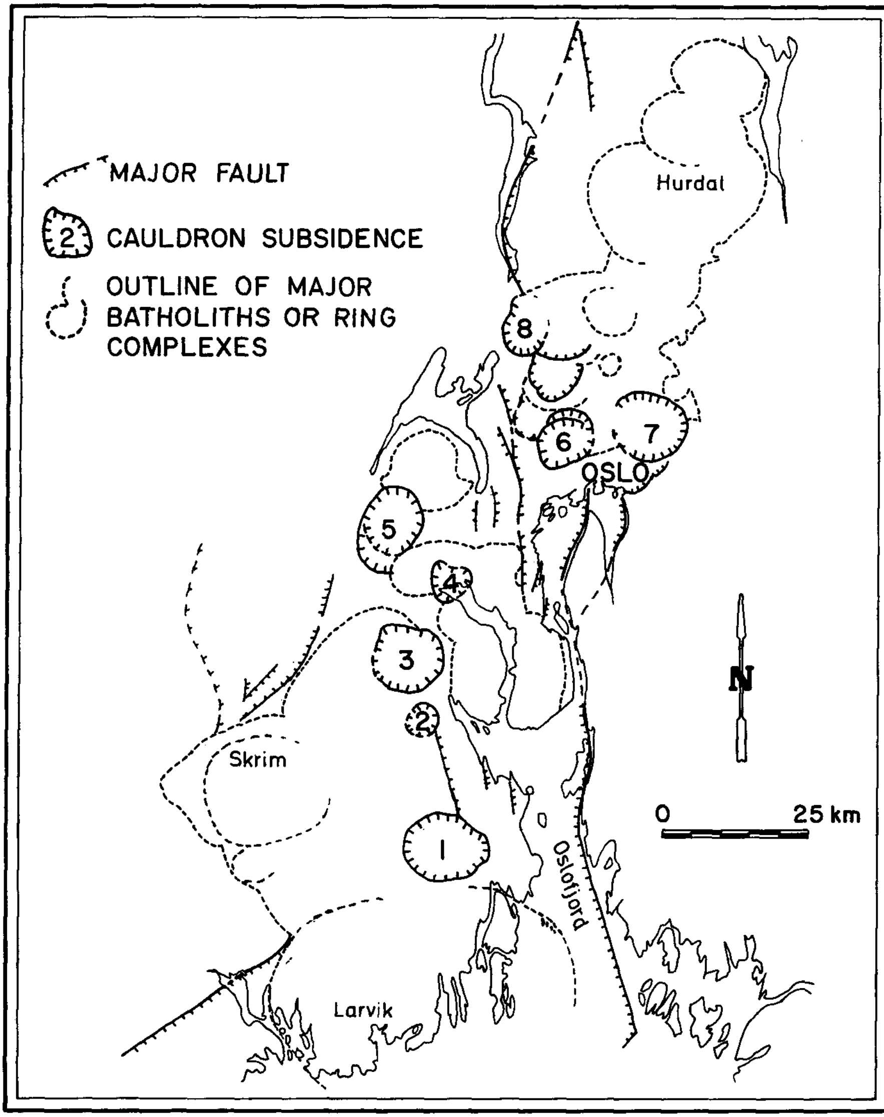


Fig. 7. Cauldron subsidences and intrusive complexes in the Oslo Region. 1. Ramnes, 2. Hillestad, 3. Sande, 4. Drammen, 5. Glitrevann, 6. Bærum, 7. Nittedal, 8. Øyangen.

studies (e.g., Nystuen 1975b, A. Gaut, pers. comm. 1972) indicate that more than one *intrusive* cycle exists.

The large plutonic massifs were regarded as laccoliths by Brøgger (1898, 1933a), a conclusion he arrived at by studying the outward-dipping and apparently concordant contact between the biotite granite and the roofing Ordovician shales at Hørtekollen north of Drammen. Most commonly, however, we have steep and cross-cutting contact relations, as seen at a number of places both in the northern and in the southern parts of the province, and Holtedahl (1934, 1935) concluded that the intrusive bodies generally have

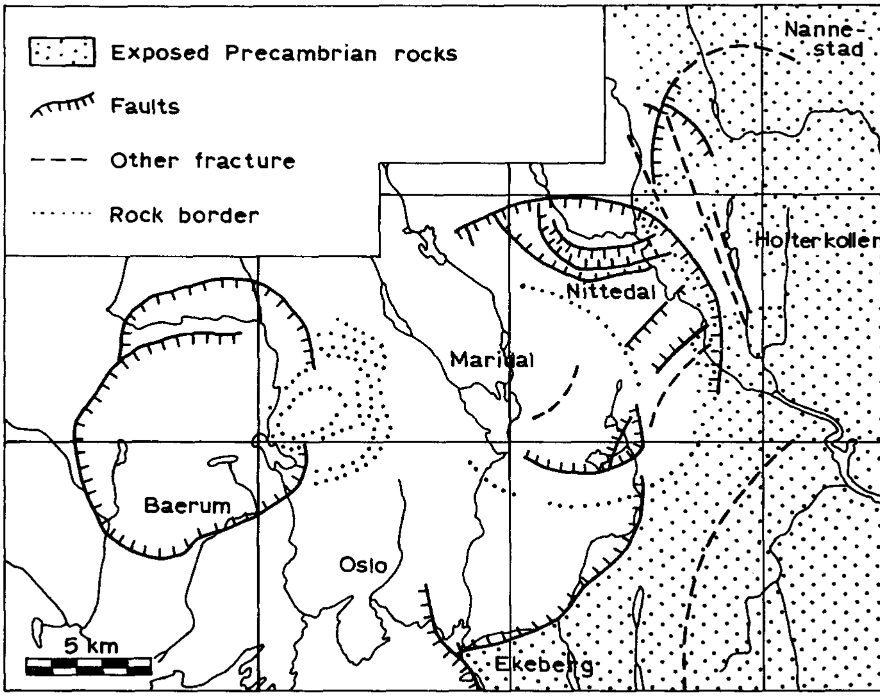


Fig. 8. Ring faults and fractures in a central section of the Oslo Region (*after* Naterstad 1971).

a stock-like or *batholith-like* form, a view widely supported by later studies.

The plutonic masses have thermally metamorphosed the adjacent sedimentary rocks. For the inner contact zone Goldscmidt (1911) subdivided the metamorphic products into 10 hornfels classes (depending on the original chemical composition). The broad contact aureoles have, together with other criteria (McCulloh 1952, p. 45–46, Sæther 1962), led to the assumption that the rising acidic masses were superheated.

The general outline. The Oslo Region is defined by (1) a system of ring-shaped fractures or boundary lines, and (2) the interaction of two principal sets of linear elements (NNW to N and NNE to NE). The *ring-shaped* features are demonstrated by many of the plutonic bodies as expressed by the ring-like boundary lines of the larvikite plutons, the Drammen and Finnemarka granites, in the repeatedly half-circle shaped western boundary of the northern syenitic batholith, etc. Typically for the Oslo Region, many of the batholiths are composite, concentrically built plutonic masses, commonly arranged with the most basic rocks as marginal zones. A number of ring complexes have been formed by cauldron subsidence (Fig. 7). A related origin, cauldron subsidence or 'ring fracture stoping', has also been suggested for the other simple or composite circular complexes (Holtedahl 1943).

Outside the areas of main intrusion, ring-shaped fractures, partly associated with Permian dikes or hydrothermal mineral occurrences, are known, e.g., in the Nittedal-Nannestad area (Naterstad 1971) and in the Ski area, indicating that the ring-fault forming forces were not restricted to the graben region proper (Fig. 8). The cone-in-cone features exhibit characteristics which indicate an origin from simple axial stress as caused by initially rising magma masses. Furthermore, the batholiths in the Oslo Region rose to the present surface largely by the process of overhead, piecemeal stoping as demonstrated by the abundance of country rock inclusions, especially closer to the borders, in all the plutonic rocks. The xenoliths vary in size from several kilometers down to a few centimeters and even to xenocrysts. Their regular distribution and mode of descent have been well illustrated by, e.g., studies in the Alnsjø area (Holtedahl 1943, see fig. 22). At many places the sedimentary rocks and lavas close to the contacts seem to have fallen into the plutons (Brøgger 1890, Holtedahl 1935, Oftedahl 1952, 1960, Rohr-Torp 1969, 1973); Fig. 9a. This can be due either to gravity sinking of the supracrustals into the usually less dense intrusives or to subsidence of the plutonic rocks after intrusion, or both.

The *linear* trends are recognized in the directions of faults, fractures, geological contacts and dikes. The downward movement on the graben side of the lineaments is demonstrated by the numerous normal faults and antithetic step faults as well as by the nature of Permian/Precambrian contact zones; for instance, in an area between the south-eastern end of the lake Hurdalssjø and Mistberget. Here the nordmarkites are bordered by Precambrian gneisses along an almost straight NE–SW line for a distance of about 5 km. Yet the border is fringed by elongate, steeply dipping, Cambro-Silurian hornfels xenoliths in the nordmarkite, their strikes always being subparallel to the nordmarkite/gneiss contact. The lack of Cambro-Silurian rocks on top of the adjacent Precambrian, the restricted occurrence of the hornfelses in a narrow zone close to the contact, and their spatial arrangement are evidence indicating a downward movement of the graben rocks at the border. No flow structures are observed in the nordmarkite, but the Precambrian is in places clearly brecciated.

The linear trend overprints the circular features, and defines the overall NNE–SSW extension of the Oslo Region. In Scotland similar cone-sheets are believed to have been intruded along shear-strain fractures (Robson & Barr 1964) or along tensional fractures (Anderson 1936, 1951). Whatever the origin of the pre-existing tectonic lines, the circular cones seem to be co-axial about a set of lines which may occupy a more or less vertical position above their source magma chambers.

The Permian (and later?) faults and fractures commonly follow older linear tectonic trends. Thus, the localization and outline of the intrusive complexes seem to have been guided by the combined effect of rejuvenated zones (planes) of weakness and younger concentric fractures associated with each center of intrusion. Nested into one another, the intrusions form the outline of the composite Oslo province (Figs. 2 & 8).

Recent rifts are quite normally associated with elevated rims and tilted graben shoulders. This is clearly not the case in the Oslo Region; as a matter

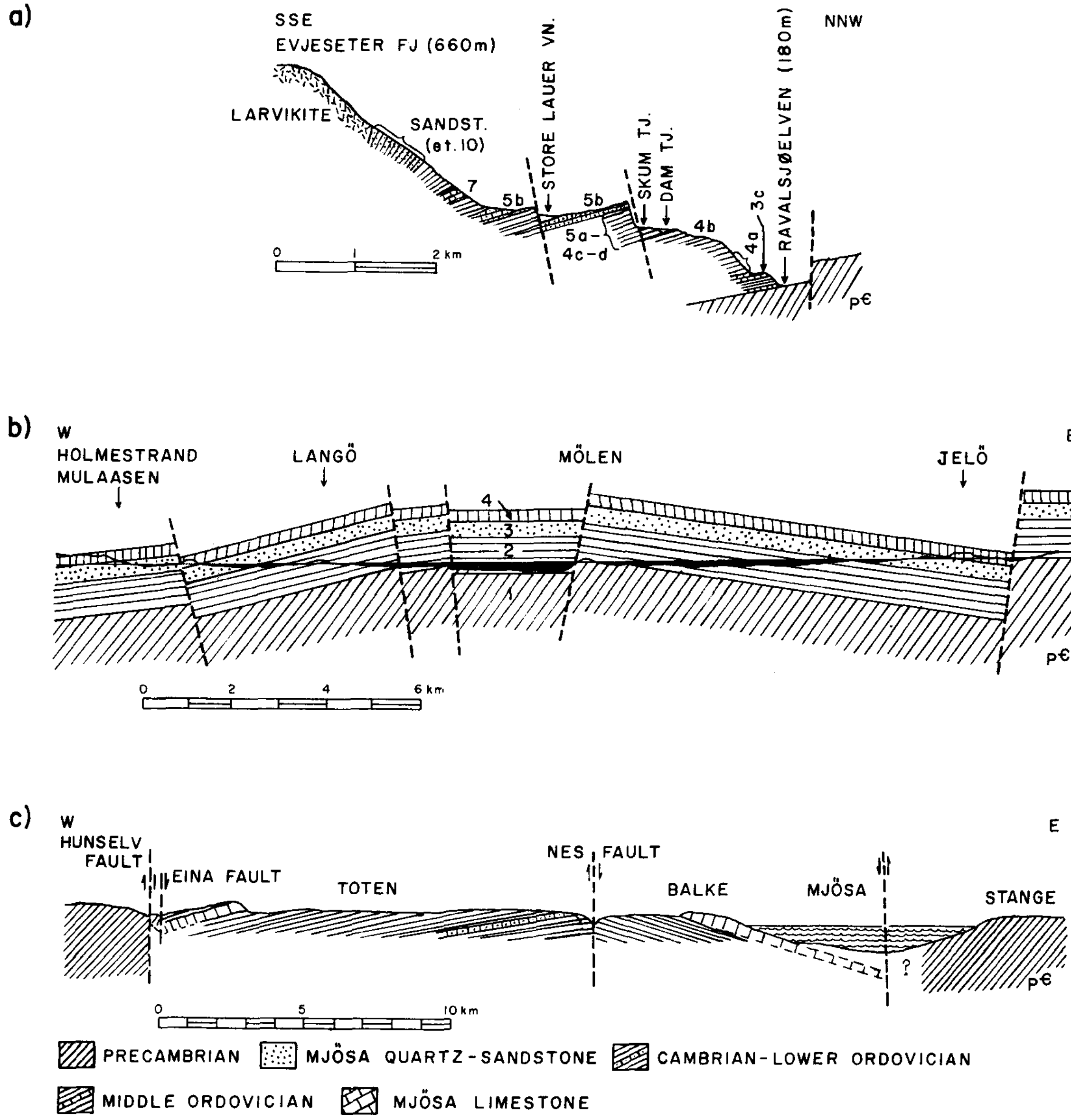


Fig. 9. (a) Geological section from the Skrim Mountain illustrating antithetic faulting and increased tilt of Cambro-Silurian blocks close to the plutons (*after* A. Heintz *in* Oftedahl 1960), (b) Schematic section across the Oslofjord demonstrating local doming and Basin and Range structure (*after* Brøgger 1886). 1 = Precambrian, 2 = marine Cambro-Silurian, 3 = Downtonian Sandstone, 4 = Lavas. (c) E–W section just north of the Oslo Graben proper showing crustal arching and normal faulting in continuation with the structural depression of the Oslo Region (*after* Skjeseth 1963).

of fact, the region occupies an area of depression in the sub-Cambrian peneplain (Strand 1960, fig. 52). The position of the sub-Permian peneplain outside the graben area is unknown. Signs of lifting and forceful intrusion are documented around some of the local intrusives like the Finnemarka granite

(Brøgger 1898, Oftedahl 1960) and the Ullernåsen volcanic necks (Dons 1952), and in the northernmost part of the region (Fig. 9c). Here, on the western side, the sub-Cambrian peneplain dips away from the intrusives at an angle of about 5°–7° (Holtedahl & Schetelig 1923, Holtedahl 1935). On the eastern side of the intrusives, basal conglomerates are found at about 500 m a.s.l. in Precambrian xenoliths close to the border, while a few kilometers to the east, at Feiring, the basal conglomerate occurs about 375 m lower (Nystuen 1975b, and pers. comm. 1973). In all the known examples of lifting and tilting in the Oslo Region, this occurs within the main tectonic lines of the graben area. This means either that wide zones of elevated rims never existed in the Oslo Region, that possible elevated rims were never detected in the Precambrian terrain, or that original temporary expansion due to concurrent anomalous thermal conditions have had time to return to normal since Permian times.

d. *Block faulting*

The Oslo Graben or Rift System is, like most continental rift systems, not unidirectional but composed of faults in several directions, subdividing the graben floor into a system of small-scale, basin-and-range-like structures. The faults are normal or antithetic step-faults, indicating a general extensional origin for the Oslo Graben. Although several trends occur, the main fault directions are NNW to N and NNE to NE.

Major boundary faults have a vertical displacement of the order of 1 km but occasionally 3 km or more as in the southernmost part of the Oslofjord (Brøgger 1886, 1933a, Holtedahl 1934, Størmer 1935, Oftedahl 1952, 1960, Holtedahl & Dons 1957). In other areas the subsidence is marked by antithetic step-faults and increasingly tilted bedrock blocks (see Fig. 9a). Major boundary faults are not always observed and may have been obliterated by subsequently rising magmas.

Minor faults occur throughout the graben and probably formed throughout the entire period of lava formation (Størmer 1935, Oftedahl 1952, 1967) since they cut even the youngest lava beds and the rhomb-porphyry fanglomerates which certainly post-date the onset of the major faults. The variety of dike rocks of varying relative age supports the existence of crustal deformation over a longer period of time. No large faults, however, have been observed within the plutonic rocks, indicating that the tensional deformation culminated no later than the crystallization of the batholiths.

Many of the faults are oblique-slip faults; measured horizontal components may amount to several hundred meters (Cloos 1928, Dons 1952). Faults and joints often display an *en échelon* arrangement.

Cloos (1928), emphasizing the predominance of the NNW- to N-striking faults, suggested that they were 'Fiederspalten' caused by a left-lateral movement of the crustal blocks on each side of the graben. The opposite movement was indicated by Størmer (1935) on the basis of the fault pattern in the outer part of the Oslofjord. Bederke (1966) pointed out that all the strike-slip and oblique-slip faults studied by him (mainly in the inner part of the Oslofjord) showed right-lateral displacement. Bederke concluded they were formed as the result of a clockwise rotational strain set up by differential movements of the adjoining blocks. However, both left-lateral and right-lateral movements have been recorded along both the NNW- and the NNE-striking faults from a number of localities, and a more detailed analysis is necessary to distinguish between and interpret the meaning of the different generations of faults that apparently exist. The sharp relief and continued seismic activity in the Oslofjord suggest that some faulting may possibly have occurred up to recent times.

e. *Dike rocks*

Igneous dikes are abundant in the Oslo Region and surroundings, especially in the Cambro-Silurian sedimentary rocks where they are also more easily spotted because of contrasted lithology and differential weathering. The dikes are petrographically highly variable, ranging from basic through monzonitic and syenitic to extreme siliceous compositions. They have been dealt with in a number of papers (Brøgger 1894b, 1898, 1933c, Sæther 1947, 1962, Dons 1952, Oftedahl 1957c, Antun 1964, Huseby 1971, Hasan 1971).

Some of the igneous dikes are clearly related to equivalent deep-seated rock-types, but in general the relations are more uncertain and many transitional dike rocks occur. The dikes are predominantly of dilatational or fissure origin. The main strikes are in the NNW–SSE and the NNE–SSW directions, the former direction possibly connected with the younger dikes (Huseby 1971). In addition to dikes, sills are also known; the most impressive are the extensive meanaite sills occurring at the base of the Cambro-Silurian sequence.

A total of at least 1650 dikes (Huseby 1971) has been described from the Oslo Region. They may occasionally constitute large parts of the present-day surface. Thus, in the Ullern-Husebyåsen area, Dons (1952) estimated that the igneous dikes cover some 17% of the map area. Similar estimates by Sæther (1947) from the lowland of Bærum give 4–6%, while in Østre Bærum the dikes may locally amount to about 18% (Werenskiold 1918). Dike rocks are also frequent in other areas, such as the Hadeland area (Brøgger 1933b), and in general the dike rocks represent a considerable volume increase of the Oslo Region.

Permian dikes also occur far outside the main area of intrusion (Oftedahl 1952). Long rhomb-porphyry dikes, in the continuation of the NNW and NNE fracture sets, intrude the Precambrian along the Skagerrak coast both on the Norwegian and on the Swedish side (Suleng 1919, Samuelsson 1971). A third major dike has been traced for about 87 km NNW-ward from the Modum area west of Oslo. Also, diabase dikes of possible Permian age occur in a wide aureol around the Oslo Region (Storetvedt 1966, 1968, Halvorsen 1970, 1972), predominantly in the SSW axial extension along the Skagerrak coast. The widely distributed occurrence of the dike rocks shows unequivocally that the Oslo magmatic activity was not restricted to the present graben area alone.

3. Formation of grabens in continental rift systems

Active continental rifts, such as those in East Africa, are generally considered to represent a part of the world rift system, the major part of which is confined to the oceanic floor. The basaltic volcanism and the fact that the rifts occur both in continental and oceanic crustal regions imply that their formation ultimately is connected with upper mantle rather than with crustal processes. In view of this it seems appropriate to note that even though the continental and oceanic rifts may represent various stages of the same process (Heezen 1960, 1969, Osmaston 1971), their development may have followed different courses, and some disagreement exists concerning the validity of a direct comparison between the two phenomena (e.g., King 1966, Girdler et al. 1969, McKenzie et al. 1970, Logatchev et al. 1972).

Even if *continental* rifts have a number of physical features in common, there is no universal agreement about their structure, mass distribution and formational evolution. The question is still open as to whether continental rifts are perhaps produced by several mechanisms. Contrasted models have been presented in the literature, involving different structural solutions and mass distributions.

For a long time, two rivalling lines of thought existed; the first assuming that the central block of the rift had been lowered relative to its shoulders between parallel normal faults due to tensional forces, possibly as a result of crustal doming (e.g. Suess 1904, Cloos 1939); the second considering that the floor of the rift is held down between thrust faults caused by compressional forces (e.g. Wayland 1930, Bullard 1936). Since then the compressional hypothesis has been disproved. The gravity anomaly, as obtained by the pioneer work of Bullard (1936) across the Lake Albert rift valley, is consistent with the tensional hypothesis as demonstrated by Girdler (1964). Borehole data (e.g. from the Rhine Graben, Illies 1967, 1970) have revealed the configuration of the inward dipping normal faults. Extension has been demonstrated by geodetic measurements of strain increments at the time of large earthquakes (Basin and Range province, Thompson 1959) while earthquake fault-plane solutions indicate that either strike-slip or normal fault mechanisms are operating in continental rifts (Balakina et al. 1969, Florensov 1969, Fairhead & Girdler 1972). Precision levelling across the Rhine Graben (Illies 1970, Kuntz et al. 1970) has revealed a continuing subsidence of the graben floor of about 0.5 mm/year, inferring that actual spreading is presently going on. The tensional origin of rift structures is also generally supported by the studies in *oceanic* rift regions, in particular by the development of the hypothesis of sea floor spreading (Vine 1966, and others) and from fault-plane solutions (Sykes 1967, Isacks et al. 1968, and others).

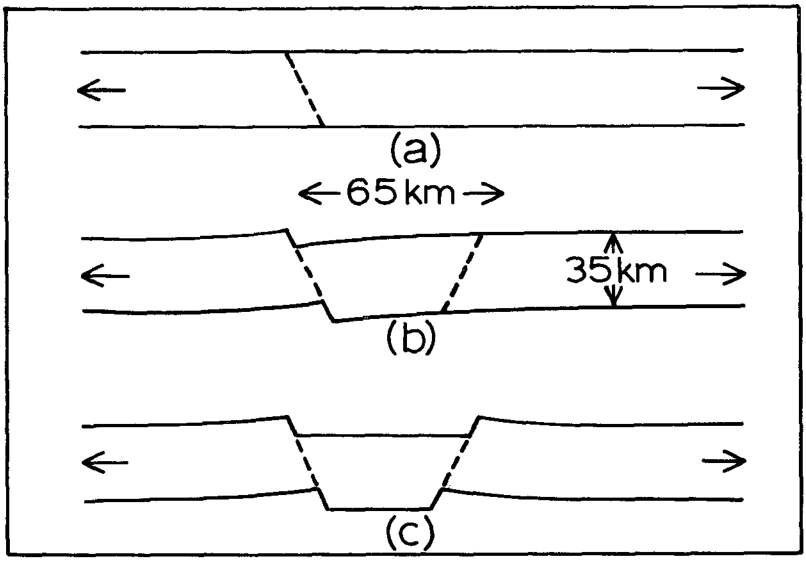


Fig. 10. Development of a graben by subsidence of an elongate 'keystone' between parallel normal faults formed in response to tensional stress which exceeded the crustal breaking strength (*after* Heiskanen & Vening Meinesz, 1958).

The development of a graben by tensile forces or stress release has been theoretically treated by Heiskanen & Vening Meinesz (1958) (Fig. 10). Assuming the occurrence of pure tension and gradationally increasing rock densities with depth, the theoretical dip of the normal faults is found to be 63°, which is close to actual values frequently observed in nature. Graben structures similar to a subsided keystone (Fig. 10c) imply increased depth to Moho and crustal interfaces, and such structures have in fact been inferred from seismic reflection data (Kanasewich et al. 1969). The normal faults which in the uppermost crust have the character of brittle fracture, must deeper down be replaced by zones of plastic deformation. Within areas of active graben formation where abnormal high temperature gradients occur (Lubimova 1969) this would be especially true. In the Rhine Graben, normal faults have been traced by reflection seismology down to a depth of about 7 km (Dohr 1957) but originally proceeded to at least 20 km depth (Illies 1970).

The subsidence of a wedge-shaped crustal block leads to a thickening of the crust under the rift. However, a number of studies seem to indicate the opposite feature, that the rifts represent the superficial counterpart to zones of crustal thinning and deep-seated plastic deformations (Artemjev & Artyushkov 1971, Meissner et al. 1970, Khan & Mansfield 1971); Fig. 11. As in the case of mid-oceanic ridges (Talwani et al. 1965) and in the central graben of the inter-continental Red Sea rift (Girdler 1958, Drake & Girdler 1964), the upper mantle below active continental rifts is characterized by anomalous P wave velocities in the range 7.3–7.7 km/sec (Cook 1969, Mueller et al. 1969, Meissner et al. 1970, Ansorge et al. 1970, Gornostaieve et al. 1970, Griffiths et al. 1971, Lepine et al. 1972). This low velocity, low density mantle material

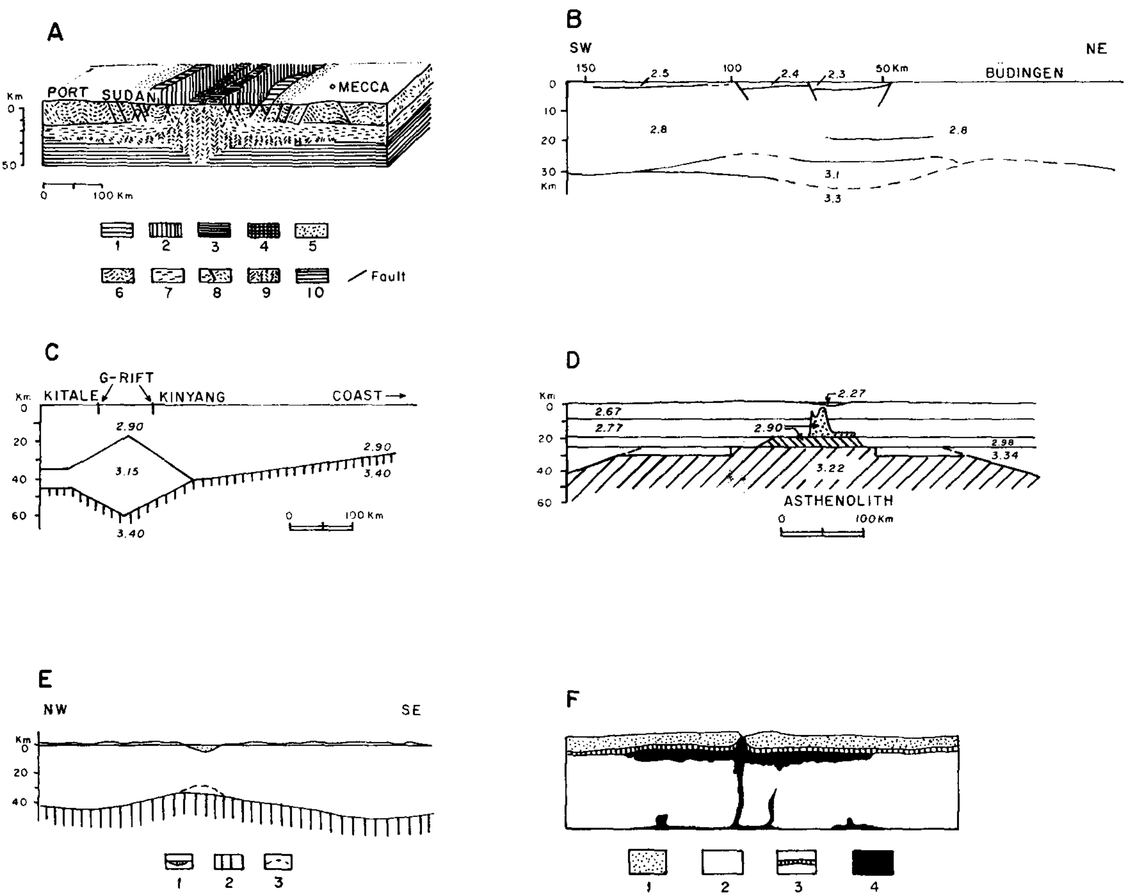


Fig. 11. Graben models: (a) The Red Sea Rift, 1 to 4: depth levels, 5 = coastal plain, 6 = basement, 7 = lower crust, 8 = crustal structure transformed by the rising sub-crustal pillow, 9 = sub-crustal body of mantle-derived material, 10 = upper mantle (Illies 1970), (b) the Rhine Graben showing crustal thinning and sub-crustal pillow-shaped body of density 3.1 based on seismic data (Meissner et al. 1970), (c) the Gregory Rift, Kenya (at 1° N) with a thinned crust and dense sub-crustal pillow (Khan & Mansfield 1971), (d) the Gregory Rift (WSW–ENE profile at about 0.2° S), showing crustal and lithosphere thinning and axial dike (Searle 1970), (e) the Baikal Rift, indicating a marked crustal thinning (Artemjev & Artyushkov 1971), (f) dynamic model run in a centrifuge, 1 = powdered wax, 2 = layering of modelling clay, 3 = painter's putty, 4 = bouncing putty (H. Ramberg 1971).

has been interpreted (e.g. Illies 1970) as a 'cushion' or pillow-shaped intrusion responsible for the upwarping and initial fracturing in rift zones (Fig. 11 a, b & c).

An alternative interpretation (Girdler et al. 1969, Searle 1970, Searle & Gouin 1972) is that the low velocity material is continuous with the asthenosphere and consequently implies not only crustal but also lithosphere thinning (Fig. 11 d). Both alternatives are consistent with the concept of an upward flow of hot, partly melted, mantle-derived material as outlined by, e.g., Wilson (1963, 1965) and H. Ramberg (1971). The injection of the ascending eruptive

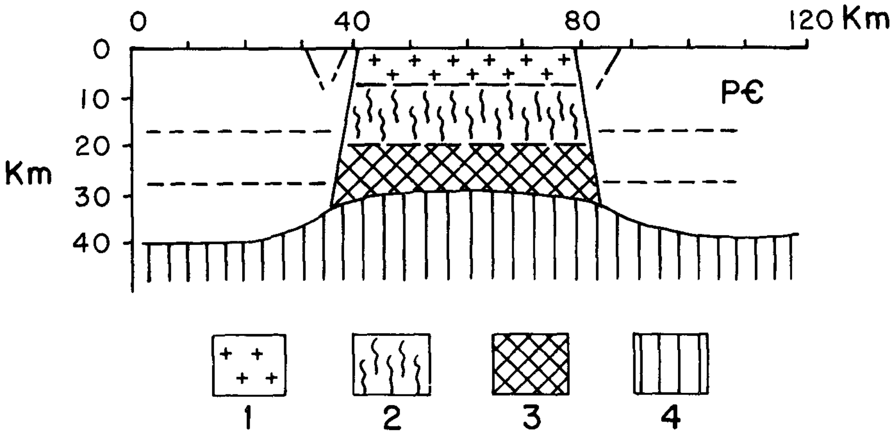


Fig. 12. Sketch of the Oslo Graben in cross-section. 1 = felsic intrusives, 2 = rocks of monzodioritic composition (gneisses and intrusives), 3 = mafic/ultramafic intrusives, 4 = upper mantle.

masses and heat into the lithosphere may lead to the broad epeirogenic arching or doming characteristically associated with continental rifts (King 1967, Holmes 1965, Beloussov 1969, Gass & Gibson 1969, Illies 1970). Artemjev & Artyushov (1971) argued, however, that axial depressions cannot be the result solely of extensive arching, but that the geological and geophysical data best of all comply with a scheme acording to which rifting is primarily connected with a stretching of the earth's crust at strains less than the ultimate strength.

Many grabens typically exhibit local negative gravity anomalies due to their sedimentary fill and sometimes broad negative anomalies due to the postulated low density asthenolith. Recently, sharp axial gravity highs have been reported from rifts in, e.g., Kenya and Ethiopia (Searle 1970, Khan & Mansfield 1971, Searle & Gouin 1972, Darracott et al. 1972). This has been attributed to intrusions of mafic rocks suggesting extreme lithosphere thinning. Similar interpretations have been advanced from the Red Sea and Gulf of Aden on the basis of axial gravity highs, and magnetic and seismic data (Girdler 1964).

Consistent with the recent rift interpretations, preliminary gravity accounts from the Oslo Graben (Ramberg & Smithson 1971, Ramberg 1971, 1972b) have indicated that the rift is associated with crustal thinning, the occurrence of dense intrusive masses at depth, and a very moderate separation. The preliminary analysis has given rise to a tentative model (Fig. 12), a model which will be reworked and refined in the chapters which follow.

4. Rock density measurements

Bouguer and terrain correction depends on the density of surface materials within the range of elevation of the survey, and the interpretation of gravity anomalies relies upon estimates of the subsurface density distribution at least down to the lower levels of the models considered. Density is an important physical property of rocks that is related to chemical composition. Also, the density studies make a necessary first step in mass calculations of rock-types and quantitative petrogenetic considerations. These reasons make it desirable to achieve a thorough knowledge of the various rock densities within the area investigated, even though general information about igneous and metamorphic rock densities are available (e.g. Clark 1966, Wenk & Wenk 1969, Smithson 1971). In the present chapter a short account of the experimental methods used and a brief analysis of the rock densities and their variation within the rock bodies are given.

4.1 EXPERIMENTAL METHOD

Estimates of rock densities can be made either from actual measurements of rock samples or by the profile method of Nettleton (1940). Actual measurements are usually beyond the scope of a regional study, but have been used here chiefly because a large number of rock samples (about 950) were already available from a geochemical sampling project (Raade 1973). The additional ca. 600 samples were collected from outcrops near gravity stations or from local areas subjected to detailed geophysical and geological studies. The outcrops provide fairly fresh rock samples in contrast to most non-glaciated areas where it might be necessary to use drill cores instead of surface samples (Baird et al. 1967). Efforts were made to obtain material as fresh and unaltered as possible; the majority of the samples were collected in road-cuts and quarries.

The densities were determined by the standard method of weighing the sample in air and in water, using a Mettler P7 balance. The precision of the method is about 0.002 g/cm^3. At low pressures, even crystalline rocks have a certain degree of porosity. At somewhat higher pressures, there are indications that the small, open cracks are closed or that the rocks are completely saturated with water (Simmons & Nur 1968). Hence, all the samples were soaked in water for about an hour before the measurements were taken. Even so, the measurements, especially of the often highly miarolitic subvolcanic varieties such as ekerite, nordmarkite, etc., are believed to be slightly low. To test the degree of saturation a few samples were left in water in a vacuum tank for several hours, but the average difference between dry and wet bulk rock density seemed insignificant, being about 0.006 g/cm^3 for the three nordmarkite samples selected for their highly miarolitic appearance (Ramberg 1972a, Table IV).

The variance within outcrop (Table 4) because of slightly varying mineralogical composition is, however, in general much greater than the variance of the standard method of weighing and the change due to incomplete saturation. In addition, only the average properties of the rock masses will normally be sensed by any geophysical technique. It is, therefore, more important to have a large number of density measurements covering the whole area rather than a more limited number of samples with extremely high accuracy.

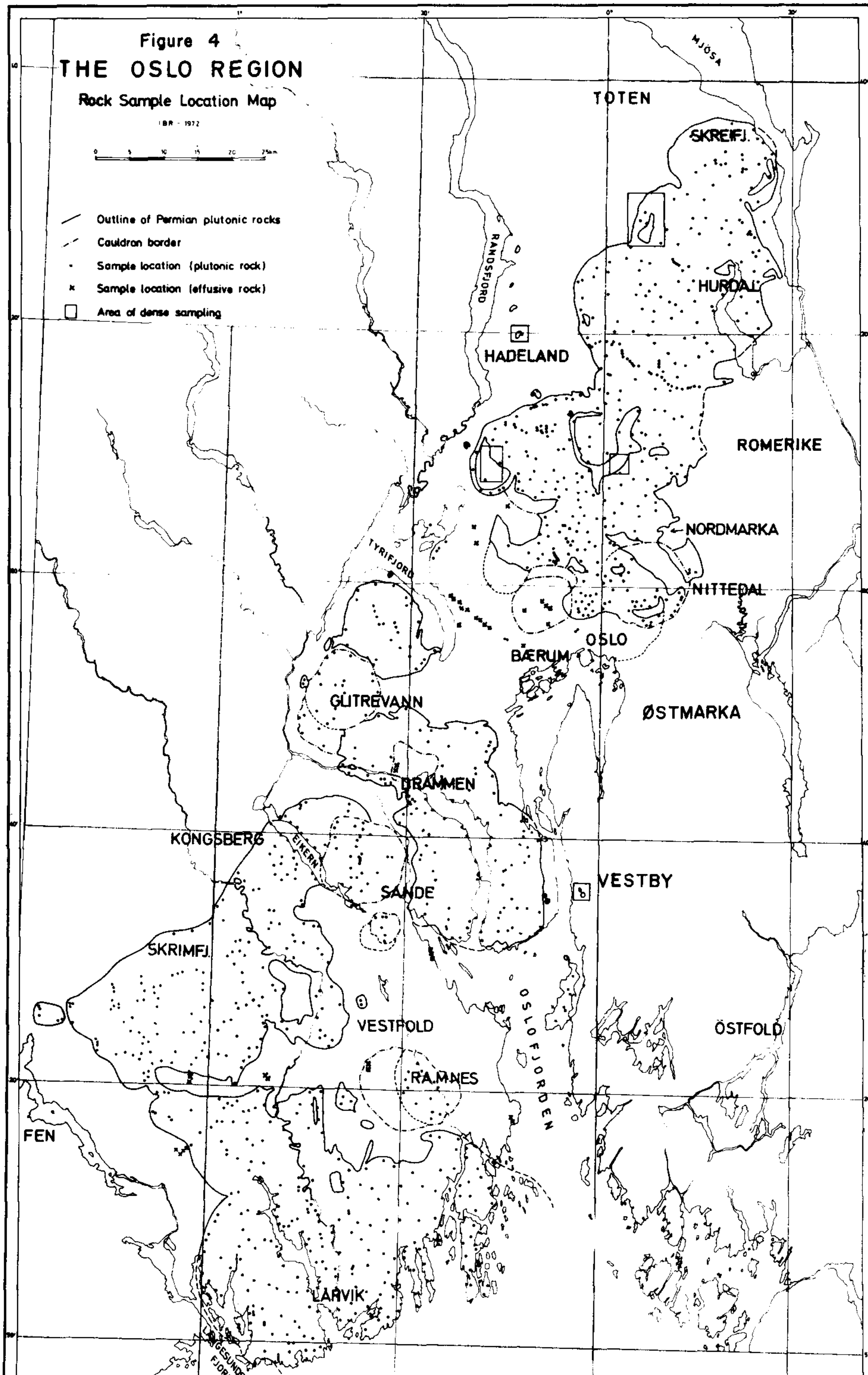


Fig. 13. Rock sample location map.

Table 4: Variance of density within outcrop*

Rock Type	Locality	No. of Samples	Mean Density g/cm^3	St. Dev. $S(g/cm^3)$	Variance S^2
Biotite granite	Hyggen quarry	7	2.596	0.009	0.000081
Syenite	Skrukkelia	6	2.625	0.012	0.000144
Pyroxenite	Vestby quarry	7	3.151	0.023	0.000529

* Arbitrarily defined as an area less than 5 x 5 m.

Table 5: Areal distribution and mean density of main rock-types in the Oslo Region (south of lake Mjøsa)

Rock-type (see also Table 3)	Areal distribution* km^2	%	Mean Density g/cm^3	No. of Samples	Stand. Dev. S	Range
C11 {Quartz porphyry / Felsite porphyry}	31	0.4	2.629	30	0.030	2.570–2.681
C12 Rhomb-porphyry	1153	13.4	2.720	24	0.043	2.610–2.787
C13 Basalt	220	2.6	2.906	41	0.117	2.690–3.130
Σ Volcanic rocks	1404	16.4				
C21 Granite	840	9.8	2.604	205	0.016	2.564–2.660
C22 Ekerite	821	9.6	2.609	123	0.020	2.540–2.662
C23 Nordmarkite, etc.	1425	16.6	2.620	364	0.029	2.531–2.750
C25 Larvikite	1670	19.5	2.711	334	0.038	2.601–2.910
C26 Lardalite	65	0.8	2.709	22	0.035	2.639–2.770
C28 Akerite	52	0.6	2.749	35	0.053	2.649–2.872
C29 Kjelsåsite	201	2.3	2.794	30	0.076	2.636–2.950
C31 'Oslo-essexite'	15	0.2	3.056	29	0.157	2.780–3.320
Σ Plutonic rocks	5089	59.4				
C41 Cambro-Silurian rocks	1545	18.0	2.783	31	0.087	2.648–3.020
C42 Downtonian sandstone	95	1.1	2.674	14	0.029	2.631–2.720
Σ Sedimentary rocks	1640	19.1				
C51 Precambrian rocks (within graben)	440	5.1	2.737	29	0.099	2.600–3.013
Total	8573	100.0		1311		
Weighted mean density of plutonic rocks C21–31:			2.66	1142		
Precambrian rocks outside the graben (various sources, see Table 7)			2.74	315		

* Mainly after Barth (1945, 1954)

Altogether 1576 density measurements have been made of various rock types from within the graben and 315 measurements from outside the graben. The locations and areal coverage of the igneous rocks are shown in Fig. 13. The average size of the samples was about 1.5 kg, and the range about 0.5–4.0 kg. The ca. 950 'geochemical' samples were all identified by means of both hand-specimen and thin-section studies (Raade 1973), while the re-

Table 6: Mean densities of some local or transitional rock-types or complexes

Rock-type	Mean Density g/cm³	No. of. Samples	St. Dev. S	Range
Foyaite (C27)	2.611	11	0.038	2.52–2.64
Pulaskite/Hedrumite (C24)	2.666	33	0.038	2.60–2.73
Vestby Volcanic Plugs (C32)	3.204	135	0.211	2.73–3.64
Viksbergene Volcanic Plug (C33)	2.978	97	0.133	2.69–3.25

Table 7: Densities of Precambrian rocks, southern Norway

Source area	Mean Density	No. of Samples	Source
Romerike-Randsfjorden	2.75 g/cm³	35	Present study
Skien-Gvarv	2.68	29	»
Vicinity of the Vestby volc. plugs	2.71	30	»
Østfold	2.75	61	Ramberg & Smithson 1971
Solør-Kongsvinger	2.73	32	Gvein, pers. comm. 1969
E of Flå granite	2.74	36	Smithson 1963a
W of Flå granite	2.78	56	»
Telemark	2.74	36	Smithson 1963b

maining ca. 600 samples were mostly classified simply by inspection of the hand-specimens.

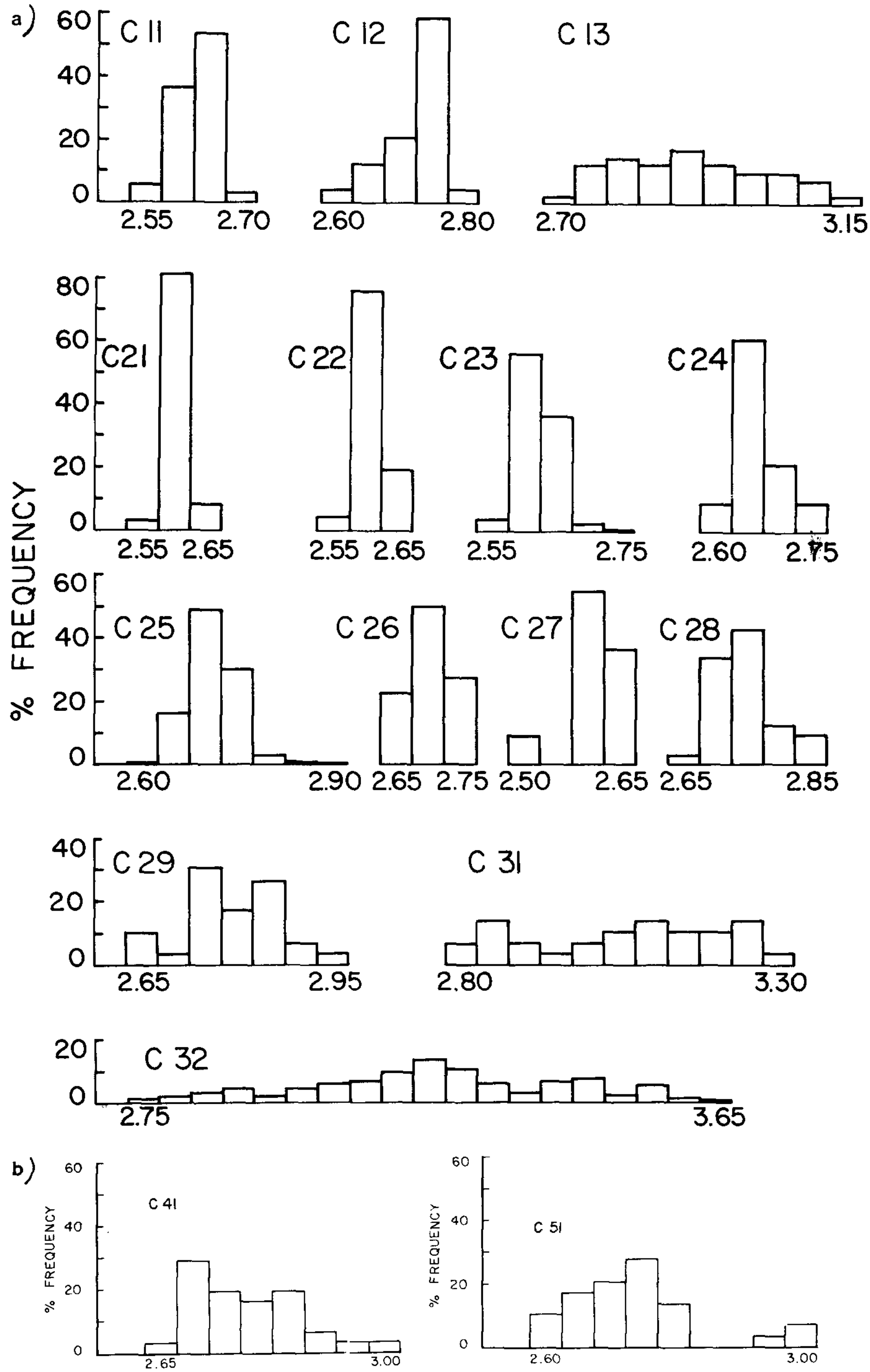
4.2 ROCK DENSITIES

The mean rock densities and the areal distribution of the main rock-types are summarized in Table 5, while the results of some density studies of local or transitional rock-types or complexes are given in Table 6.

The calculated areal distribution in Table 5 considers only the graben area south of Lake Mjøsa, and does not include the occurrences of rhomb-porphyry and sandstone in Brumunddalen (Rosendahl 1929). In the northern part of the area the Randsfjorden-Hunselven fault zone has been assumed to define the graben border toward the west, thereby including a large block of Precambrian and Cambro-Silurian rocks not included in earlier calculations (Barth 1945, 1954).

The 315 Precambrian rock samples from outside the graben (Table 5) are split into their various source areas in Table 7. The mean density (M_{315} = 2.744 g/cm³) is strikingly close to that of the 29 samples of Precambrian rocks from within the graben borders (M_{29} = 2.737 g/cm³). If one adds 54 samples from the Bamble area SW of the Oslo Region (Smithson 1963b), which includes many omphibolites and hyperitic rocks, one arrives at a mean of M_{369} = 2.754 g/cm³. Recent sampling of Bamble rocks, however, suggests that the previous Bamble average (2.81 g/cm³) is too high because of over-representation of mafic rocks.

Woollard (1962, 1969a) proposed that 2.74 g/cm³ would be a better mean density than 2.67 g/cm³ for the upper crust, and Smithson (1971) has suggested



14. Frequency distribution of rock densities from various rock-types (categories), (a) Oslo igneous rocks: C11 = quartz and felsic porphyry, C12 = rhomb-porphyry, C13 = basaltic rocks, C21 = biotite granite, C22 = ekerite, C23 = nordmarkite, etc. C24 = pulaskite-hedrumite, C25 = larvikite, tønsbergite, C26 = lardalite, C27 = foyaite, C28 = akerite, C29 = kjelsåsite, sørkedalite, C31 = 'Oslo-essexites' (various occurrences), C32 = Vestby volcanic necks, (b) Cambro-Silurian (C41) and Precambrian rocks (C51) from within the graben.

that rock densities from most metamorphic terrains (except granulite facies gneisses) fall in the range 2.70–2.79 g/cm^3. The data from Southern Norway support these conclusions, and a mean density of 2.74 g/cm^3 has accordingly been chosen for the Precambrian shield rocks surrounding the graben. This value accords to an overall intermediate composition and agrees with the estimated average chemical composition of shield areas, approximating that of granodiorite, mica gneiss or graywacke (Poldervaart 1955, Shaw et al. 1967).

In contrast to the Precambrian rocks the weighted mean density of the Permian plutonic rocks is about 2.66 g/cm^3, resulting in a negative density contrast of 0.08 g/cm^3. The remaining rocks within the graben borders, volcanics and sedimentary rocks, and Precambrian gneisses, will not affect this contrast very much, and the Oslo Region should as a whole be an area of negative gravity anomalies.

Of the various plutonic rocks the granites and syenites, having the largest negative density contrasts to the Precambrian, will give rise to negative anomalies, while larger masses of kjelsåsite and the basic volcanic necks will similarly be expected to cause positive anomalies. The monzonites, which show slight negative density contrast with respect to the Precambrian, will theoretically be associated with small negative anomalies, but the frequent occurrence of transitional rock-types into the more dense akerites and kjelsåsites (especially north of Oslo) or into the pulaskites and nordmarkites, renders this conclusion of little practical value. The density studies suggest that neither the mass nor the geometry of the monzonitic plutons can possibly be outlined by means of gravity measurements.

Tables 5 and 6 show that most of the igneous rocks form well-defined density populations with relatively small standard deviations. Exceptions are the basic volcanic rocks and the basaltic rocks which exihibit polymodal distributions, while the rest are close to normally distributed (Fig. 14). Also the akerites and kjelsåsites form somewhat ill-defined density populations, a feature which reflects the fact that in some areas (Skrukkelia – Høvernsjøen, Sørkedalen – Heggeli, Sande cauldron) they constitute rock complexes ranging from monzonitic to gabbroic composition.

The granites and nordmarkites exhibit the lowest standard deviations and most extreme peakedness, and their average densities are significantly different from those of the monzonitic rocks. The transitional pulaskite-hedrumite group seems to form a well-defined density population between that of the nordmarkites and the monzonites. However, a t-test reveals that the differences between the pulaskites and the nordmarkites on one side and the monzonites on the other side, are not big enough to contradict the prediction that they are equal at a probability level of 5%. Similarly, the monzonite and the akerite densities are for all practical purposes equal (5% level), while the same conclusion does not apply to the monzonites when compared with kjelsåsites or any other more mafic group.

The trachyte-rhyolite group and the rhomb-porphyries show average densities close to those of the acidic plutonic rocks and monzonites, respectively. The

a)

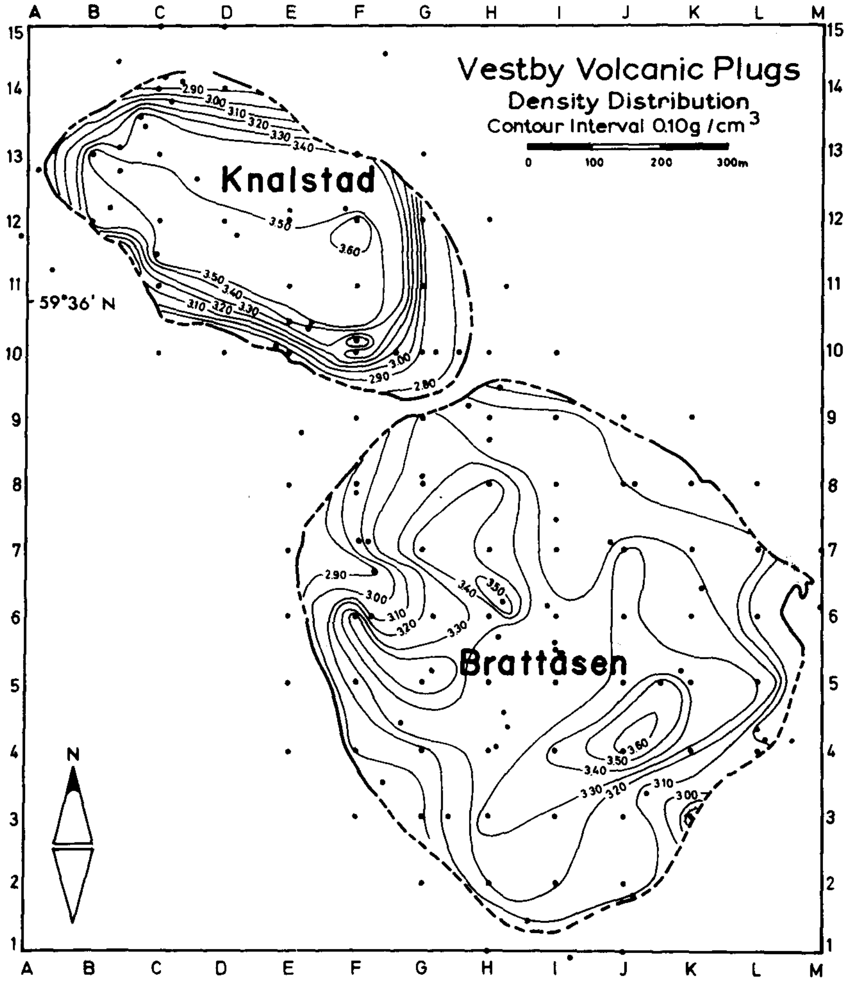


Fig. 15a. Surface density distribution within the Vestby volcanic necks.

basaltic rocks, however, show considerably lower average density than their possible deep-seated equivalents in the volcanic pipes.

4.3 LATERAL DENSITY VARIATIONS WITHIN THE ROCK COMPLEXES

Subhorizontal (= present surface) density variations are exemplified in the most extreme way by some of the volcanic necks. In the Vestby volcanic necks (Fig. 15a) a concentric pattern with a large central plateau is apparent both in the Knalstad neck and the more circular Brattåsen neck. Apart from relations close to the border, sharp contacts between large masses of contrasting rock-types are rarely observed. The density pattern therefore reflects radial gradational transitions in the chemical and mineralogical composition of the intrusive complexes. The extremely high density values found at Vestby are due to the high content of magnetite in some of the pyroxenitic and gabbroic varieties.

b)

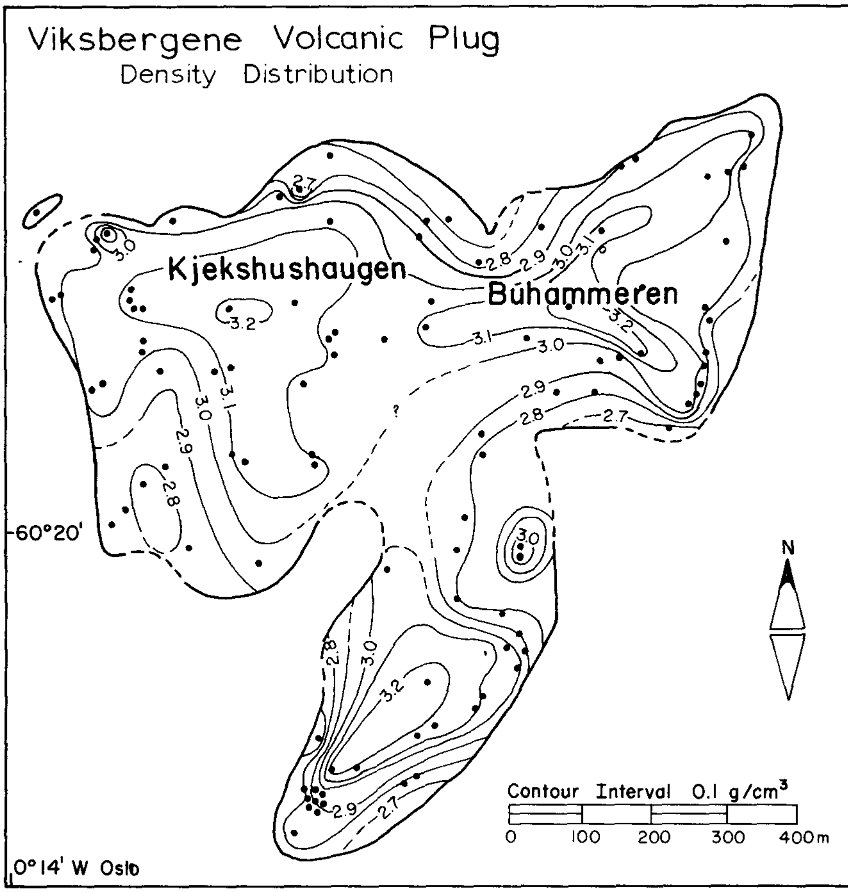


Fig. 15b. Surface density distribution within the Viksbergene volcanic neck (outline mainly *after* V. Singh, pers. com. 1972).

This phenomenon is also reflected in the sharp vertical ground magnetic anomalies over the necks, reaching peak values of about 12,000 γ.

The concentric pattern in Vestby seems to be typical for many of the basic to ultrabasic pipes in the Oslo Region, for instance Viksbergene in Hadeland (Fig. 15b). In general, the same concentric pattern is recognized, but clearly it is more complex as revealed by the existence of at least three density (intrusion?) centers. These coincide with ground magnetic anomalies of maximum 7000 γ, indicating that the increased density may correlate with concentrations of magnetic minerals.

The tendency for decreased density towards the margins also occurs in some of the larger plutonic bodies, but much less pronounced than in the basic pipes. In the sub-circular monzonite pluton around Larvik (Fig. 16) this phenomenon is recognized in the western and northern part of the pluton, although heavily distorted by later syenitic intrusives in the north-western part. The shape of the pluton and the density-distribution pattern indicate that another half of the pluton is to be found below the sea towards the south-east as previously inferred on the basis of aeromagnetic data (Sellevoll & Aalstad 1971).

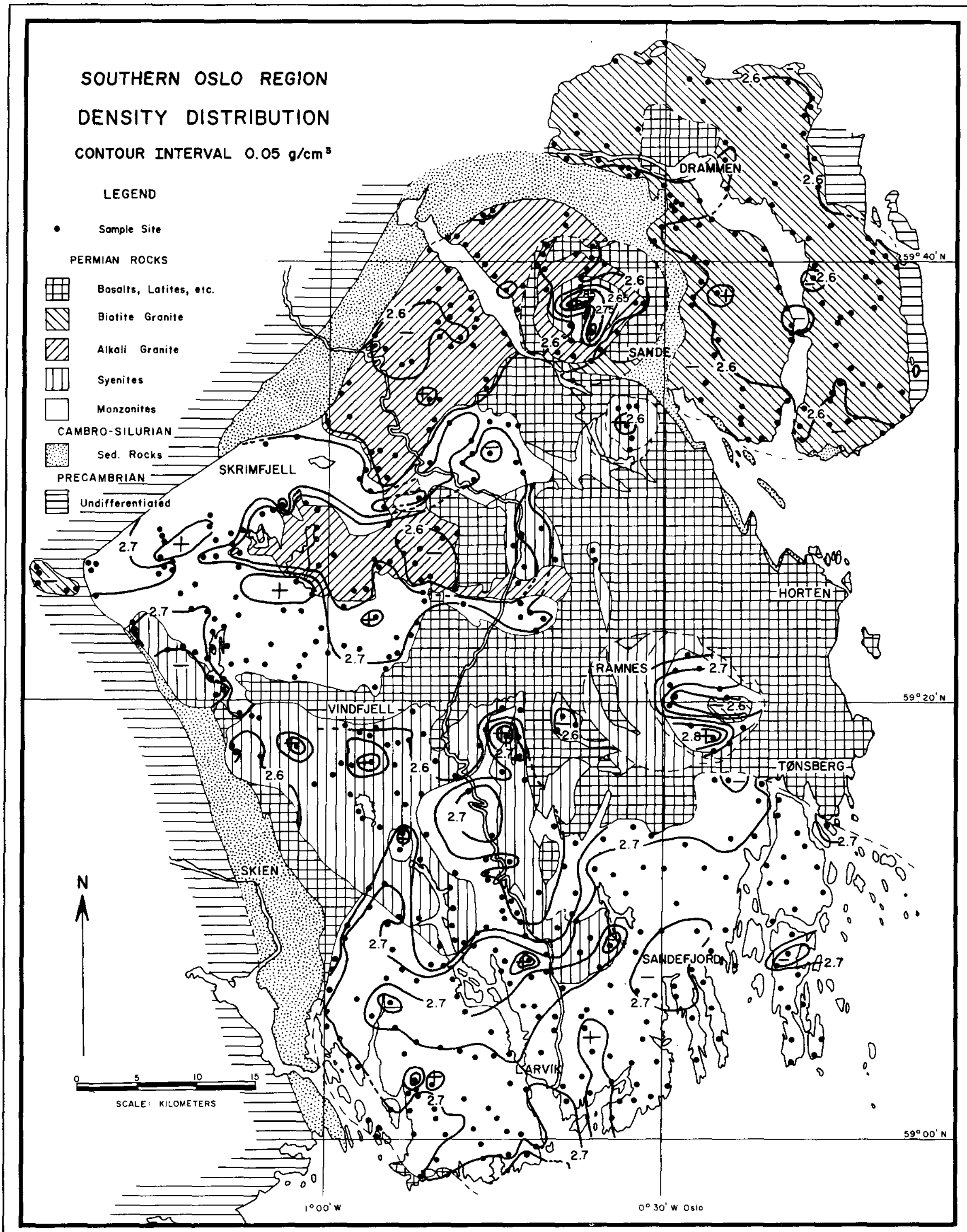


Fig. 16. Surface density distribution within the plutons of the southern Oslo Region.

As can be seen from Fig. 16, density lows occur north and south of Larvik and in the Sandefjord area, indicating that large-scale compositional variations do occur within the central parts of the pluton. The density variations are not confined to the surface only, but persist to some depth since the 'anomalies' are closely associated with variations in the gravtiy field (Chapter 5).

A different pattern is revealed in the Skrimfjell monzonite pluton (Fig. 16) which is intruded by a very irregularly shaped body of soda granite (ekerite), outlined by smooth density curves. The density low in the monzonite east of the river Lågen is caused by rocks transitional into syenite. The opposite conclusion applies to the nordmarkite east of Skien. Here, transitional types to larvikite are frequent, as inferred by the density distributions.

In the Drammen granite, a low-density, marginal belt could be associated with a porphyritic facies of the granite, but detailed mapping of the various sub-types within the granite has presently only been carried out in smaller areas (A. Gaut pers. comm. 1973). A similar, porphyritic facies in the southern, marginal part of the Finnemarka granite is not associated with any notable change in density. The density pattern in the Drammen granite may therefore suggest reversed zoning.

For the majority of the remaining plutons the density values, in general, seem to be randomly distributed, and the most significant density changes occur at borders between contrasting rock-types. The geological map (Fig. 2) combined with the average density values (Tables 5 and 6) can therefore be applied to obtain the horizontal density distribution for the larger part of the igneous province.

4.4 DENSITY VARIATIONS WITH DEPTH

Experience has shown (e.g. Birch 1960, 1961, Christensen 1965) that seismic velocities usually increase with increasing rock density or mean atomic weight. The empirical relation is roughly known, and the seismic velocities can be used with a certain degree of confidence to deduce probable density values (Woollard 1959, Talwani et al. 1959a).

Seismic studies in Southern Norway (Sellevoll & Warrick 1971, Kanestrøm & Haugland 1971, Vogel & Lund 1971) show that P velocities in the range 6.0–6–3 km/sec occur down to depths of ca. 15–18 km and that P_b velocities of ca. 6.5 km/ sec are characteristic of an intermediate crustal layer. Any clear intermediate or Conrad discontinuity is not reported, and a gradational transition between the two layers is possible. This means that the observed mean density at the surface (~2.74 g/cm^3) is probably representative for approximately the uppermost ten kilometers, since the elimination of an initial porosity (usually <3 % in crystalline rocks) is almost exactly offset by thermal expansion at depth (Woollard 1959), and that a slight increase in density because of change in crustal composition is to be expected in the lower part of the upper crust. In the intermediate layer at depths of about 17 to 27 km the average density has increased to about 2.90 g/cm^3. This represents an increase in density from

Table 8: Velocity and density distribution model for southern Norway

Depth (Km) Vp (Km/s)	Average rock density, (g/cm³)					
Kaneström & Haugland (1971)	1	2	3	4	5	6 Chosen Values («Standard Section»)
6.2	2.81	2.91	2.84	2.90	2.73	2.74 / 2.88 } 2.80
17 — C						
Crust 6.55	2.91	3.00	2.94	2.98	2.80	2.93 } 2.98
27 — CM						
7.1–7.3	3.05–3.12	3.14–3.19	3.09–3.15	3.10–3.15	3.00–3.08	3.05 } 2.98
34 — M						
Mantle 8.20	3.41	3.42	3.37	3.35	3.38	3.41
Mean density of crust,	2.90	2.98	2.93	2.97	2.81	2.90
Density contrast crust/mantle,	0.51	0.44	0.44	0.38	0.57	0.51
Mass per unit area at 40 km depth (kg/cm²)	11 890	12 210	11 980	12 105	11 600	11 865

1. Nafe – Drake curve, Talwani et al., (1959)
2. Nafe – Drake curve, McCamy et al. (1962)
3. Woollard (1959)
4. Woollard's curve reproduced in Mc Camy et al. (1962)
5. Volarovich et al. (1967)

surface to the upper part of the lower crust of about 5 % which is grossly similar to the density increase expected in granitic rocks recrystallizing into granulite facies mineralogy (Green & Lambert 1965). A deep crustal layer with P wave velocities in the range 7.0–7.4 km/sec has been reported from Scandinavia (Sellevoll 1966, Kanestrøm & Haugland 1971, Vogel & Lund 1971, Gregersen 1971). This would equal densities in the range 3.0–3.15g/cm³.

A tentative crustal model for Southern Norway, based on the seismic refraction profile Trøndelag-Oslofjorden (Kanestrøm & Haugland 1971), is presented in Table 8. Mean densities and density contrasts have for comparison been calculated on the basis of various density-velocity curves. The scatter in the calculated densities reflects the uncertainty involved, and the chosen values (Table 8, col. 6) are based on the original Nafe-Drake relationship (Talwani et al. 1959a, M. Talwani pers. comm. 1970) and on isostatic considerations.

Assuming that the earth's crust and upper mantle are regionally in a state of isostatic equilibrium, it follows that regionally the crust and upper mantle have the same mass per unit area above a certain depth of compensation. If one rather arbitrarily assumes that isostatic equilibrium is achieved within the upper 40 km, 11,840 kg/cm² has been suggested as a reasonable value for normal crustal mass per unit area for such a column (Hess 1955, Woodside & Bowin 1970). If the chosen section of Southern Norway (Table 8, col. 6) is taken as a standard *sea-level* section, the mass per unit area is 11 865 kg/cm^2.

Since regional variations in the depth to the various crustal interfaces have not been reported from Southern Norway, one may assume, as commonly employed (e.g. Woodside & Bowin 1970, Artemjev & Artyushkov 1971), that the obtained regional gravity anomalies are connected mainly with variations in the depth to the Moho discontinuity. The average density contrast between crust and upper mantle is, therefore, of prime importance.

The average densities and density contrast between crust and upper mantle have been discussed in a number of papers and along various lines of evidence (Woollard 1962, 1969a, Drake et al. 1959, Talwani et al. 1959a, Dorman & Ewing 1962, Ringwood & Green 1966, etc.). Either approach infers that the crust is significantly denser than 2.67 g/cm^3, and the accepted value seems to be found in the interval 2.88–2.93 g/cm^3. Similarly, most estimates of upper mantle density fall within the range 3.35–3.42 g/cm^3, while the density contrast crust to upper mantle is most likely in the range 0.38–0.52 g/cm^3. In Southern Norway, the crust/upper mantle density contrast (based on crustal thickness and surface elevation) was tentatively set to 0.50 g/cm^3, while a contrast of 0.45 g/cm^3 gave the best fit between calculated and observed regional Bouguer anomaly along the seismic profile Trøndelag–Oslofjorden (Kanestrøm & Haugland 1971).

From Table 8 (col. 6) it is seen that the *average* density contrast is as high as $\Delta\rho = 0.51$ g/cm^3, while the density contrast at the crust/mantle interface (Moho) is 0.36 g/cm^3. The contrast between the lower crust as a whole (2.98 g/cm^3) and the upper mantle is 0.43 g/cm^3. For the calculations, therefore, a contrast of 0.45 g/cm^3 has been generally applied and other values have been used for comparison.

5. Gravity measurements and anomaly maps

5.1. GRAVITY STATION COVERAGE

About 5 300 gravity stations form the basis of the present analysis. During the field seasons 1966–1971 about 2 650 new gravity stations were established in the Oslo Region and surroundings. These stations, together with gravity stations from the Norwegian Geographical Survey (NGO) and other sources (Smithson 1961, 1963a and b, Ramberg & Smithson 1971), have been plotted on a bedrock map of the Oslo Region (Ramberg 1972a, fig. 1). The overall coverage for the area surveyed is about 1 station per 5 km^2. In central parts of the area, stations are more closely spaced leaving the surrounding Precambrian area correspondingly less well-covered.

Detailed surveys have been accomplished in particular areas (see Ramberg 1972a, fig. 1), some of which are dealt with in the present paper. Others have been treated separately: the Fen area (Ramberg 1964, 1973), the Gjerdingen area (Grønlie 1971), the Gran area (Lehne 1972), and the Øyangen area (Ramberg & Larsen, in prep.). In order to further improve the coverage, some 490 new gravity sites were occupied in 1972 especially in areas of cauldron subsidences and other local features in the southern half of the region. The locations of these gravity stations are shown in Figs. 33 and 34.

5.2 FIELD METHODS AND INSTRUMENTS

A detailed description of field methods has been given in an earlier report (Ramberg 1972a), and therefore only a brief summary will be presented here.

A Worden Master gravity meter (WM 653) was used throughout the survey. After 1970 a LaCoste Romberg gravity meter (LCR 214) was also used both for base station readings and general field operations. The overall reproducibility of the instrumental readings is determined to be about ± 0.15 mgal (WM 653) and about ± 0.08 (LCR 214). A number of base stations were established in the area surveyed, all of which were tied to the Oslo Fundamental Gravity Station at the Geological Museum at Tøyen (g = 981 928.15 mgal, Sømod 1957; see discussion in Ramberg 1972a, p. 5).

Elevations north of Oslo were generally established by means of 3 seven-inch precision barometers (Paulin), while in the southern part of the region they were mostly based on spot elevations from detailed topographic maps (scale 1:5,000). Standard deviations (S_E) are estimated to be about ± 4.0 m and ± 2.0 m for the two methods, respectively. In local surveys, and whenever available, bench marks and levelled elevations were applied ($S_E = \pm 0.01$ m). The horizontal control as determined by the scale and quality of the maps available (limiting factor: maps series M711, new edition, scale 1:50,000) is probably better than ± 50 m (or about ± 0.03′).

5.3 GRAVITY REDUCTIONS AND UNCERTAINTIES

Bouguer anomalies, Δg_b, were calculated using the formula:

$$\Delta g_b = (g + g_f - g_b + TC) - g_o, \text{ or}$$
$$\Delta g_b = (g + 0.3086\,h - 0.04185 \cdot h\rho + TC) - g_o$$

where g is the observed gravity, g_f is the free air correction, g_b is the Bouguer correction, TC is the terrain correction, g_o is the theoretical gravity given by the 1931 International Gravity Formula (Heiskanen & Vening Meinesz 1958), h is the elevation in meters, and ρ is the rock density.

Full terrain corrections, TC, were computed by the zone chart method of Hammer (1939) to a distance of about 22.0 km for all stations north of Oslo. The corrections are on the average less than 1 mgal (range 0.05 to 8.5 mgal) with an estimated accuracy of about 20 percent which equals an average uncertainty of about ± 0.2 mgal. An automatic terrain correction method based on digitized topographic maps (Grønlie & Ramberg 1973) was applied to a detailed study (the Ullernåsen volcanic neck). The observed rock densities (Table 5) have been used throughout to obtain the terrain corrections. Terrain corrections have not been computed from the southern half of the region where the relief is generally quite moderate (most sample TC < 0.6–0.7 mgal).

A density value of 2.67 g/cm^3 was chosen for computing the Bouguer anomalies. This is consistent with common procedure and enables a direct comparison with Bouguer anomaly maps from neighboring areas. Alternatively, the *observed* density values could have been chosen resulting in a more effective stripping of the gravity effects of the rocks above the plane of reference (sea level) while in the former case the gravity effect of such lateral density variations will be included in the computed Bouguer anomalies. Within the range of elevation (0–700 m) and with the average height above sea-level less than 300 m, as in the present case, the difference is not significantly large. Both alternatives have been calculated and the results from the region north of Oslo are presented in Ramberg (1972a, Table IX). The two methods rarely differ by more than 1 mgal. Hence, the use of 2.67 g/cm^3 as Bouguer density is satisfactory except in strictly local studies involving small mass differences.

The various factors involved in the computation of the Bouguer anomalies have been considered and the standard deviation (S_b) has been calculated for various combinations of instruments and methods (Ramberg 1972a). For the northern half of the region, we have for the most unfavourable case (WM gravity meter, barometric altimetry), that $S_b = \pm 0.83$ mgal for the new gravity sites, which is similar to the standard deviation determined for the NGO stations from the same area. From the southern part of the region (south of Oslo), $S_b = \pm 0.76$ mgal, including the increased uncertainty in not having added terrain corrections. For local studies and areas with small terrain corrections, the standard deviations are considerably lower. Consequently the Bouguer anomaly map of the Oslo Region (Plate 2) has been drawn with a contour interval of 5 mgal, while closer contour intervals are justified for various local maps.

5.4 BOUGUER ANOMALY MAPS

The correct isolation of the Oslo Region gravity high is critically dependent upon a more regional view. Hence a Bouguer anomaly map from Southern Norway has been prepared and presented as Plate 1. The map includes all the available regional gravity survey data (especially from NGO) and some of the more local survey data (various sources) up to 1971. Marine gravity data (Bedsted Andersen 1966) have been included for the Skagerrak area, while Wideland's (1946) gravity map of Southern Sweden has been consulted in the border area to the east. Terrain corrections have not been entered.

Apart from numerous local anomalies, the following important large-scale features (Plate 1) pertain to the present study:

1) Outside the Oslo Region the Bouguer gravity values correlate inversely

with topography; the isogal lines run closely parallel to the smoothed coastline except in the graben area. This relationship, together with the observed relation between Bouguer anomalies and depth to Moho along a N–S profile across Southern Norway (Kanestrøm & Haugland 1971) suggests that the crust in Southern Norway might be generally supported hydrostatically by the Airy hypothesis of isostacy. The apparent anomalous crustal conditions met with in the Oslo Region have been inferred earlier on the basis of various seismic and gravimetric evidence (Schwinner 1928, Crampin 1964, Husebye et al. 1971, Ramberg 1972a).

2) The Oslo Region is marked by a very distinct gravity *high,* trending continuously towards the NNE from the Skagerrak (at about 9° 30′ E and 58° N) to perhaps as far north as Trysil (at about 11° 30′ E and 61° 20′ N). The gravity high is not causally connected with surface geology since it is significantly wider and continues far outside the graben region in the axial direction towards the NNE. Also, simple graben subsidences will normally lead to negative gravity anomalies (Girdler 1964) and the overall negative density contrast of the graben rocks with respect to the surrounding gneisses will similarly lead to negative anomalies which, however, are seen as purely local features superimposed upon the regional high.

3) A broad gravity low follows the main NE to NNE trend of the Caledonian mountain belt. However, the lowest values of the trough occur over the foreland to the east of the central trough (as defined by the Jotun Nappe and the Trondheim Region). The axis closely coincides with the miogeanticlinal ridge as indicated by H. Ramberg (1966, fig. 1); the most negative anomalies being connected with the many basement domes or windows, a feature that can be traced all the way northward along the east front of the fold belt. A similar low is situated to the NW of the Jotun complex above the Møre basal gneiss culmination or eugeanticlinal ridge.

4) Dividing the wide 'Caledonide' low in two, is a marked gravity high associated with the major nappes situated in the 'Faltungsgraben' (Goldschmidt 1916) from about 7° E/60° 40′ N and north-eastwards. The anomaly clearly marks the dense rocks of the Jotun nappe (Smithson et al. 1974) as well as the Trondheim Region in a depressional zone between the geanticlinal axes. Three seismic refraction profiles that cross the structural depression (Sellevoll & Warrick 1971, Kanestrøm & Haugland 1971, Vogel & Lund 1971) indicate no excessive or reduced depth to Moho below the Jotun nappe and the Trondheim Region. The maximum depth to Moho is located close to the axis of maximum negative Bouguer anomalies and to the miogeanticlinal ridge. The excessive masses in especially the Jotun complex, if in isostatic equilibrium, must be regionally compensated.

The above-mentioned relations imply that isostatic balance is approximately maintained over Southern Norway, except for the Oslo Region and possibly also the rather narrow Jotun and offshore Møre highs (see Plate 1). In accordance with the Airy concept of isostacy, a crust of constant density, ϱ_c, but with variable thickness, is hydrostatically balanced by a denser substratum

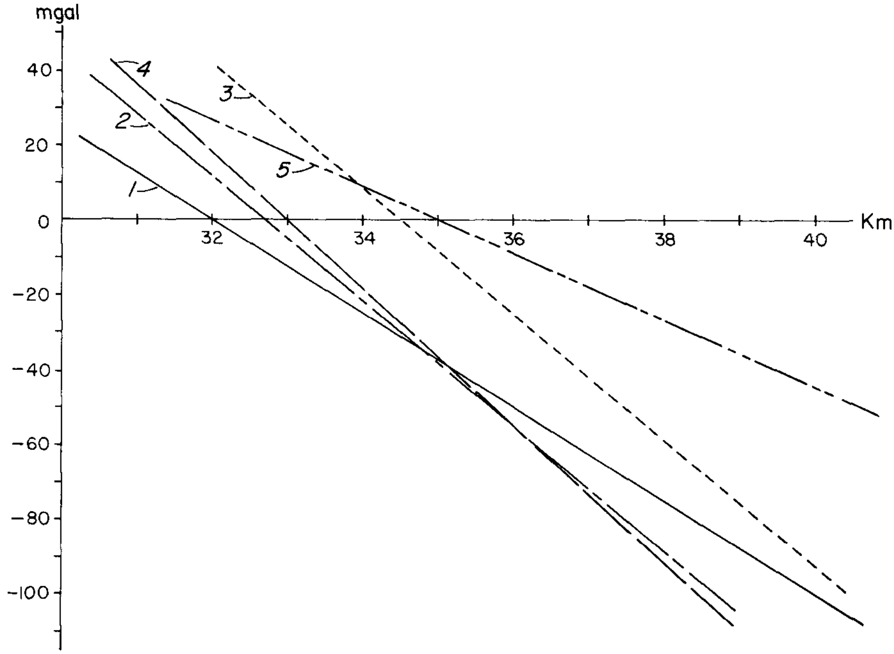


Fig. 17. Bouguer anomalies versus crustal depth based on various empirical relationships as given by (1) Woollard (1959) for the whole earth, (2) R. Kanestrøm (pers. comm. 1972) for the Fedje-Grimstad gravity and seismic profile, (3) Kanestrøm & Haugland (1971) for the Trøndelag-Oslofjord profile, (4) Worzel & Shurbet (1955) for plains, (5) Demenitskaya & Belyaevsky (1969) for the whole earth.

of density ϱ_m (Woollard 1969b), and empirical formulae for the relation between crustal thickness (H), topographic elevation (h), and Bouguer anomalies (g_b) have been given by a number of investigators. Fig. 17 shows various relations for Bouguer anomalies versus crustal thickness; Woollard's (1) and Demenitskaya & Belyaevsky's curves (5) are for the whole earth; Worzel & Shurbet's curve (4) is for plains. Kanestrøm's relations (2 & 3) from Southern Norway are based on more restricted data along two seismic profiles; even so, they are very similar to the other relations except no. 5. From Kanestrøm's original plots, it is evident that without the gravity values from the assumed anomalous Oslo Region, the slope values and the intercepts would come closer to those in Woollard's curve.

Fig. 18 gives the thickness of the normal crust in Southern Norway as derived from Woollard's (1959) empirical relationship $H = 32.0 + 0.08\Delta g_b$. The relative features of the main trends of the map would not have been greatly changed by applying some of the other relationships given, but the absolute depths would vary within the range of a few kilometers, as apparent from Fig. 17. The Bouguer anomalies are averaged values of 1^0 by $30'$ quadrangles (equivalent to about 55 km by 55 km in central parts of Southern Norway) with a 50 % overlap. Each dot on the map (Fig. 18) represents the center of a quadrangle. Though not significantly different, a choice of somewhat larger quadrangles would in effect smooth out the relief, while on the other hand

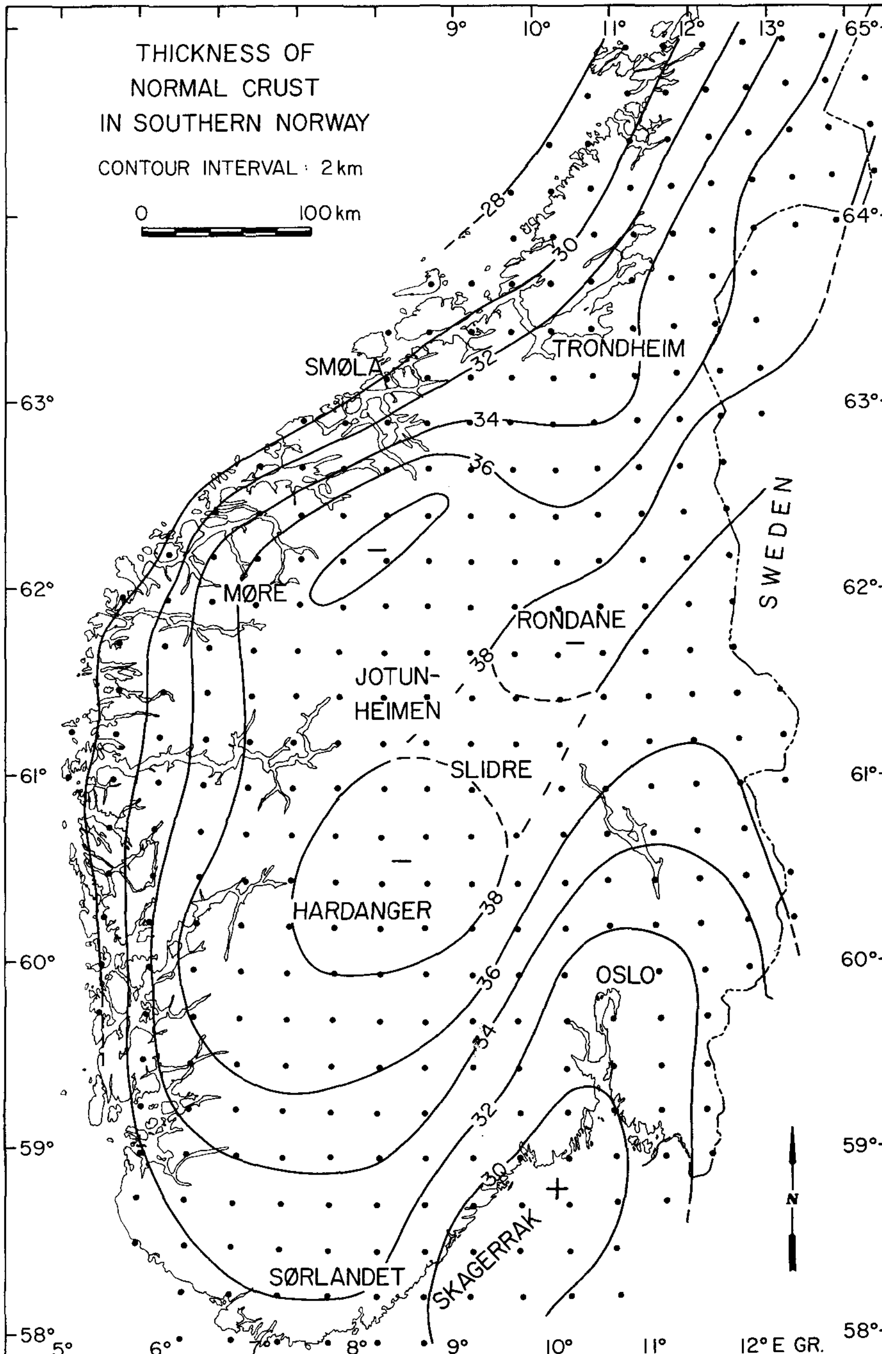


Fig. 18. Gravity normal crust of southern Norway. Based on Woollard's (1959) relationship (see text).

smaller quadrangles would tend to be affected by small-scale, local gravity features.

From Fig. 18 it is apparent that the normal crust is thickest parallel to the main Caledonian axis and the shallows towards the west coast and the Oslo Region. These trends are in general agreement with preliminary Moho contouring based on seismic refraction data (Sellevoll & Warrick 1971, Kanestrøm & Haugland 1971). The anomaly seen in the Slidre area is most likely caused by a local concentration of dense subsurface rocks (Smithson 1964); the axis of crustal depression is most probably continuous and parallel to the geanticlinal ridge throughout the area. Further to the east, into Sweden, similar calculations indicate a relative deepening again of the normal crust, but not to the same extent as below the mountainous central part of Southern Norway. Differences in seismic and normal crustal thicknesses may be due to a number of factors (Demenitskaya & Belyaevsky 1969) but will, in general, indicate unusual crustal and upper mantle composition and/or structure.

In conclusion, the relative thinning of the crust below the Oslo Region is indicated by both seismic and gravity data. This infers that the Oslo gravity high is partly, or completely, due to mantle upwarping below the graben axis; this question will be elaborated upon in Chapter 7.

The Bouguer anomaly map of the Oslo Region (Plate 2) reveals the regional high and many local anomalies in greater detail. The gravity gradient on the *west* side of the Oslo Region reaches its maximum value (up to 4 mgal/km) close to the contact between the Telemark and the Kongsberg/Bamble regions (see also Fig. 2). This signifies that the gradient is affected by the density contrast between the outcropping Telemark and Kongsberg/Bamble rocks, and is caused not only be deeper sources. Stripped for local effects, the 'regional' gradient is therefore much less than 4 mgal/km. In areas where the contact between the Oslo Region and the Telemark rocks is not screened by the Kongsberg/Bamble rocks (such as in the Notodden area and also further north in the Randsfjorden area,) the gradient is more moderate (1.5–2.0 mgal/km) and constant over a wide area.

To the *northeast* we find the largest gradient associated with Magnusson's (1937) 'mylonite zone'. Here, however, rock denser than the average occur on both sides of the mylonite zone. South of the zone, monzodioritic to dioritic gneisses commonly occur, while to the north, in the Solør district, we frequently meet with 'hyperites' or amphibolites similar to those in the Kongsberg and Bamble regions. The zone of 'hyperite' intrusives in the Tangen district is hardly reflected at all. Hence, the gravity gradient in this area is most likely due to subsurface density contrasts.

From the Kongsvinger area and southwards, the enveloping contours of the regional high are deflected from their main trend in two places (Plate 2), that is, at the intersections between the approximately NNE trending regional high and the more NNW–SSE trending gravity troughs and ridges associated with large-scale density variations in the Precambrian gneisses. These variations occur in belts parallel to the main NNW structural grain of the shield. They comprise two relatively low-density belts in the area of Iddefjord–Østfold–Vestby and the area Setskog (59° 45′ N/1° E of Oslo) – Southern Romerike, alternating with relatively high-density belts. Granites occur more frequently in the low-density belts and gabbroic rocks in the high-density belts, but the overall density differences between the belts are small. The gravity pattern indicates that this slight petrographic contrast between the alternating belts persists to some depth. These feaures are still better seen on the residual map (Plate 5); here also a similar trend of alternating gravity highs and lows is recognized on the NW side of the Oslo Region. The many *local* anomalies superimposed on the regional gravity high are also more clearly depicted in the residual map and will be dealt with in detail in later sections.

5.5 SEPARATION OF REGIONAL AND RESIDUAL EFFECTS

"Often, one of the most troublesome parts of the interpretation process is finding a way to isolate anomaly patterns of the kind one is looking for from the remainder of the field. This is referred to as regional-residual analysis," (Grant 1972, p. 651). In certain cases, like that of the present study, this problem might be overcome simply by interpreting the complete observed gravity field directly. However, despite obvious uncertainties involved in the process of separation, it is considered a major advantage to focus upon one feature at a time and to interpret the relatively shallow and possible deep-seated structures independently. The regional gravity anomaly is by definition always represented by a smooth gravity field without minor irregularities. The regional variations may be caused by deep-seated geological structures, as commonly assumed, or by more shallow features which are relatively thin and broad. Usually, there is no exact way of calculating the regional anomaly which may be different depending on the method applied. Depending on what kind of structure is to be studied, the regional may be *chosen* to include gravity effects of all geological features below a certain level which may be everything from a few kilometers below sea-level to the crust-mantle boundary or below. In general, anomalies that are filtered out from the regional field will then appear in the residual, and vice versa, and no gravity information is lost.

A number of methods have been used to separate regional and residual effects; some are graphical, others are analytical (Dobrin 1960, Grant & West 1965). The graphical methods are subjective, empirical, but highly flexible in that they may take into account available geological information. The analytical methods involve routine numerical operations excluding personal judgement and possible prejudices. The numerical smoothing procedures are somewhat arbitrary and generally require regularly gridded gravity readings. Vajk (1951) argued that no unique determination of the regional effect is possible and no mechanical method for its removal can be found. The determination of the regional effect is in itself an interpretational problem unavoidably involving geological preconceptions and judgement of the interpreter. However, accessibility to large computers has contributed toward automatic processing and semiautomatic methods of interpretations in the last decade. This development has to some extent also included the field of regional-residual analyses, including: 1) trend surface analysis, 2) computer modelling, and 3) digital filtering (Grant 1972).

Trend surface analysis (Merriam & Cocke 1967), like the older numerical methods, identifies the 'regional' with some kind of arbitrary local average. Which polynomial order to use is also a subjective decision; and since no geological information goes into these processes, the trend-fitting approaches are not so much used in regional-residual analysis. Another method is computer modelling (Grant 1972), applying interactive programming and the aid of visual graphics. Modelling geology for potential field analysis has thus become more rapid but requires a well-mapped surface geology and a maximum of subsurface control.

Recently, digital filtering has been attempted to modify or to factor the anomaly spectrum (Strahkov 1964, Naidu 1966, 1967, Spector & Grant 1970). Although this may well be the method of the future, the approach still seems to be in an early stage of development. The main difficulty is designing an effective filter from the structure of the anomaly spectrum or 'noise' field.

For the Oslo Region study, two different methods of regional-residual separation have been chosen:

I) The regional was *defined* as the gravity effect caused by variations in crustal thickness on the *assumption* that no lateral density variation occur and that the depth to the Moho is known.

II) The regional was obtained by means of simplified computer modelling. Critical analysis of gravity data as related to known geological structures as well as other geophysical data was entered in an interactive manner, gradually modelling the large-scale geological structures consistent with a smoothed regional anomaly.

Method I: This approach will theoretically provide a regional without uncertainties if the assumptions are fulfilled, and the gravity effects of all crustal inhomogeneities will ideally be picked up by the resulting residual field. The requirements are not satisfied but, to a *first* approximation, the average crustal density may be considered a constant within the sections in question. The major uncertainty is considered to be the depth to Moho.

The configuration of the Moho discontinuity in Southern Norway is grossly known and may be inferred from the following sources:

a) Preliminary Moho contouring based on two seismic refraction profiles (Sellevoll 1967, Sellevoll & Warrick 1971). One of these profiles (Flora-Åsnes) was not reversed, and the contouring is indicated only for the area west of the Oslo Region.

b) Detailed Moho contouring below the Norwegian Seismic Array (NORSAR) and a seismic refraction profile Trøndelag-Oslofjorden (Kanestrøm 1969, 1973, Kanestrøm & Haugland 1971).

c) Isostatic considerations, applying empirical relations between mean topographical height and crustal thickness (Woollard 1969a, 1970). Mean topographical elevations (h) have been calculated from 75 km by 75 km rectangles, their centers spaced at 50 km intervals. On the assumption of an Airy type of isostatic compensation in S. Norway, the depths to Moho (H) have been estimated. In Fig. 19a the relations $H = 33.2 + 8.5\ h$ (Woollard 1969a) and $H = 31.7 + 5.9\ h$ (Kanestrøm, pers. comm. 1972, calculated for the Fedje-Grimstad profile) are plotted as lines C_1 and C_2, respectively. The lower slope factor in Kanestrøm's relation implies an increase in density contrast between crust and upper mantle.
d) Gravity data, applying empirical relations between mean Bouguer anomaly and crustal thickness (Woollard 1959, Demenitskaya & Belyaevsky 1969); see Fig. 18.

Fig. 19a summarizes the results and compares the Moho depths as obtained from the four different sources (a–d). The depths are plotted along the regional profile IV (Plate 2) which is perpendicular to the Oslo Region and also crosses the southernmost part of the NORSAR. The various depth profiles conform in a general fashion but display some spread in the actual depths. The accordance will be improved by disregarding Woollard's isostatic profile (C_1). It is noted, however, that the c- and d-curves show a smoother appearance than the seismic Moho depths, especially in the Oslo Region. This may mean that different empirical relations (for c and d) apply better to Southern Norway, or that the Oslo Region is not isostatically compensated. It also reflects the nature of these methods; they will in practice tend to 'average out' all anomalies, having much the same effect as automatic smoothing of the gravity profiles.

Despite the rather similar Moho configurations, direct computation of a regional anomaly (I) does not seem justified on the basis of the more preliminarily known Moho depths *outside* the Oslo Region as compared to the more detailed NORSAR model and the consistent data along the Trøndelag-Oslofjorden profile (Kanestrøm 1969, 1973, Kanestrøm & Haugland 1971). Hence, on the assumptions of a crust/mantle density contrast of 0.45 g/cm^3 (Chapter 4), a regional anomaly (I) was computed only for the Oslo Region and with emphasis on the NORSAR model. Outside the immediate graben area, it was attached to the regional anomaly as obtained by the method II described below. Using the information from source (b), regional profiles (I) were constructed for all the 15 profiles in Plate 1 (see Chapter 7).

The regionals of the 15 profiles were combined to give a regional map (I) for the Oslo Region (Plate 3). This regional will largely reflect the gravity effect of variations in crustal thickness below the Oslo Graben within the uncertainty range of the seismic interpretation on which it is based.

Method II: Fig. 19b shows the configuration of the regional gravity (II) as determined by computer modelling in a trial and error procedure. For compari-

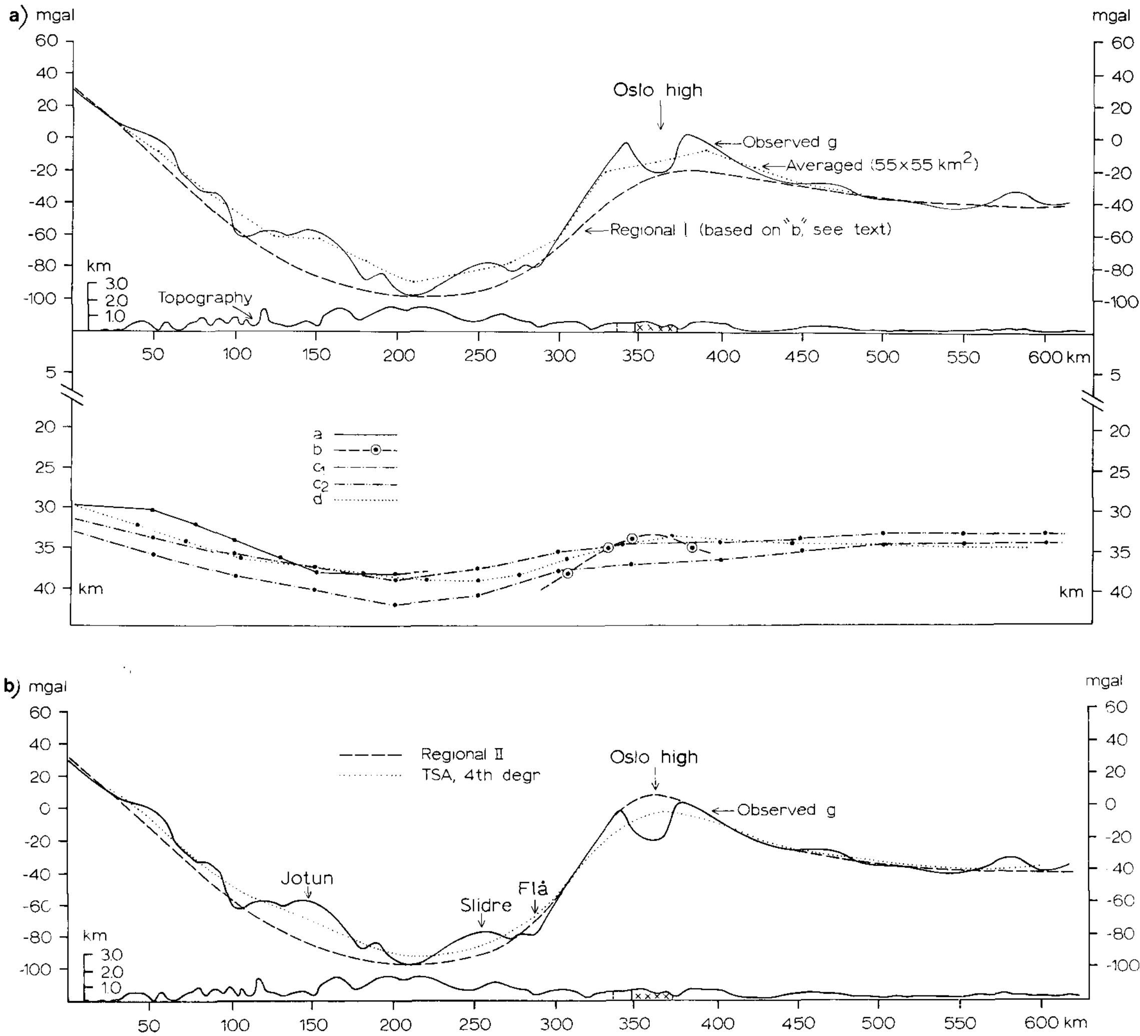


Fig. 19. (a) Various Moho depth estimates and regional gravity anomaly (I) along profile IV. Explanation for letters *a* to *d*, see text. (b) Various regional gravity anomalies along profile IV, see text.

son, the figure also shows a regional curve constructed by means of trend surface analysis (TSA, 4th degree). As can be seen, the automatic averaging process will in general cut all gravity highs and lows in an arbitrary fashion, smoothing out the gravity anomalies no matter what their cause.

The computer modelling was carried out by careful application of all available geological and geophysical information along each profile, the analysis being aimed to reveal whether, for instance, a gravity anomaly is caused by a local density increase or is simply the effect of an area of average density in between zones of lower than normal density. In Fig. 19b two local highs occur on each side of a sharp low above the Oslo Region. No rocks are found that can explain the highs, but the outcropping, light, syenitic rocks of the Oslo

Province are most likely the cause of the low. Hence the regional may be computed, as shown, as a curved line from the top of the one high to the other. Similar reasoning for the areas of the Flå granite (Smithson 1963b), the Slidre high (Smithson 1964) and the Jotun nappe (Smithson et al. 1974) have determined the direction and the magnitude of the regional anomaly (II); see also Section 7.2. In effect, regionals I and II are different only in the graben area.

The method II was applied to all the 15 gravity profiles traversing Southern Norway normal to the graben axis (Fig. 20). Regional II clearly reveals the broad gravity high above the Oslo Region and outside it in its axial direction. The combination of the profile regionals together with 3 profiles parallel to the rift axis, making a network of gravity profiles, gave the regional gravity map (II), Plate 4.

5.6 REGIONAL ANOMALY MAPS

The two contrasted approaches to regional-residual analysis have resulted in two regional gravity maps, each one revealing information about different parts of the deep Oslo Region. The regional anomaly map I (Plate 3) shows a smooth gravity high plunging gently towards NNE roughly parallel to the graben axis. The axis of the high is situated somewhat to the east of the graben axis, and the gravity gradient is larger on the west side towards the mountain chain than towards the east. If the anomaly, as intended, primarily reflects the variation in crustal depth, then the axis of the Moho bulge is shifted somewhat to the east relative to the graben axis. If rather constant mean crustal densities prevail in Southern Norway, then it is also implied that the Moho is dipping down to deeper levels to the west than to the east of the graben. However, seismic studies from the Precambrian shield in Sweden east of the Oslo Region indicate crustal thicknesses comparable to those to the west of Oslo, and also imply the possibility of an appreciably shallower Conrad or intermediate discontinuity in these regions (Dahlman 1971). Hence, large-scale, lateral, crustal density variations do occur, and the relatively thin crust below the Oslo Region is thickened by about the same amount on both sides.

In the case of regional anomaly II (Plate 4), the anomaly axis almost coincides with the graben axis. It also shows more variation in gravity gradients and some dependence on the local changes in graben outline, thus, indicating that parts of the anomaly result from a not too deep-seated mass concentration. One factor is the slight density contrast along the Kongsberg/Telemark region contact. The major cause is, however, to be sought at deeper levels. As a matter of method, the regionals were constructed so that the *difference* between regional I and II is ideally regarded to reflect anomalous dense masses *above* the Moho discontinuity, but *below* the lower level of the outcropping felsic bodies causing the local anomalies. This hypothetical mass has, on the average, to be situated directly below the graben.

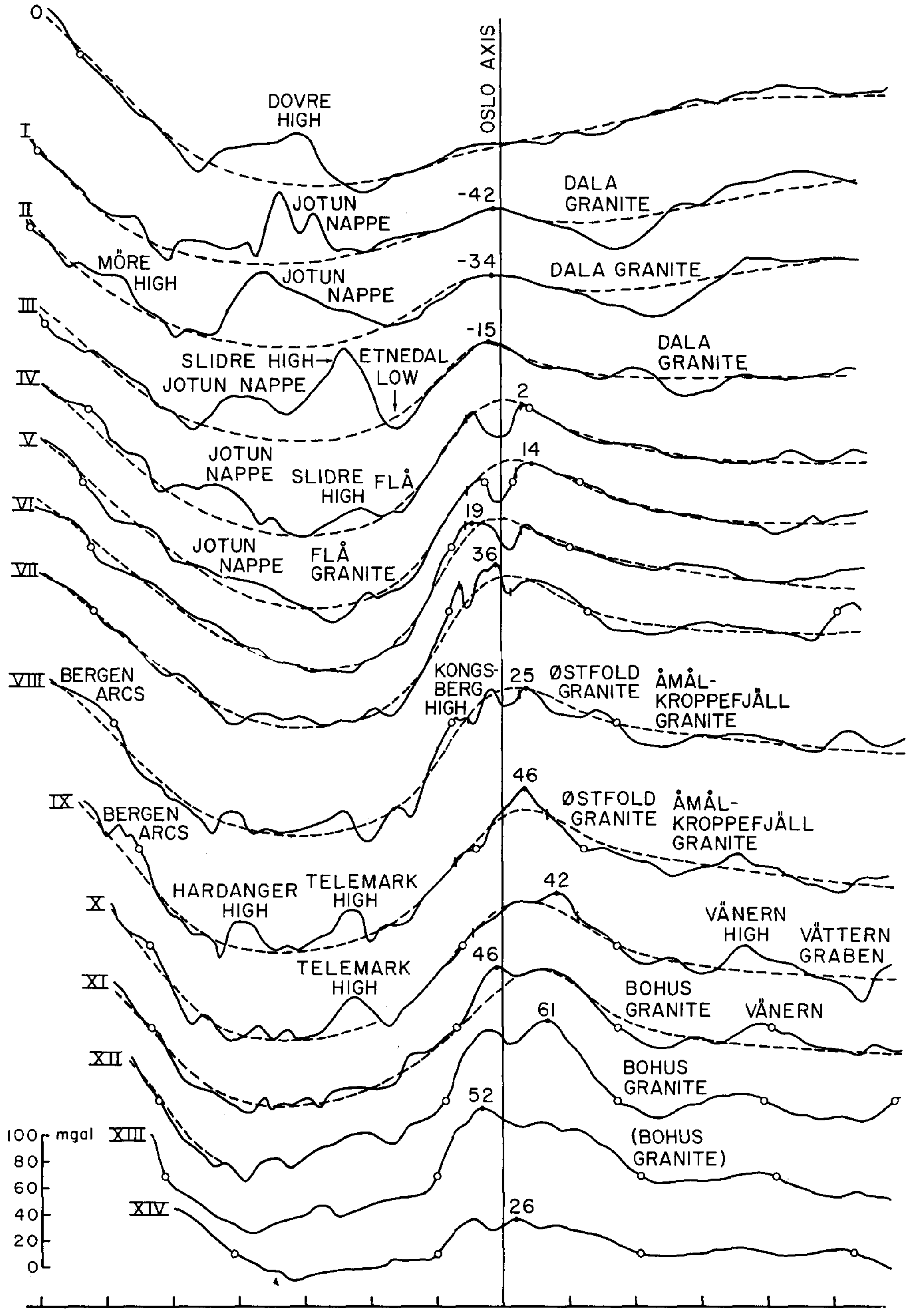


Fig. 20. Fifteen gravity profiles perpendicular to the Oslo Axis consistently show a broad gravity high of about 55 mgal above the graben, except for the profiles to the far north and south (location of profiles: Plate 1). Maximum gravity value above the graben (or its axial extension) is indicated for each profile. Zero anomaly values are marked with open circles. Dashed lines show regional anomalies obtained according to the method (II) described in text.

5.7 RESIDUAL ANOMALY MAP

Subtraction of the regional II (Plate 4) from the Bouguer map (Plate 2) revealed the residual anomaly map of the Oslo Region as presented in Plate 5. This map intentionally and ideally reflects the gravity effects of density anomalies in the uppermost crust. That this goal is close to fulfilled is indicated by (a) the fact that only a very simple regional contour map had to be applied, (b) the very close relationships between the many residual anomalies and the local geological features, and (c) a general lack of long-wavelength, typical regional gravity features in Plate 5 (with the exceptions noted below).

If, instead of subtracting the regional anomaly II, the regional I (Plate 3) had been applied, the resulting residual map would have included the effect of density variations also in the deeper parts of the crust.

Plate 5 shows that some rather long-wavelength anomalies do occur outside the graben area. They are of supposedly shallow origin since (1) they exhibit at places rather sharp changes in gradient, and (2) they coincide with observed petrographical changes at the surface. The anomalies are: (a) the Bamble trend parallel to the coast line and the 'Great Friction Breccia' in the southwestern corner of the map (see also Plate 1), and (b) the elongate gravity highs and lows in the Precambrian to the east of the graben, and partly also to the north-west. These alternating gravity trends coincide with structural and petrographic zones as previously described, and also with magnetic trends apparent on the NGU aeromagnetic maps over Southern Norway. The most marked magnetic trends are the ones from Indre Østfold-Ski-Østmarka which is also well-marked on the gravity map; another is a parallel trend SSE-ward from Minnesund, also associated with a slight gravity high.

Both on the gravity residual and on the magnetic maps two main trends predominate, ca. NNW and NNE. These trends occur both within and outside the graben. From a simple inspection of the residual maps, it seems that the NNW trending features can be traced from the Precambrian terrain into the Permian graben and then found again in the Precambrian on the other side. This is especially true for the Indre Østfold-Ski-Østmarka-Nordmarka-Hadeland trend. Monzonitic to dioritic inclusions occur within the syenitic rocks at the intersections between the graben trend and the NNW gravity ridges. Similar geometric and petrographical relations are apparent for the gravity lows, too (Plate 5), although exceptions or deviations from the strictly linear trends do occur. The significance of the above observations is uncertain. On the other hand, remelting and remobilization has been suggested as the chief cause of the Oslo Igneous Province (Schwinner 1928, Barth 1954). The remarkable gravity and magnetic trends observed might be interpreted as evidence in favour of this hypothesis which, however, is not supported by the overall evidence discussed in Chapter 8.

A number of positive and negative residual anomalies occur within the Oslo Region. As is commonly the case, rocks of granitic composition *(sensu lato)* produce gravity lows since they normally exhibit negative density contrasts with respect to the surrounding rocks in the range –0.05 to –0.2 g/cm^3 (Bott & Smithson 1967). From Table 5 it is evident that the average density contrast between the granitic to syenitic Oslo plutons and the Precambrian gneisses is about –0.12 to –0.13 g/cm^3. The six *major* negative anomalies (Plate 5) are therefore all associated with felsic intrusives; they are from north to south: 1) The Nordmarka-Hurdalen low above the composite syenite batholith in the northern half of the region, 2) the Finnemarka granite low, 3) the Glitrevann cauldron low, 4) the Drammen granite low, 5) the Eikeren low above the ekerite (soda granite) body, and 6) the low about the syenitic rocks just east of Skien in the Siljan district.

In general there is a very close connection between the outline of the outcropping intrusives and the gravity contour lines. Also, the isogals are usually much smoother than the sometimes irregular appearance of the contacts between intrusives and country rocks, suggesting that the exposed intrusives are parts of more regularly shaped bodies in depth than indicated by their accidental surface outline; see, for instance, the southernmost part of the Eikern low.

Major deviations are seen to the SE of the Nordmarka-Hurdalen low where the isogals imply a continuation of the syenitic rocks beneath a thin veneer of gneisses. In the inner part of the Oslofjord (the Nesodden district) an elongate gravity low, possibly continuous with the Drammen granite low, is not matched by any exposed rocks or structures that can adequately explain its existence.

Markedly in contrast to the residual lows within the Oslo Region, the major residual highs do not easily connect with any obvious geological features at the surface. Consequently, the anomalies must be caused by sub-surface mass concentrations, the depths to and geometry of which are subject to some degree of uncertainty. This applies to: 1) the Asker and Lier highs above the Cambro-Silurian some 15–25 km WSW of Oslo, 2) the Horten-Tønsberg high above the volcanics and monzonites along the west side of the Oslofjord, and 3) the Narrefjell high (at about 59° 25′ N and 1° 20′ W of Oslo) within the Skrim monzonite massif. If we consider the average densities of the Precambrian gneisses (2.74), the Cambro-Silurian (2.78) and the monzonites (2.71), the anomalies must be caused by rocks of dioritic to gabbroic composition.

Other local gravity highs are found connected with the Skien basalts and with several of the cauldron subsidences, that is the Sande, Drammen, Bærum, Nittedal and Øyangen occurrences. Gravity highs also mark many of the basic volcanic necks; however, the contour interval applied is commonly too coarse for these small-scale features, and detailed studies have been carried out on a number of them (section 6.4).

The density contrast between the monzonites and the surrounding gneisses (about –0.03 g/cm^3) is insignificant, and no major gravity anomalies are associated with the larger monzonite plutons in the southern part of the region except for the aforementioned Narrefjell high. Within the Larvik pluton, smaller gravity anomalies coincide with rock density variations (Plate 5 and Fig. 16) and infer petrographical variations in these areas within the complex. The gravity variations have not been worked out in any detail but seem to be a promising approach for future studies of the larvikite.

In the northern part of the region, the monzonitic to monzodioritic rocks occur as inclusions in the syenite rocks and stand out gravimetrically as small highs relative to the large syenite low.

Three prominent gravity highs occur just outside the Oslo Region or at the regional border. These are: 1) the Tyrifjord high, 2) the Kongsberg high and 3) the high above the Fen carbonatite complex. They are all located on a straight line parallel to the western border of the Oslo Region and coincide with the western line of volcanic necks of Permian age. Because of their possible connection with the Oslo Graben and associated rocks, these anomalies will be dealt with in some detail in the present paper (except for the previously studied Fen anomaly, Ramberg 1973).

A number of minor anomalies occur throughout the region but are not apparent with the degree of resolution offered by the 5 mgal contour interval of the gravity maps. Some of the minor anomalies, however, will be touched upon in the following interpretational chapter.

6. Gravity models of the residual anomalies

"The removal of regional effects is one of the two most important problems in gravity interpretation" (Dobrin 1960, p. 149). The other chief problem, the deduction of geological information from the residual field, is the goal of the present chapter. In the present study a third, equally important, problem will also be considered: the cause of the regional effect (Chapter 7).

A gravity interpretation falls into two parts: 1) Fitting simple geophysical models to the observed data, and 2) transforming the geophysical models into meaningful geological solutions. The first step, which involves solutions to the 'inverse' potential problem, is by nature ambiguous. The second step is based on available geological and other geophysical information which itself may involve a number of assumptions and implications. In summary, the gravity interpretations are inevitably ambiguous; their value and possible improvement is dependent on the amount and reliability of limiting controls from physical factors other than gravity potentials. Often, the chief value of such calculations is that they may "definitely throw out possibilities that previously had looked plausible and bring to mind unthought of new ones that are much more plausible" (Barton 1945, p. 64).

6.1 CALCULATION METHODS

A computer program by Talwani et al. (1959b) has been used for problems where a two-dimensional configuration was assumed. For some problems involving only two density values the two-dimensional program was made fully automatic by applying iterative processes to match the calculated and observed gravity profiles (Cordell & Henderson 1968, Henkel 1968). In cases of three-dimensional geological problems the program of Talwani & Ewing (1960) has been employed.

The two-dimensional programs have been applied for a series of problems, and to save time also in surveying clearly three-dimensional problems. Ideally, no geologic body fulfills the requirements of two-dimensional mass distribution. To demonstrate the error involved in assuming a two-dimensional structure, Bott & Smithson (1967) computed gravity anomalies for a number of two-dimensional and corresponding three-dimensional models, two of which are reproduced as Fig. 21. Bott & Smithson emphasize that (p. 863) "for any given cross-section, the two-dimensional model usually provides an underestimate of the thickness."

While a given gravity anomaly in general can be explained by a variety of structural models, it has been proved on the basis of Gauss' Theorem (Hammer 1945) that the magnitude of the causative anomalous mass can be uniquely determined from gravity data alone. For the anomalous mass (ΔM) and the total mass (M), respectively, Hammer (1945, p. 55) arrived at the following formulae:

$$(1)\quad \Delta M = \frac{1}{2\pi\gamma} \cdot \iint_P \Delta g \cdot ds, \text{ and}$$

$$(2)\quad M = \frac{1}{2\pi\gamma} \cdot \frac{\varrho_1}{\varrho_1 - \varrho_2} \cdot \iint_P \Delta g \cdot ds.$$

where γ is the gravitational constant, P is the plane or total horizontal extent of the anomaly over which the integration is to be carried out, Δg is the gravity anomaly within the segment ds, ϱ_1 is the density of the anomalous body, and ϱ_2 is the density of the enclosing material .The mass computation is done either graphically directly on the anomaly maps or by feeding the gravity anomaly into computer programs for two- or three-dimensional cases.

The critical point in the mass estimates is that the residual anomaly has to be correctly resolved. If, e.g., the residual anomaly reflects the net gravity effect of two bodies, one lighter and one heavier than the country rock, the formula only gives the *difference* of the two anomalous masses. The actual mass of a given body is, of course, also dependent on a correct estimate of the densities ϱ_1 and ϱ_2.

6.2 THE FELSIC INTRUSIVES

Granitic and syenitic rocks cover a total of about 3085 km^2, that is about 60 % of the area occupied by the Oslo Region plutonic rocks. An important question, however, is whether or not the observed, large *areal* extent of the felsic

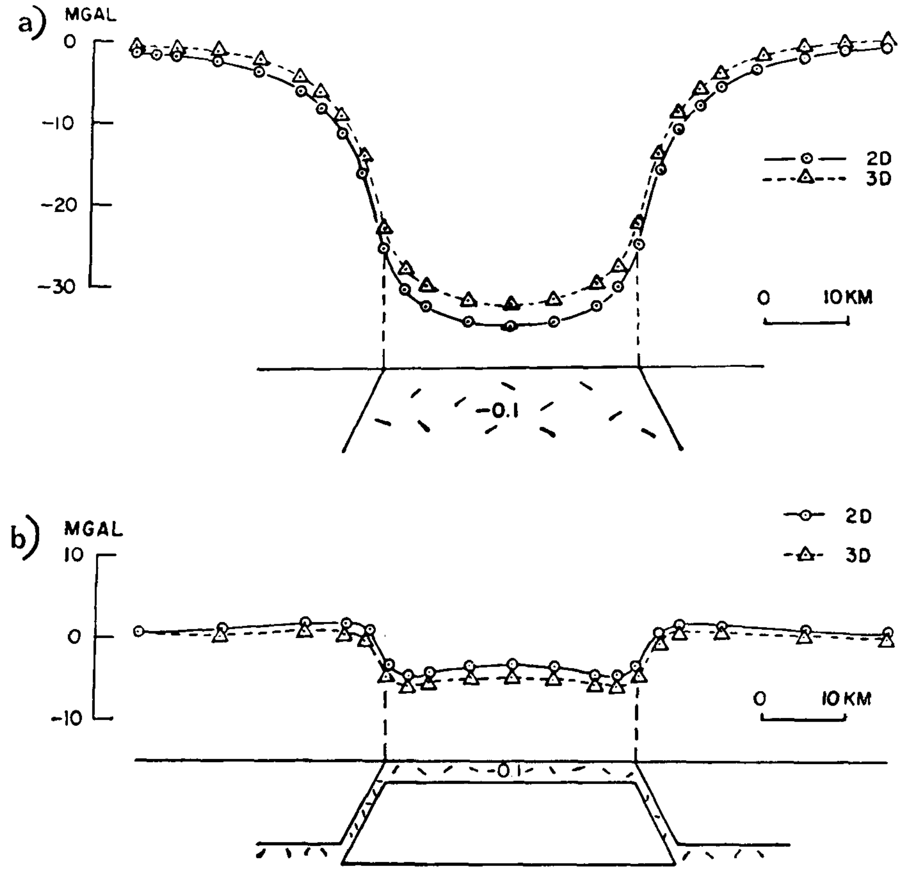


Fig. 21. Gravity profiles calculated for 2- and 3-dimensional models, for comparison. (a) Pluton represented by mass deficiency only, (b) Pluton with presence of subsided block (*after* Bott & Smithson 1967).

rocks is matched by a corresponding *depth*. The solution of this problem has both a structural and petrogenetic impact. It is also crucial when discussing the 'room problem' and the emplacement of the felsic intrusives.

The volume relations and mass distribution of the felsic rocks may be read from the characteristic gravity cross-sections. A few theoretical examples are briefly discussed (for a more thorough treatment, see Bott & Smithson 1967). Fig. 21a shows the typical section across a thick pluton represented by mass deficiency only, while Fig. 21b demonstrates the completely different anomaly encountered over rather thin plutons or plutonic rocks emplaced by cauldron subsidence and with foundered block present. The Oslo plutons seem almost ideal examples of post-kinematic intrusives being emplaced by piecemeal stoping. In accordance with Daly's (1933) idea that stoped granites have simply traded place with the country rock, Holtedahl (1952 p. 90) concludes with respect to the mechanism of emplacement of the Oslo batholiths that "huge subterranean crustal blocks sank to an unknown depth along curved fracture lines, with magma occupying the vacated space." This situation (Fig. 22a) will

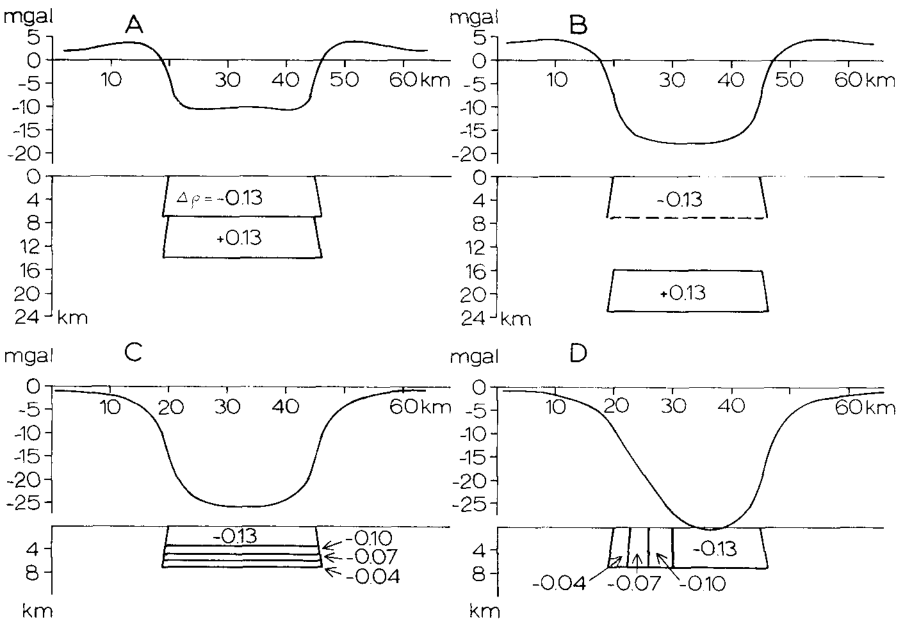


Fig. 22. Theoretical models of felsic plutons, (a) Pluton underlain by mass excess (e.g. stoped material), (b) Plutons underlain by mass excess at greater depth (e.g. residuum or stoped material). Note decrease in positive gravity effect compared to first case. (c) Pluton with horizontal density stratification. (d) Plutons with lateral density variation.

produce a flat-bottomed anomaly or, for big batholiths, one which shows convex upward-shaped central parts of the anomaly. Synkinematic, stoping intrusives may behave differently at greater depth, with forcible intrusion leading to shouldering and uplift of country rock (Buddington 1959). Less mass will consequently trade place with the ascending magmas, and Fig. 22b shows a model stoping at the top and accompanied by forcible intrusion at the lower level. The anomaly curve resembles the simple case in Fig. 21a. Somewhat similar anomalies would occur if the displaced mass in Fig. 22a sank to a greater depth or became dispersed over a larger volume. Whatever structure exists, if the displaced mass underlies the felsic rocks, the *felsic rock anomaly* must be greater than the observed value and the felsic rocks even thicker than the thickness encountered by direct computations as in Fig. 21a. The possibility of a density stratification with depth (Fig. 22c) may be difficult to distinguish by means of anomaly profiles from some of the other theoretical cases. Such stratification could represent a layered complex, a gradual vertical change in the density of the intrusion and/or country rock, or be due to an increased number of dense inclusions towards the lower level of the pluton. Lateral density variations will lead to asymmetric gravity profiles (Fig. 22d); the same effect might be caused by, e.g., an irregular distribution of denser xenoliths.

a. The Nordmarka-Hurdalen syenite complex

This huge composite batholith, forming the northern half of the Oslo Region, can conveniently be considered as one single body. Also, the residual anomaly (Plate 5) forms a continuous gravity low closely following its general outline but with some minor deviations mostly associated with local petrographical variations.

A series of gravity cross-sections through the syenite complex is shown in Fig. 23. The smoothest, most symmetrical lows are revealed by the central sections (profiles N3 and N4) where, according to the published geologic maps, the most uniform areas of the felsic pluton occur. In the northern (N1, N2) and the southern areas (N5, N6) larger portions of syenodioritic rocks are found together with intermediate to basic effusives and Cambro-Silurian inclusions. The maximum gravity anomaly varies from about –20 to –30 mgal in the various profiles, and reveals that the complex is of considerable thickness (compare with Fig. 21).

In areas of uniform density contrast the overall attitude of the contact between gneiss and felsic intrusives can be estimated from the shape of the gravity anomaly. If the anomaly over the contact is close to A/2 (A = max. grav. anom.) then the contact is nearly vertical; if it is <A/2, then the contact slopes inward, if >A/2 it slopes outward. From Fig. 23 it is clear that in profile N4 the 'contact anomalies' are almost exactly A/2 and that the gneiss/syenite contacts are consequently close to vertical, in perfect accordance with the general field impression. The same applies to N3, with a possible slight outward sloping at the east and a similar inward sloping at the west contact. An extreme outward sloping is indicated on the east contact of profile N5. This is in the area of Romerike-Nittedal where, e.g., the Holterkollen granite occurs outside the main complex and also ring faults are found in the Precambrian (Fig. 8). To the north, this test has to be applied with more caution. The gravity profiles N1 and N2 may be interpreted as indicating a pronounced inward sloping on the west side and a possible outward to nearly vertical contact in the east. However, a reduced lateral density contrast towards the west contact should give a gravity profile similar to that of an inward-sloping contact. The last alternative (with lateral density variations) is also the most likely interpretation of the west side of profile N5 (Nordmarka W); here the 'intermediate' pulaskites (Table 5) and monzonitic rocks predominate at the surface. The interpretation most likely also applies to the southernmost profile N6 which, however, interferes with local gravity highs (Tyrifjord and Ski-Ekeberg ridge), resulting in a somewhat confused gravity profile.

On the assumption of a two-dimensional mass distribution, various interpretations of the profile N4 are given in Fig. 24. The simple prism-like model (case I) does not appear to be applicable. A series of other models were calculated and a wedge-shaped model (Case II), obtained by a 'cut-and-try' approach, matched the observed data best. Dealing with an anomaly caused entirely by an *outcropping* body of known, *uniform* density and a *simple* form,

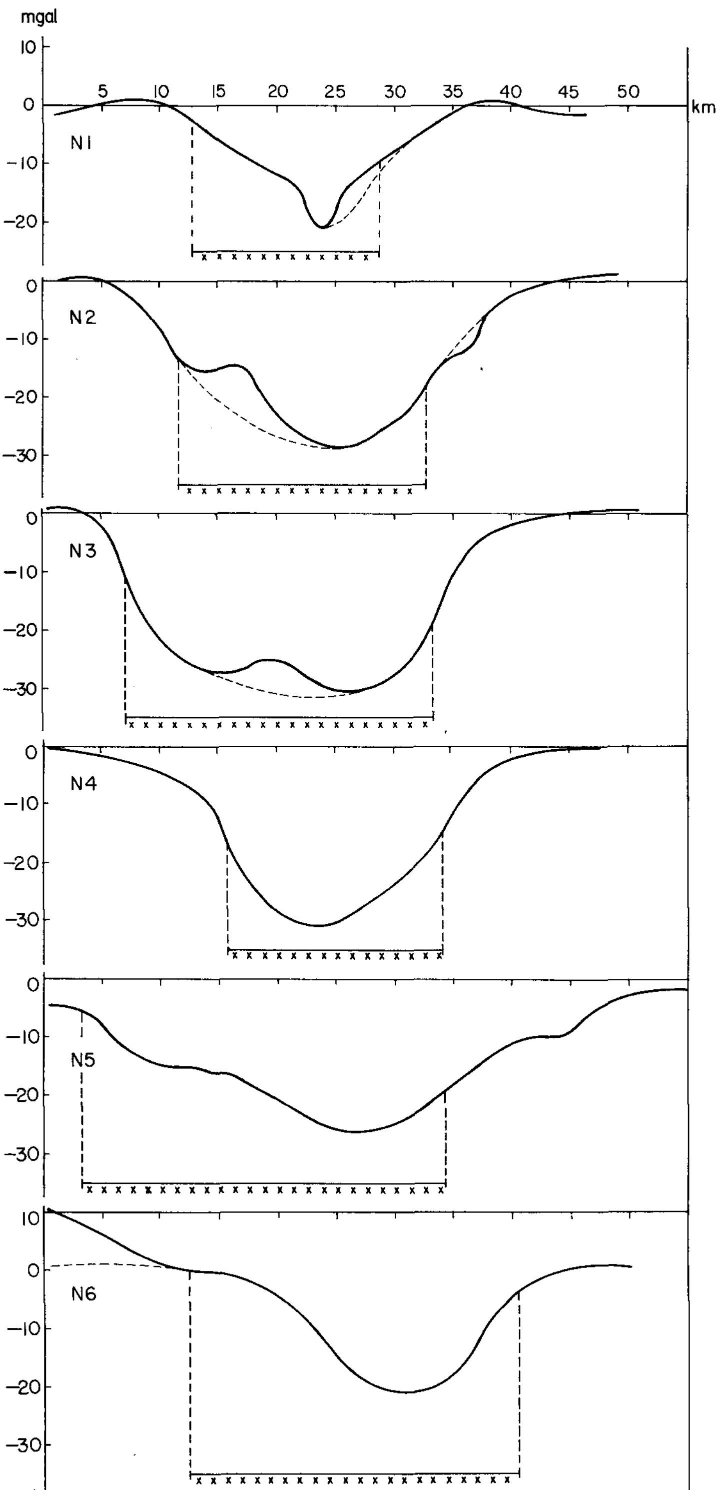


Fig. 23. Residual gravity profiles across the Nordmarka-Hurdalen syenite complex, regularly spaced at 12.5 km intervals; see Fig. 26. Profiles N2, N4, N6 are parts of the longer profiles IV, V, and VI, respectively.

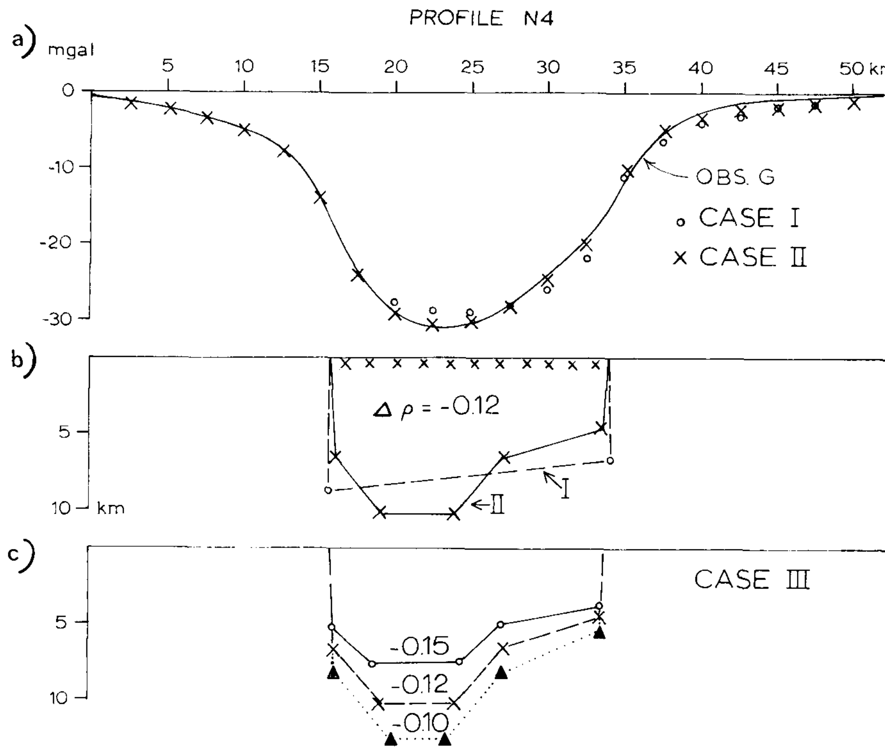


Fig. 24. Gravity interpretations of profile N4. Position of profile shown in Fig. 26.

a theorem by Smith (1961) shows that a body of only one shape could cause the anomaly. Thus, using the observed density contrast (0.12 g/cm^3), it seems reasonable that the root-like depression is real and (considering the two-dimensional approach) that it reaches down to at least 10 km below sea level. However, since the density has been determined from surface samples only (Fig. 13), the effect on the model of possible changes of the overall density contrast is depicted in Fig. 24, case III. A possible slight increase in the density of the Precambrian gneisses with depth (see Table 8) is likely to be at least compensated by a corresponding vertical density increase of the miarolitic syenitic rocks. Also, any petrographical variation in the intrusive complex at depth will include more dense rock types. Both these hypothetical phenomena would eventually result in a reduced density contrast and an accordingly increased thickness of the batholith.

In Fig. 25 various models of the profile N3 are presented. Again, the wedge-shaped model (Case II) gives the best fit between the calculated and the observed gravity data. In the central part of the profile there is a local high not matched by any exposed petrographical variations in the surface rocks. The outline of the profile somewhat resembles the theoretical case (Fig. 22a) with excess mass (stoped material or residuum) below the felsic rocks. However, this high, which reveals rather sharp gradients, has to be accounted for by a quite shallow, dense mass (depth to top of the body less than 1.5–2 km ac-

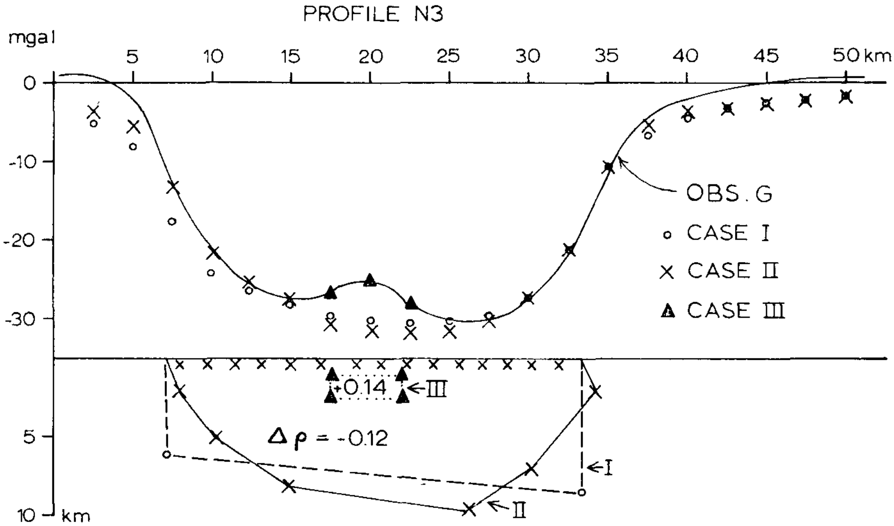


Fig. 25. Gravity interpretations of profile N3. (See Fig. 26).

cording to the maximum depth formulae of Bott & Smith 1958). Case III represents a solution with the top of the body at 1 km depth, totally enclosed in syenitic rocks. The density has been chosen as that of syenodioritic rocks, and the hypothetical body is on an almost straight SSW line with the outcropping masses of syenodiorites at Høvernsjøen and Skrukkelia west of Hurdal.

Based on the foregoing discussion, a three-dimensional model has been computed by means of trial and error for the whole Nordmarka-Hurdalen complex, applying the observed density contrast of –0.12 g/cm^3. In the NW part of the area a subsurface body of density 2.76 g/cm^3 has been added. The model (Fig. 26a) that gives the best fit (Fig. 26b) with the residual anomaly (Plate 5), is funnel-shaped with its deepest part (contour VII) reaching down to about 12 km in the central area.

If other density contrasts between gneiss and intrusive complex are chosen, the contour lines will mark different depth levels. With density contrasts of, e.g., –0.15 and –0.10 g/cm^3, respectively, the contour line VII represents approximately the 10 km and 15 km depth levels.

The depth contour lines in Fig. 26a closely reflect the general outline of the complex in most areas, but indicate a marked inward sloping from the surface and down in the NW (Toten) and the SW (Nordmarka W) areas, and a conspicuous outward sloping in the Nittedal-Romerike area to the southeast. Since the three-dimensional model was only designed to depict the configuration of a hypothetical body of uniform density contrast (–0.12 g/cm^3), it seems appropriate to point out some alternative solutions for these anomalous regions.

In Fig. 27, case I represents a simple two-dimensional model for the gravity profile, N1, similar to the three-dimensional model. A better fit is obtained by introducing local geological features as observed in the field (case II). The

a)

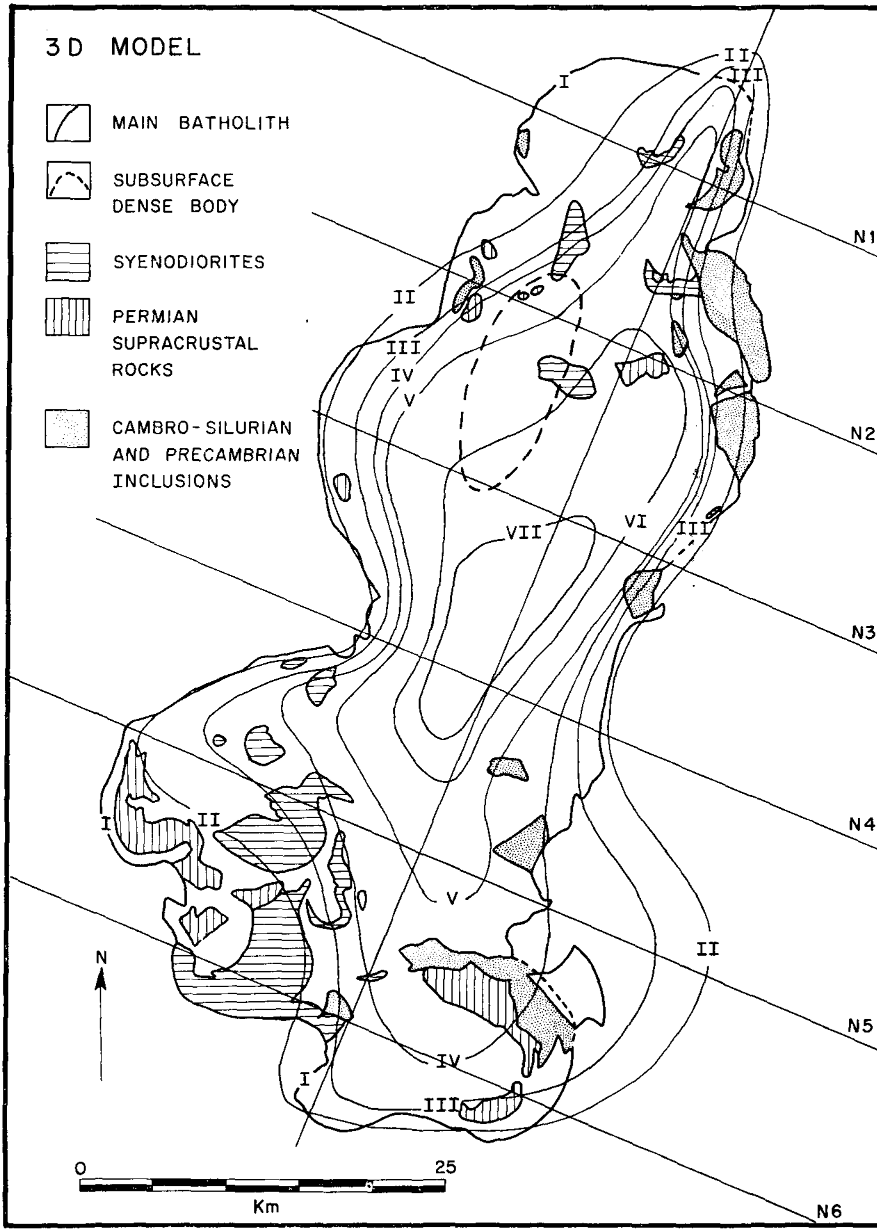


Fig. 26a. Three-dimensional interpretation of the Nordmarka-Hurdalen syenite complex. Geophysical model. For the various density contrasts (–0.10, –0.12, and –0.20), the depth contours are (in km): I = 0 (surface outline), II = 3, 2, 1; III = 5, 3, 2, IV = 8, 5, 3; V = 11, 7, 4; VI = 14, 10, 5; VII = 15, 12, 6. Subsurface dense body located at 0.5 to 5 km depths with a density contrast of 0.14 g/cm³ to syenite.

root-like depression will in this case reach down to about 10 km. Alternatively, case II could be reinterpreted by assuming lateral density variation, as in the theoretical example, Fig. 22d. In case IIIA, the possible lateral variation is explained by the occurrence of syenodioritic rocks at depth on both sides,

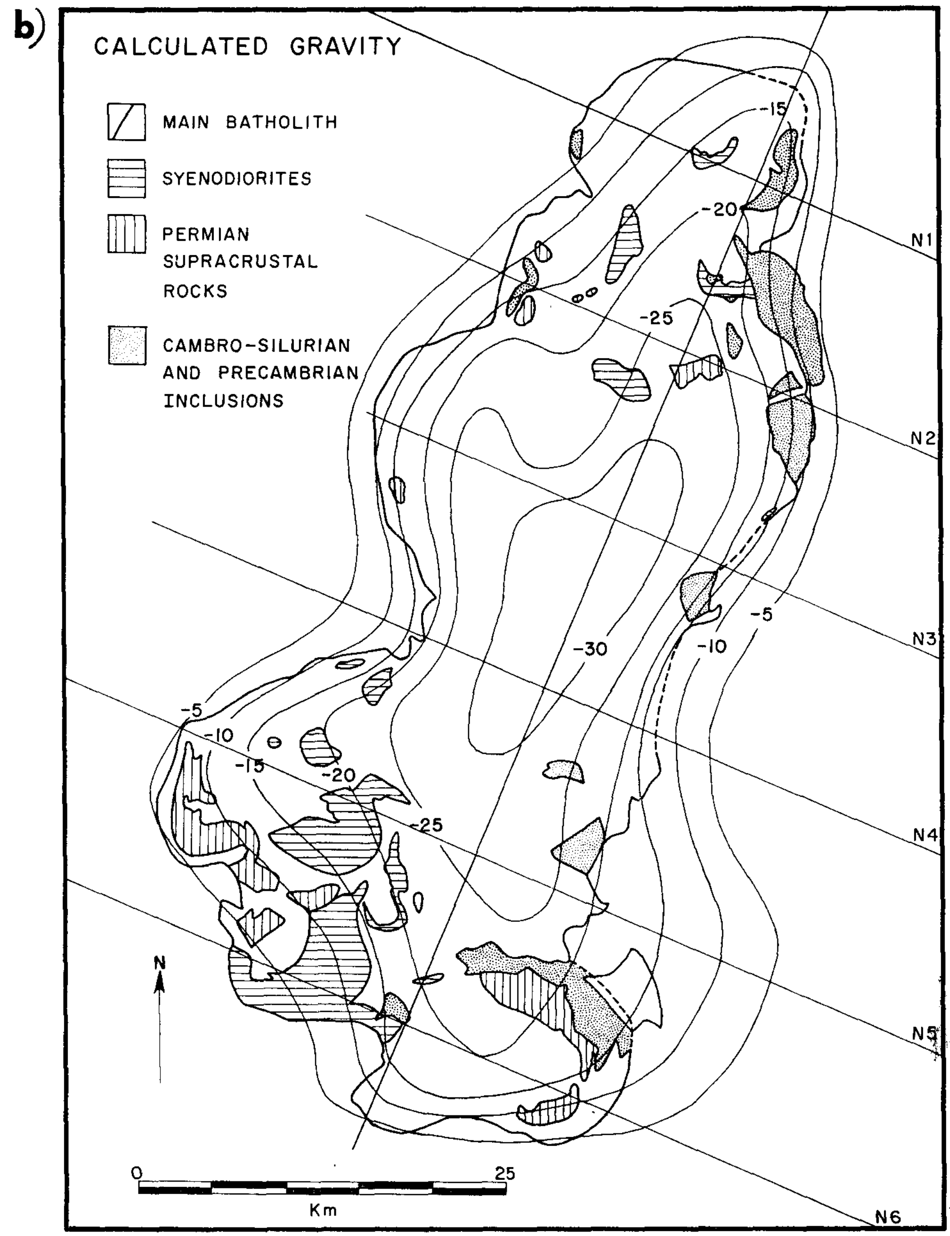


Fig. 26b. Calculated gravity shown for a density contrast of –0.12 g/cm³ (equiv. to observed contrast). For the actual calculations the body was subdivided into 14 lamellae.

thereby forming a zoned intrusive complex of a more simple geometric net form than in case I and II. Examples of a normally zoned intrusive complex in the Oslo province are found, e.g., in the Finnemarka complex, the Sande cauldron central intrusion, and in several concentric complexes in the Nordmarka-Hurdalen batholith.

Alternative case IIIB (Fig. 27) explains the possible lateral density variation

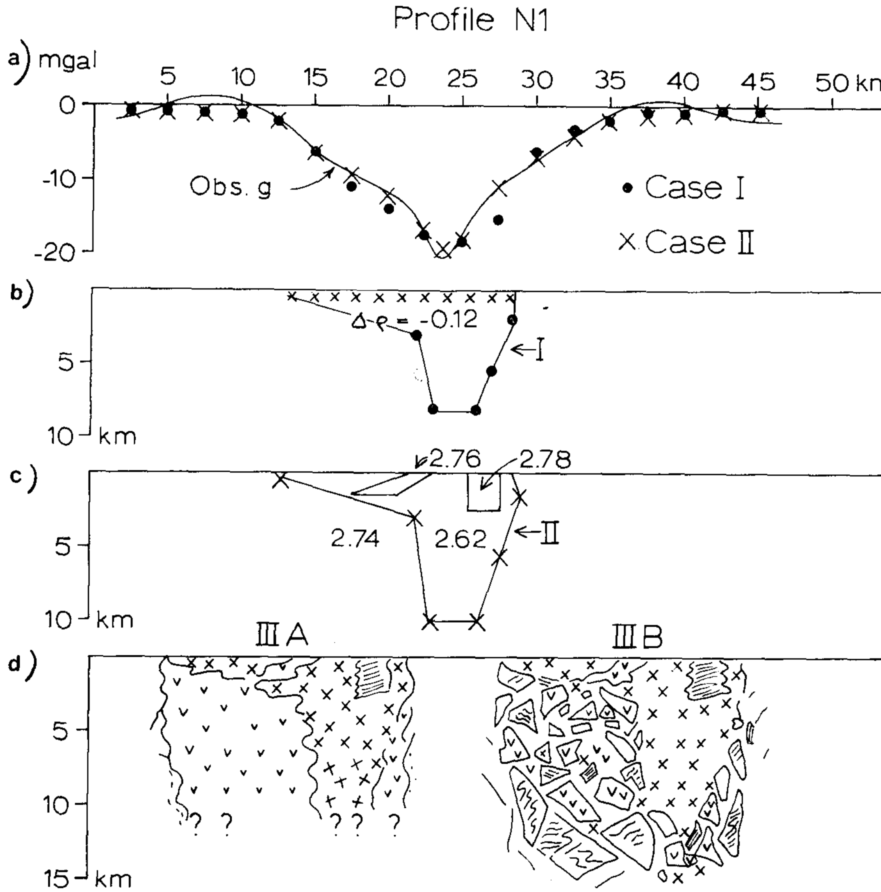


Fig. 27. Gravity and geological interpretations of profile N1 (see Fig. 26).

as being caused by the accumulation of stoped blocks of earlier crystallized syenodiorites, Cambro-Silurian and Precambrian rocks. This alternative offers, in the author's opinion, the most likely explanation of the mode of emplacement of the syenities; it creates a smaller room problem and it is in harmony with the general observation that at least the upper portions of the Oslo plutons primarily ascended by means of magmatic stoping.

Similarly, the profile N5 (Fig. 28) can be satisfied either by an irregularly shaped body of uniform density (case I) or by models assuming horizontal changes in density towards the west, as exemplified by case II. The west part of the profile passes over Nordmarka W with large monzonitic areas (e.g., at Katnosa and Gjerdingen), pulaskites etc., and, further to the west, over (the northern tip of) the Øyangen cauldron remnant close to the contact. Gravity and magnetic studies of the Gjerdingen occurrence (Grønlie 1971, Kristoffersen 1973) indicate that the density and magnetic contrast continues to a depth of about one kilometer below the surface. Stratigraphical studies in the Øyangen cauldron indicate possible lava thicknesses of up to 3000 m (Sæther 1962),

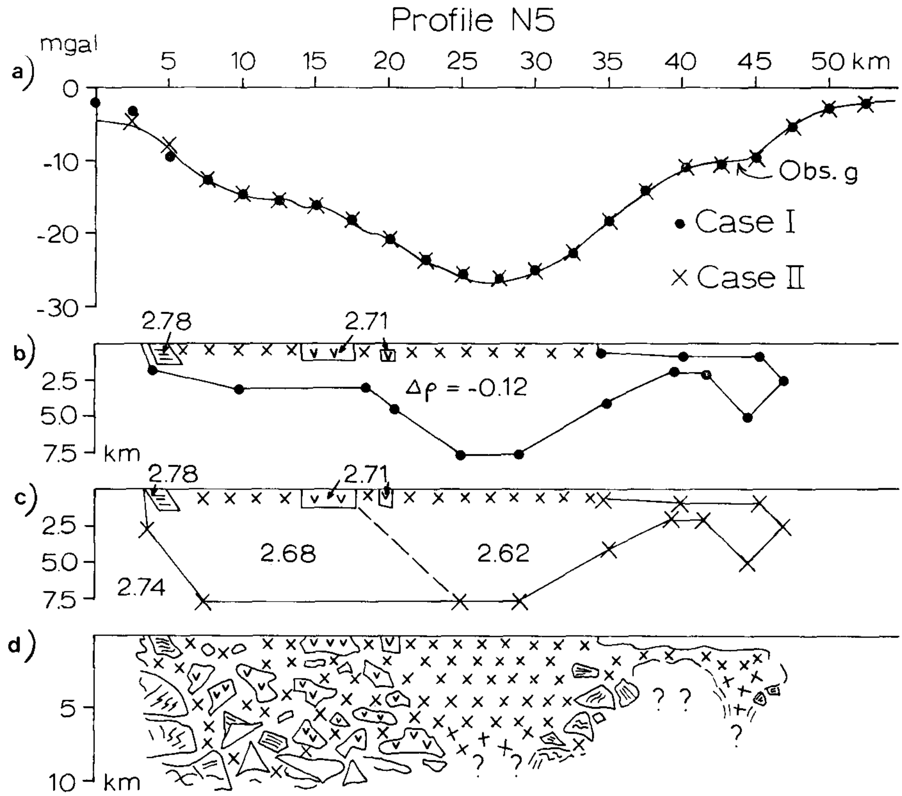


Fig. 28. Gravity and geological interpretations of profile N5 (See Fig. 26).

implying considerable depths also for the enclosing syenitic to monzonitic material. The geological interpretation of the given geophysical models is critical. One possible solution is offered in the lower part of Fig. 28; here, it is assumed that the western part of the section has a somewhat higher average density because of the presence of numerous monzonitic and other rock fragments within the syenites.

The east part of the profile (Fig. 28) reveals large masses of felsic rocks far outside the surficial contacts. Further, with the applied density contrast (–0.12 g/cm^3), batholiths up to 5 km thick may exist beneath the ring fractures in the Precambrian to the east. The exact configuration of these masses is uncertain. They may be continuous with the main complex to the west (as in the models) or represent separate intrusions. The nearby Holterkollen granite complex also comprises different rock-types, including monzonites. In the model, such rocks will give even larger thicknesses. The same effect would be caused by the presence of xenoliths. The Holterkollen complex exhibits extreme outward sloping of the contacts to the east. The gravity gradients (Fig. 28 & Plate 5) show that the subsurface felsic rocks extend to about 7–8 km east of the complex, the total section being considerably wider than that of the surface exposure of the igneous province.

In conclusion, the gravity anomaly shows that the outcropping Nordmarka–Hurdalen syenite complex extends downward to maximum depths of about 12 km when a constant density contrast of -0.12 g/cm^3 is assumed. A body with the said uniform density contrast has a wedge- or funnel-shaped geometry.

From the residual anomaly (Plate 5) the mass anomaly represented by the syenite complex has been estimated to $\Delta M = 11.8 \cdot 10^{17}$g. With the observed density contrast of $\Delta\rho = 2.74 - 2.62$ g/cm^3 $= 0.12$ g/cm^3, it is estimated that the total mass is about $M = 2.57 \cdot 10^{19}$g and the total volume is $V = 98.09 \cdot 10^{17}$cm^3 (9809 km^3).

It is generally accepted that the plutonic series rose chiefly by piecemeal stoping; inclusions from the size of a few millimeters to several kilometers are found throughout the complex and especially closer to the contacts. One might speculate, therefore, that because of temperature distribution (lower toward the contacts) and viscous drag, the stoped blocks will tend to slow down and accumulate along the sides of the ascending batholith. This will affect all the models discussed (Figs. 23 to 28) where wedge- or funnel-shaped configurations have been shown, and which were based on a uniform density contrast. A reduced density contrast toward the walls will lead to a widening of the models with depth toward that of a vertical prism as shown in the geological interpretations of Figs. 27 & 28.

Because of the implied process, stoped blocks might be expected to reside also in central parts of the batholith but here generally at deeper levels. This situation is somewhat similar to the vertical density stratification shown in Fig. 22c, and the depths calculated in the models (e.g., Fig. 26a) will consequently represent minimum values. There will be no sharp floor to the batholith which most likely grades into a mixture of syenitic rocks and more basic derivatives together with stoped blocks and *in situ* gneiss fragments. Since the large majority of the xenoliths have densities close to that of the Precambrian gneisses, this interpretation will not violate the mass calculations above. The likely occurrence of subcircular, old monzonitic complexes in the Nordmarka and Hurdalen areas implies that the emplacement of the Nordmarka-Hurdalen syenite complex was preceded by high-level intrusions of more mafic plutons.

b. The Drammen granite complex

The subcircular Drammen complex of various granites and porphyries occupies a rather central position in the Oslo Region around the town of Drammen (Fig. 2). The residual gravity anomaly (Plate 5) clearly reflects a mass deficiency as expected from the observed density contrast of -0.13 g/cm^3 between the granites and the Precambrian gneisses. A local gravity high, associated with the Drammen cauldron subsidence within the complex, is considered separately (p. 106).

The area of maximum anomaly (approx. -20 mgal) corresponds roughly to an area of high K_2O and Th content (Raade 1973) and to an area of medium-

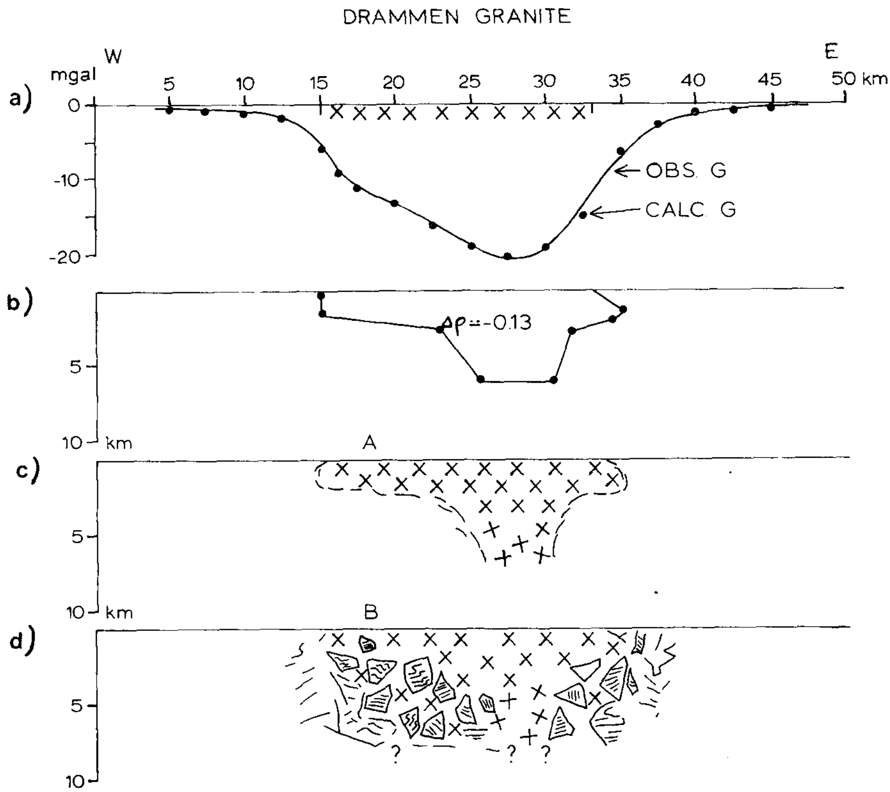


Fig. 29. Gravity and geological interpretations of an E–W profile across the Drammen granite complex.

to fine-grained granite with a relative abundance of quartz (El Bouseily 1971, unpubl.). The gravity low is, however, not caused by a local density reduction since it is not associated by any significant variations in the horizontal density distribution (Fig. 16). Hence the low in the east central part of the granite complex must largely be due to an increased thickness of granitic material in this area.

The residual gravity anomaly clearly shows that the granite occurs at shallow levels below the Cambro-Silurian and Downtonian sedimentary beds of the Konnerud area SW of Drammen. Similarly, the granite extends beyond its surface borders underneath the Precambrian in the Røyken-Slemmestad area east of the granite. The anomaly is also continuous with the gravity low along the Nesodden peninsula on the east side of the Oslofjord, but more data are needed in the Oslofjord area and on the islands in order to define this anomaly more clearly, and no definite solutions can be offered for its cause at this stage. It is possible that the elongate low marks a subsurface extension of the Drammen granite in a NNE direction parallel to the Nesodden and other fault lines. The inner Oslofjord seems to define a minor graben by itself, and the gravity low may alternatively be due to the mass deficiency caused by the local subsidence.

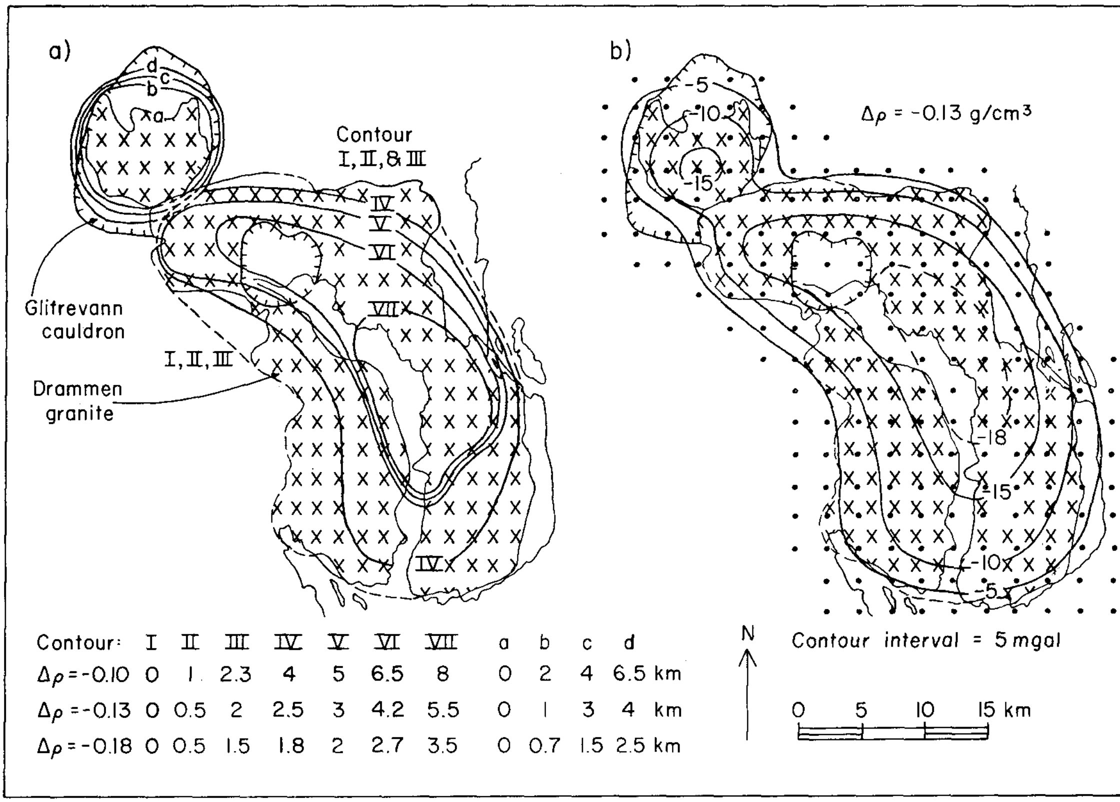


Fig. 30. Three-dimensional interpretation of the Drammen granite complex and the Glitrevann cauldron central pluton. (a) Geophysical model. Contour lines I, II, and III, identical to surface outline of the complexes except where dashed, (b) Calculated gravity anomaly (for obs. $\Delta\rho = -0.13$ g/cm^3). The denser rocks of the Drammen cauldron have been disregarded in these calculations. For the calculations, the Drammen pluton was sliced into a stack of 19 horizontal lamellae, the Glitrevann pluton into 14 lamellae. Black dots indicate positions of calculated gravity values.

On the assumption of a uniform density contrast of -0.13 g/cm^3, simple two-dimensional (Figs. 29 & 51) and three-dimensional (Fig. 30) geophysical models suggest that the Drammen granite complex is a thin (2–4 km), tabular intrusion with a rather wide, vertical extension in the east central part. Another possible vertical extension is located to the west of the Drammen cauldron subsidence. Fig. 30 also illustrates the changes in thickness within the rather extreme density contrast range of -0.10 to -0.18 g/cm^3.

The irregular floor of the models is compatible with a funnel- or mushroom-shaped body with feeder channels located below areas of minimum Bouguer anomalies (interpretation A, Fig. 29c). Considering its large horizontal extension, rather moderate thickness and general form, the structure is not very dissimilar to a laccolith, a form originally suggested for the Drammen granite by Brøgger (1890, 1933a); but the cross-cutting relations to the surrounding Precambrian gneisses suggest that tabular batholith would be a better term. However, in accordance with the Nordmarka-Hurdalen syenite complex interpretation, the Drammen granite complex may alternatively be interpreted as a major batholith with an increasing quantity of stoped blocks accumulated below surface and towards the contacts (B, Fig. 29d). In the latter case, no sharp boundary is expected to exist; the relatively homogeneous granitic body grades into a mixture of xenoliths and intrusives, the overall density approaching that of the surrounding gneisses. As with the syenite complex, the central 'roots' may in any case represent major avenues of ascent and represent a measure of the width of real crustal separation in that area. Hence, the emplacement of the large felsic bodies may be ascribed to a combination of dilatation and stoping.

In conclusion, the Drammen granite complex is interpreted as a subcircular batholith grading downward into a mixture of stoped blocks and intrusives. Alternatively, it is a tabular or mushroom-shaped intrusion floored by Precambrian gneisses and with one or more root-like extensions. In either case, disregarding the questionable Nesodden low, the anomalous mass is estimated to about $\Delta M = 2.35 \cdot 10^{17}$g, that is approximately one fifth of the syenite massif. With the observed density contrast of -0.13 g/cm^3, the total mass is estimated to $M = 4.73 \cdot 10^{18}$g and the corresponding volume to $V = 18.11 \cdot 10^{17}$ cm^3 (or 1811 km^3).

c. *The Finnemarka granite complex*

This subcircular intrusive complex crops out some 25–40 km west of Oslo. It consists of a large core of biotite alkali granite ($\rho = 2.61$ g/cm^3) fringed by rocks of intermediate composition in the west, north and south (Fig. 2). The intermediate rocks have been subdivided into an inner zone of 'kjelsåsite-larvikite' (Brøgger 1933a) or 'granodiorite' (Czamanske 1965) having an average density of 2.71 g/cm^3 (7 samples) and an outer zone of 'akerite' of average density 2.77 g/cm^3 (3 samples). Both rock-types may be classified as monzodiorites according to Streckeisen (1967). The contacts between the three petrographical varieties are sharp and well-sealed; inclusions of the more mafic types are found in the more felsic intrusives. Porphyric varieties have been described by Czamanske (1965) from one locality but seem almost ubiquitous along the southern border regions where no rim of intermediate rocks occurs. Various aspects of the petrogenesis of the complex have been discussed by Czamanske (1965), Czamanske & Mihálik (1972), Czamanske & Wones (1973), and Raade (1973).

An E-W gravity profile (Fig. 31a) has been drawn across the Bouguer map of Fig. 33. The regional gravity profile has been taken from the regional map (Plate 4) to produce the residual anomaly that reaches a maximum of about -15 mgal over the central part of the complex (Fig. 31b). The west flank of the profile interferes with the Tyrifjord high, while the east end of the profile is afected by a marginal part of the Asker-Lier high.

The residual anomaly clearly reveals the extreme outward sloping border especially to the east of the granite, as observed by Brøgger (1933a) and others.

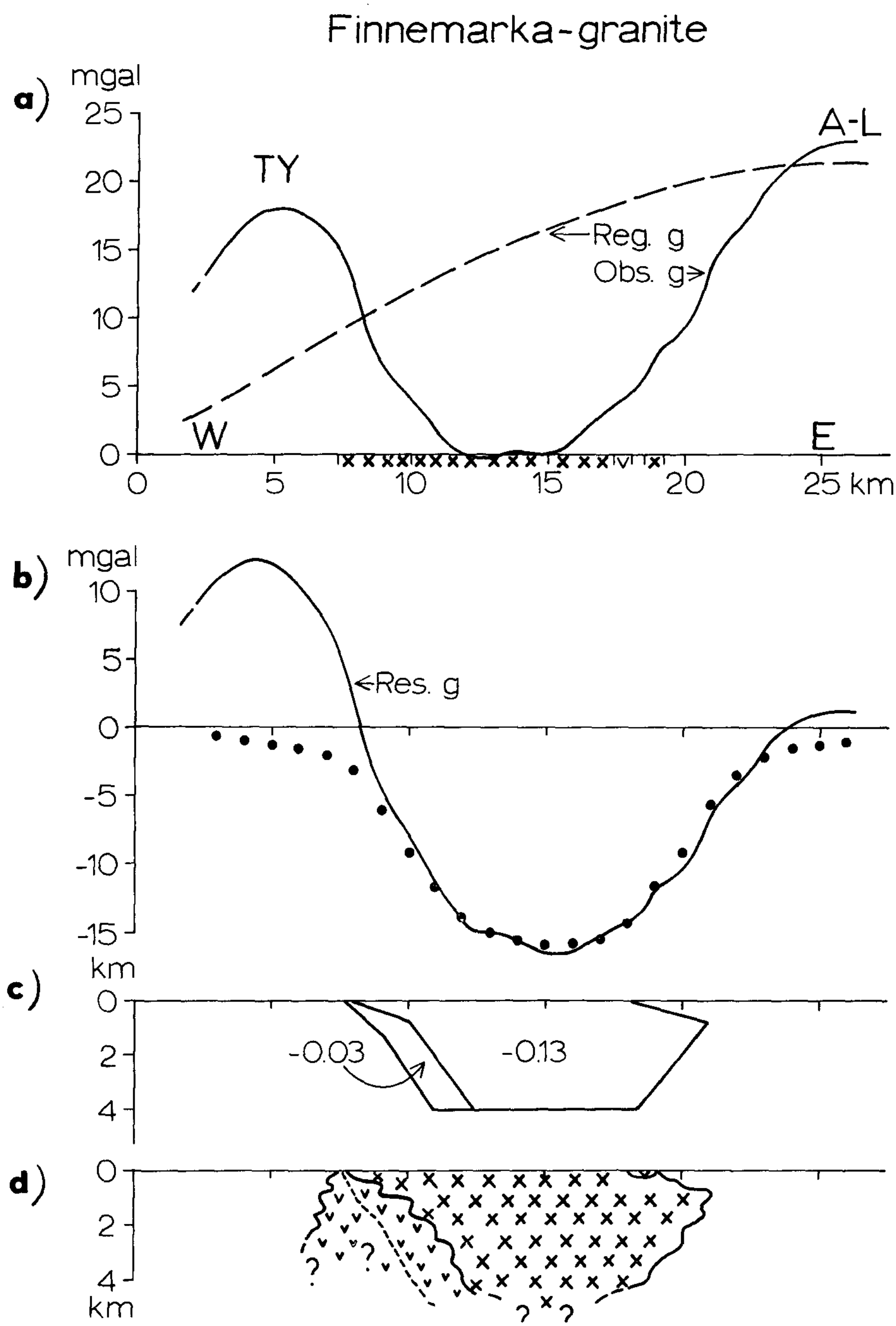


Fig. 31. Gravity and geological interpretations of an E–W profile across the Finnemarka granite complex. Regional anomaly taken from Plate 4. TY = Tyrifjord, A–L = Asker–Lier.Position of profile shown in Fig. 33.

A similar outward sloping has been observed for the monzodiorites to the west (Czamanske 1965). Here, however, the gravity anomaly implies an inward

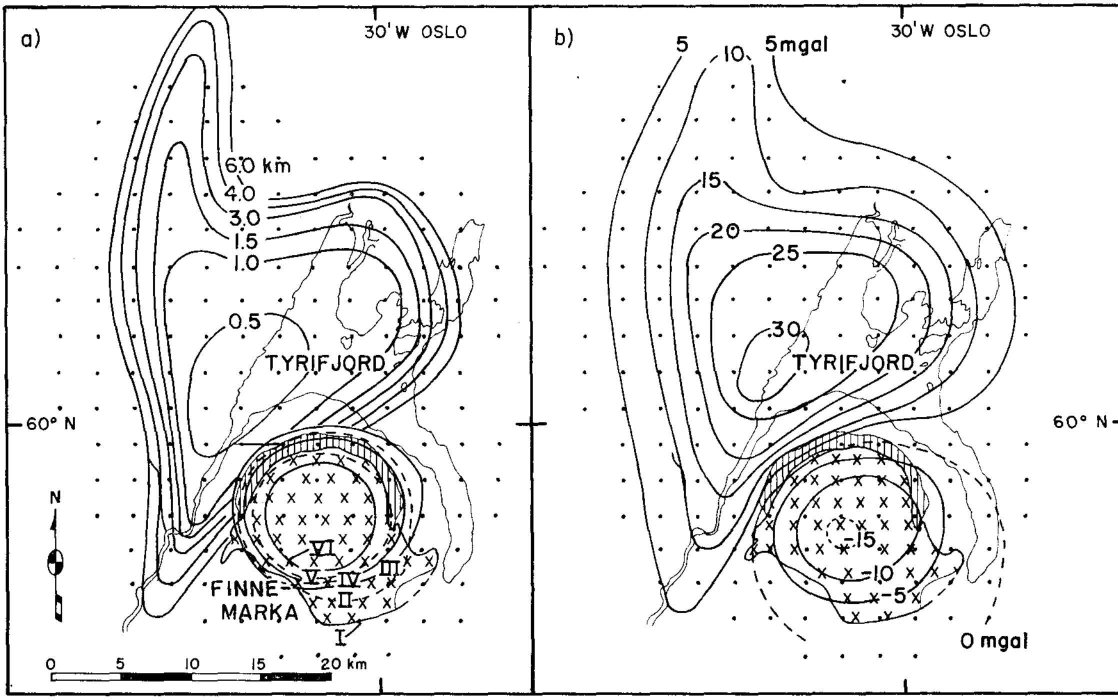


Fig. 32. Three-dimensional model of the Finnemarka granite complex, and including the adjacent Tyrifjord anomalous body. (a) Geophysical model. For the observed density contrast (–0.13 g/cm^3) the contour lines represent the following depths (in km): I = 0, II = 0.5, III = 1.0, VI = 3.0, V = 4.5, VI = 7.5. For density contrast of –0.1 and –0.15 g/cm^3, contour line VI represents the 9.5 and 5.2 km depth level, respectively. (b) Calculated gravity anomaly (for obs. $\Delta\varrho$ = 0.13 g/cm^3). For the calculations, the Finnemarka pluton was subdivided into 21 lamellae, the Tyrifjord body ($\Delta\varrho$ = + 0.2 g/cm^3) into 14 lamellae. In both figures, exposed granite is shown by X'es, monzodioritic rim by vertical ruling. Black dots indicate positions of calculated gravity values.

sloping of the lighter material, but the same gravity effect may partly be due to lateral density changes because of the observed petrographical zoning. Because of the interference with the Tyrifjord high and the rather small density contrasts between the monzodiorites and the adjacent Precambrian gneisses, no quantitative estimates have been attempted for the more dense rocks of the complex except for the simple two-dimensional model of Fig. 31 indicating a widening of the monzodiorites with depth at the expense of the granites. To the east (and certainly also north-east) the monzodiorites have to be cut by the underlying, outward sloping granite surface.

From the three-dimensional models in Fig. 32 it is seen that a mass of uniform density contrast of –0.13 g/cm^3 will extend downward to ca. 7.5 km below sea-level. Other density contrasts give different thicknesses but fit the 'observed' residual less well. In any case, the striking concentric configuration of the subsurface contour lines shows that the shape of the Finnemarka granite may be that of a simple cone. However, considering the presence of petrographical zoning and the possibility of subsurface xenoliths along the flanks, the complex may in fact be cylindrical or even widening with depth.

A lower density contrast in the deeper part of the complex may be caused by (1) an accumulation of stoped blocks, and/or (2) a gradual change in composition of the felsic intrusive. Both effects lead to vertical density stratification in the lower part of the batholith which then might possibly be considerably thicker than what has been estimated for the rather uniform granitic material.

In conclusion, the gravity data have put restraints on possible configurations of the complex. The subcircular complex is not a shallow laccolith-like pluton, nor is it underlain by a shallow subsided block. It represents a large mass deficiency and certainly extends to considerable depths. By assuming a uniform density of 2.61 g/cm^3, its thickness is 7.5 km.

From a residual map constructed on the basis of the detailed observations in Fig. 33, the mass anomaly of the Finnemarka granite has been calculated to $\Delta M = 0.44 \cdot 10^{17}$ g. Considering the areal extent of the granite outcrops, this number seems small compared with the estimate for the Drammen granite ($2.35 \cdot 10^{17}$ g). This emphasizes the likely cone-shaped configuration of the Finnemarka *granite* as well as the tendency for the Drammen granite to extend beyond its borders in the subsurface at many places. Also, the southernmost part of the Finnemarka granite is apparently very thin (Fig. 32). By applying the observed density contrast (2.74 – 2.61 g/cm^3 = 0.13 g/cm^3), its total mass and volume has been estimated (Table 9, p. 83).

The monzodioritic rim of the complex may be ascribed to assimilation of Cambro-Silurian and Precambrian rocks by the granite magma. According to recent geochemical data, this process seems unlikely since the monzodiorites have approximately the same, low Th/U-ratio (mean 3.57) as any other kjelsåsite-larvikite rocks in the Oslo Province (Raade 1973). More likely, the contact relations and types of inclusions imply that the complex formed by multiple intrusion, the various petrographic types representing members of a differentiation series as suggested by Czamanske (1965). Since the various members of the complex have been emplaced largely by stoping and only to a minor degree by forcible intrusion, the stoped blocks of the original monzodioritic intrusion and other rock fragments must have been settled along the margins and below the 7.5 km depth level. The Finnemarka complex is, therefore, seemingly not underlain by a solid block of Precambrian gneisses, but by Permian intrusives and older rocks with an overall intermediate composition.

d. The alkali granite (ekerite) at Eikeren-Skrim

A large outcrop of alkali granite with aegirine and/or alkali amphibole (ekerite) is found in the Eikeren-Skrim area southwest of Drammen (Fig. 2). More recent petrographical accounts on the ekerite have been presented by Dietrich et al. (1965) and Raade (1973).

The ekerite pluton is marked by an elongate, negative residual anomaly with

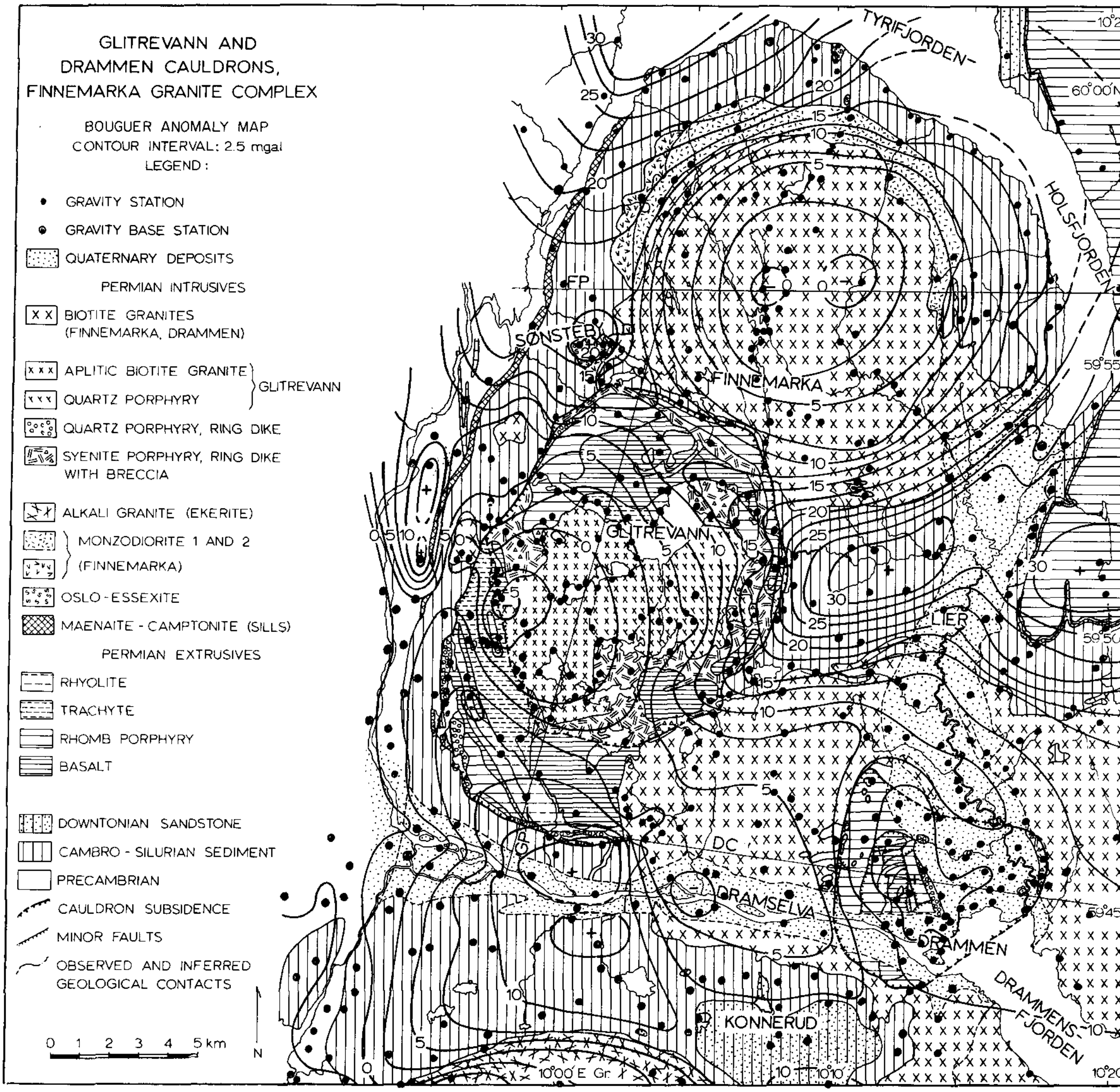


Fig. 33. Detailed Bouguer anomaly map of the areas around the Finnemarka granite complex, Glitrevann, and Drammen cauldron subsidences.

a minimum of about –21 mgal (Plate 5). Isogal contours indicate that felsic rocks occur at shallow levels also outside the main body below a cover of larvikite and effusives to the south-east; additional measurements (from 1972, see Fig. 34) confirm this conclusion. Similarly, the very irregular outcrop pattern to the south is apparently underlain by a much more regular mass distribution. In the southernmost part of the pluton, around Myklevann within the Skrim complex, the ekerite is seen to be intrusive into the larvikite/kjelsåsite, forming many dikes and breccias. From the residual map it is clear, however, that only rather restricted amounts of felsic rocks exist below the surface. The field impression of observing the very top of an extensive ekerite central plu-

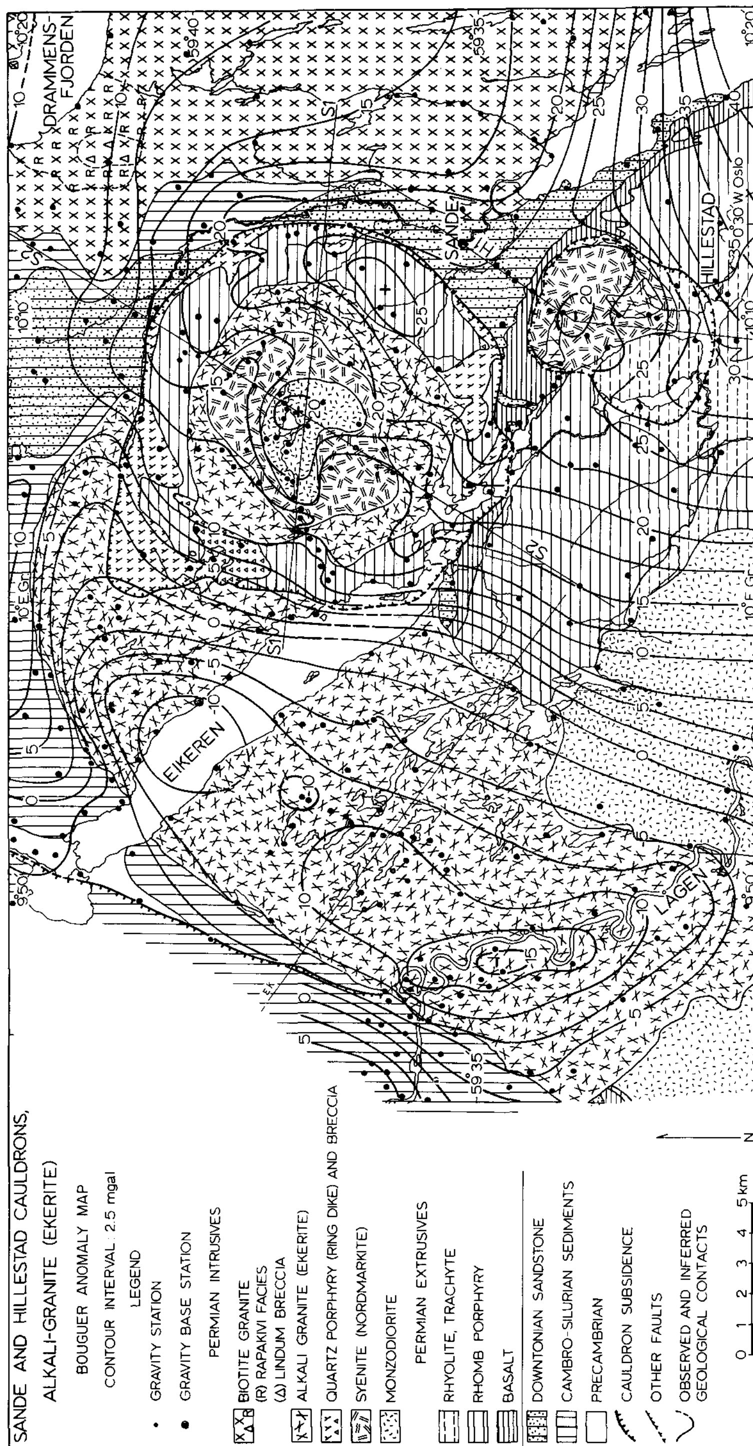


Fig. 34. Detailed Bouguer anomaly map of the areas around the Sande and Hillestad cauldron subsidences, and the Eikern alkali granite (ekerite).

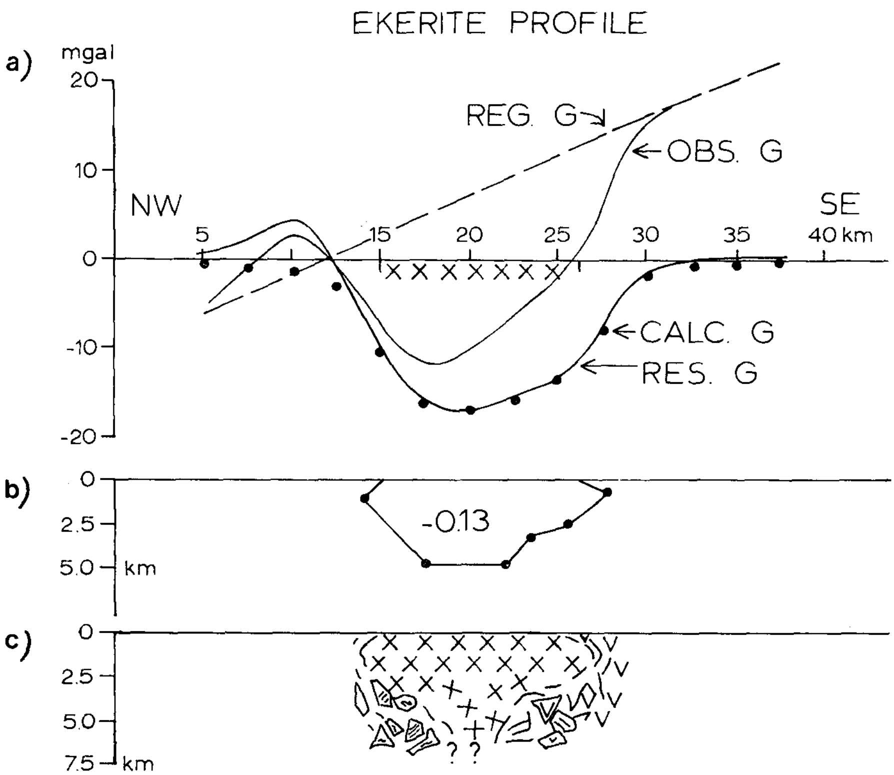


Fig. 35. Ekerite profile, gravity, and geological interpretation. Regional anomaly from Plate 4. Position of profile, see Fig. 34.

ton is not supported for this part of the region. The large masses of ekerite typically occur in the Eikern body; the ring complex around Myklevann, on the other hand, is composed largely of monzonitic and more mafic rocks.

The maximum gravity anomaly occurs at the intersection of the ekerite occurrence with the river Lågen. In Fig. 34, it is seen that the negative anomaly is caused by the interaction of two anomaly trends: 1) in the NNE direction, following the axis of the elongate ekerite body and associated with the mass deficiency represented by the felsic rocks, and 2) in the SSE direction and probably caused by the Quaternary river deposits. By considering only those gravity stations that were located on solid rock, the negative effect of the Quaternary deposits has been estimated to a maximum of 2–3 mgal, equivalent to a sedimentary thickness of about 150 m ($\Delta\rho = -0.5$ g/cm^3). This is in accordance with detailed gravity studies in the river valley (Grønlie & Jørgensen, pers. comm. 1973). Hence, the maximum gravity anomaly caused by the ekerite is assumed to be about −18.5 mgal above the Lågen valley.

A NW–SE gravity profile across the ekerite north of the Lågen valley is shown in Fig. 35 with a simple two-dimensional interpretation. Two three-dimensional models for the entire ekerite body are presented in Fig. 36. The models reveal essentially the same information as obtained for most of the other felsic bodies, that the light rock masses seem to form wedge- or cone-

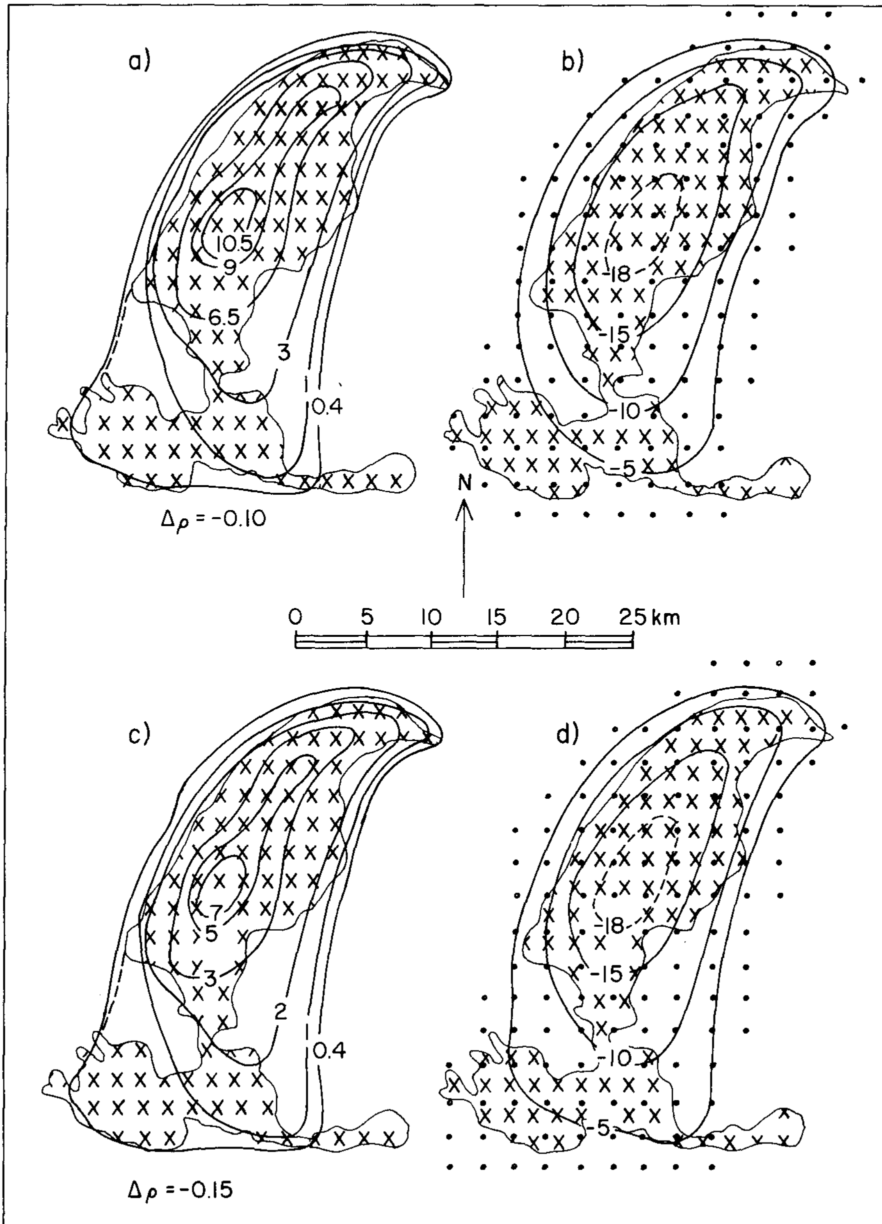


Fig. 36. Three-dimensional models of the Eikern-Skrim alkali granite (ekerite). (a) and (c): Geophysical models with density contrasts –0.10 and –0.15 g/cm^3, respectively. Contour line I = 0 km, identical to surface outline except where it follows dashed line. Extreme SE appendix omitted. (b) and (d): Calculated gravity anomalies. Calculations conducted with models subdivided into 22 lamellae. Black dots indicate positions of calculated gravity values.

shaped bodies with possible feeders below their central parts. The felsic rocks may be bordered and floored by Precambrian gneisses, stoped blocks and/or syenitic to monzonitic intrusives.

In conclusion, the ekerite occurrence SW of Drammen apparently forms an elongate body with a wedge-shaped cross-section. The effect of possible variations in the overall density contrast is shown in Fig. 36; with a density contrast of -0.10 g/cm^3 the thickness will be about 10.5 km; with a contrast of -0.15 g/cm^3, the thickness will be about 7 km. For the same reasons as previously stated, the preferred geological interpretation is that the felsic rocks grade into a substratum of stoped blocks and (more mafic?) intrusives.

On the basis of Fig. 34, excluding the negative gravity effect caused by the Lågen valley deposits, and applying the observed density contrast of -0.13 g/cm^3, the anomalous mass, total mass and volume of the ekerite has been estimated to $\Delta M = 1.79 \cdot 10^{17}$ g, $M = 35.9 \cdot 10^{17}$ g, and $V = 13.76 \cdot 10^{17}$ cm^3, respectively.

e. *Other felsic intrusives*

Additional occurrences of felsic intrusives are found within the following regions: 1) within some of the cauldron subsidences; these will be dealt with separately in a later section, 2) the Siljan area in between the monzonitic masses of the Skrim and Larvik batholiths, 3) the Hvittingfoss area (at about 59° 28′ N/45′ W of Oslo), and 4) the Nordagutu area (59° 26′ N/1° 26′ W of Oslo). The two latter occurrences, the Hvittingfoss syenite and the Nordagutu granite do not exhibit any notable gravity anomalies. It can be concluded that neither of the two occurrences is underlain by any large masses of felsic rocks of the types seen at the surface.

The residual map (Plate 5) indicates that the Siljan syenitic rocks are also associated with only a minor mass deficiency. In the western part of the area the gravity low is superimposed by the gravity high above the Skien basalts which, therefore, have to be considered along with the interpretation of the Siljan syenites. A two-dimensional interpretation of a SW-NE profile (largely parallel to the main road Skien-Siljan) across the Skien basalt and Siljan syenites (Fig. 58) shows that the latter form a rather shallow body floored by rocks of intermediate composition, supposedly larvikite. Since, however, transitional rocks frequently occur within the Siljan massif as seen, e.g., from the density distribution map of Fig. 16, it might be more appropriate to use a lower density contrast. Hence, along the presented profile the maximum depth of the 'syenitic' body is at least 2.5 km, probably more. It is not unreasonable to assume that the massif grades into monzonitic rocks with depth or contains inclusions of subsided basalt, or both. If we exclude the positive gravity effect of the Skien basalt and assume an overall density contrast of -0.11 g/cm^3 between the syenites and the gneisses, the mass and volume of the Siljan massif can then be estimated (see Table 9).

Table 9: Anomalous mass, total mass, and volume estimates of the Oslo Region felsic rocks

Rock complex	Anomalous mass ΔM (grams)	Total Mass M (grams)	Volume V (cm^3)
Nordmarka-Hurdalen	11.77 x 10^{17}g	257.0 x 10^{17}g	98.09 x 10^{17} cm^3
Drammen granite	2.35 »	47.3 »	18.11 »
Finnemarka granite	0.44 »	8.8 »	3.36 »
Eikern-Skrim granite	1.79 »	35.9 »	13.76 »
Siljan syenite	0.26 »	6.7 »	2.56 »
Cauldron plutons (see Table 12)	0.61 »	12.1 »	4.57 »
Sum felsic rocks	17.22 x 10^{17}g	367.8 x 10^{17}g	140.45 x 10^{17} cm^3

1 km^3 = 10^{15} cm^3

f. Concluding remarks

The large areal extent of the Oslo Region felsic rocks is to some extent matched by the volume as interpreted from the residual anomalies. If one includes the masses of felsic rocks assumed to occur within the cauldron subsidences, then the estimated mass and volume of the Oslo Region syenitic and granitic intrusives may be read from Table 9.

Even though great care has been taken to compensate for interferences with gravity highs (e.g. the Kongsberg high just east of the ekerite low) and to account for badly defined gravity flank anomalies, the above mass and volume estimates have to be considered as rough approximations. Mass calculations by other methods (e.g., from the computed 3D models) may in some cases differ by up to 15 %. The applied method (outlined on p. 60) gives consistently low values. This is probably due to the fact that occurrences of included dense rocks will as a result give mass estimates of the *difference* between the light and dense rocks; and masses of denser rocks are likely to occur within several of the complexes tabulated in Table 9. By always applying the same method of mass calculation, a relative consistency is obtained.

There are some significant differences in the apparent structures (and masses) of the various felsic batholiths. The huge Nordmarka-Hurdalen batholith and the smaller ekerite batholith reach down to considerable depths (8–12 km or more); uniform density models imply wedge-shaped or root-like extensions in their central parts. Similarly, the Finnemarka granites form a deep inverted cone, while other granites (e.g., Drammen) apparently form tabular batholiths with possible feeders. The apparent structural divergences are accompanied by differences in their overall petrographic composition and geochemistry. If this simple structural interpretation is valid, this might indicate that the various felsic rocks derived from different sources and at various depth levels following somewhat divergent trends of evolution. The wedge- to mushroom-shaped form for all the major felsic intrusives seems a reasonable solution for these originally low-density, buoyant and fluid bodies.

Gravity models only estimate the depth of the assumed density contrast;

they do not resolve what rock-types underlie the felsic rocks at the lower level of the density contrast. It follows, therefore, that different possibilities exist, and the final decision has to rely upon other geophysical methods and geological deductions. There is ample evidence that the Oslo intrusives ascended by means of stoping, and that the actual crustal separation is rather small. Evidence of large-scale doming is not observed. To solve the room problem it is therefore necessary that the felsic intrusives more or less traded place with blocks of Precambrian gneisses, monzonites, etc. Alternatively, therefore, the felsic rocks are not floored by solid gneisses but may represent the upper portion of large, inhomogeneous batholiths with vertical density stratification according to changes in composition of intrusives, increased amount of stoped blocks and possible accumulates. An interesting aspect is that if the displaced mass has descended through the emplaced syenites and granites, the subjacent layer must have been molten, too. Thus energy requirements will be even higher than those for the 'simple' model of intrusive felsic rocks.

6.3 THE MAFIC INTRUSIVES

Mafic intrusives occur as dikes and sills, and in the volcanic necks as 'Oslo-essexites' (Barth 1945). The volcanic necks or pipes are distributed along possible fracture zones (Fig. 5). From geological evidence they clearly predate the main plutonic series (kjelsåsite-granite).

The 'Oslo-essexites' (Table 2) are regarded not as a direct part of the main series, but to form an independent differentiation series (Fig. 6a). Previous authors seem to agree in a deep mantle-derived origin for the 'Oslo-essexites' (Barth 1945, 1954, Oftedahl 1952, 1959, 1967), but part of the differentiation may well have taken place in separate chambers closer to the surface. Recent geochemical studies imply that also the differentiated kjelsåsite-granite series originate from the deeper part of the crust or the upper mantle (Heier & Compston 1969, Finstad 1972). Hence, the diversity between the 'Oslo-essexites' and the main series might be due to different paths of development rather than contrasted source rocks.

The volcanic necks are generally supposed to extend vertically as subcircular feeders. Gravity studies may help confirm their existence or disclose structural relations in general. Specifically, closely spaced measurements have been carried out in the Hadeland area to see if there is any indication of *larger,* subsurface, dike-like connections between the individual necks, but this was not found (Fig. 39a).

The Vestby model. Detailed measurements have been carried out in the Vestby region which offers favorable terrain conditions. The Vestby twin volcanic neck (Fig. 37) reveals a concentric isogal pattern with a maximum anomaly of about 7.5 mgal above the circular Brattåsen neck and about 6 mgal above the oval Knalstad neck. The Bouguer map apparently indicates a slight tilt of the

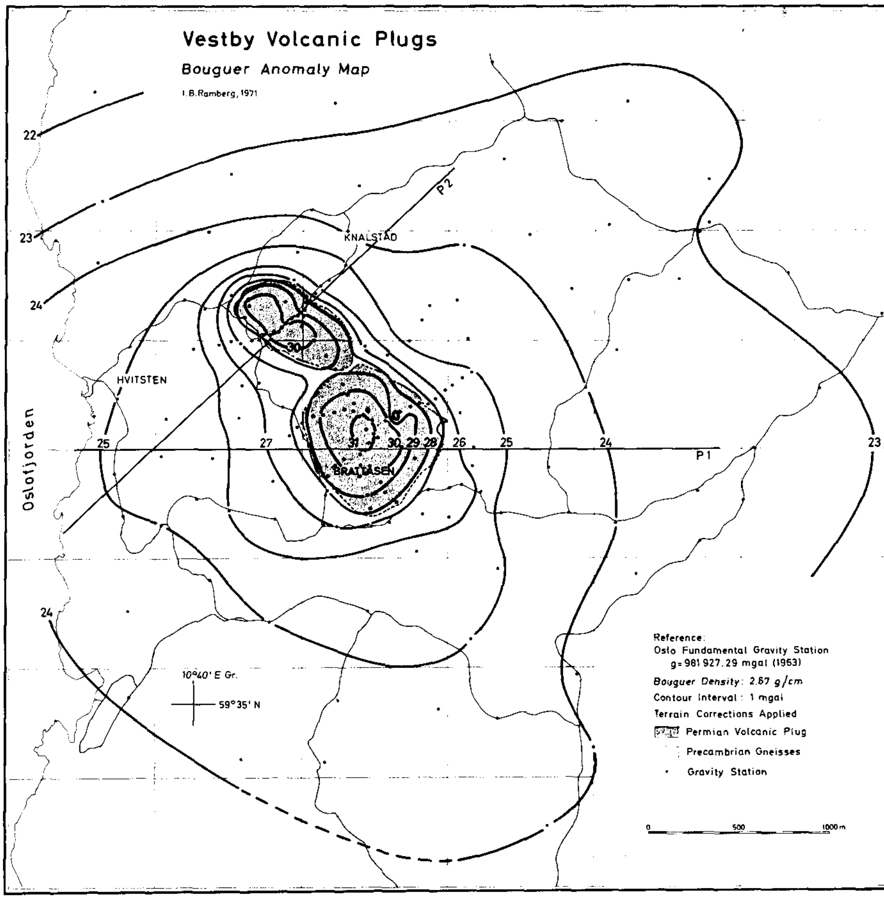


Fig. 37. Bouguer anomaly map of the Vestby volcanic necks.

pipes toward east; this effect is, however, partly reduced when compensating for the regional trend in the area (Plate 4).

Three-dimensional models have been calculated for the Brattåsen and Knalstad necks; Fig. 38 shows results plotted along E–W and NE–SW profiles, respectively. The best fit between observed and calculated gravity anomalies was obtained for barrel-shaped models with a thickness of only about 1.5 km. Because of the dense monzodioritic to pyroxenitic rocks prevailing in the Vestby necks (Table 6) and the somewhat lighter than average surrounding gneisses (Table 7), the high density contrast of $+0.5$ g/cm^3 (3.20–2.71) had to be used. Even so, the peak values differ somewhat in the Knalstad anomaly; a deeper model would preferentially widen the calculated anomaly on the flanks while a somewhat higher density contrast for the uppermost part of the body would greatly improve the fit. Therefore, the commonly occurring, extremely dense, pyroxene-magnetite rocks (e.g. in old iron ore quarries) are probably enriched in a subsurface 'layer' in the Knalstad neck.

The important geological implication that can be derived from the models

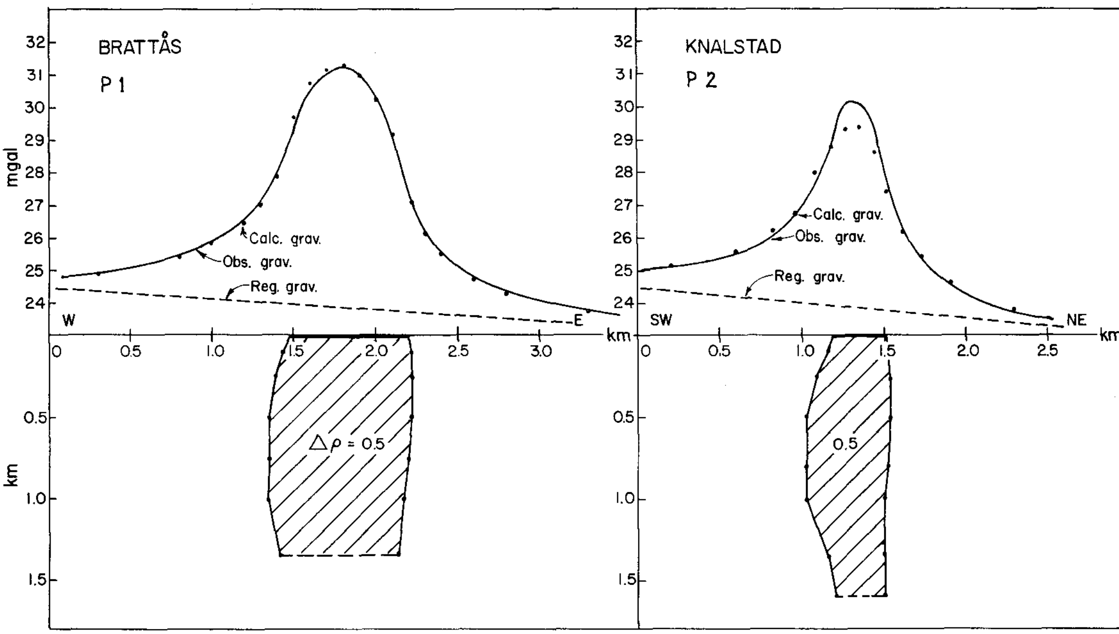


Fig. 38. Three-dimensional models of the Vestby volcanic necks viewed along an E–W (Brattåsen neck) and a NE–SW (Knalstad neck) profile. Regional anomaly from Plate 4; positions of P1 and P2 seen in Fig. 37.

depicted in Fig. 38 is that the Vestby necks do not seem to extend downward as pipes of uniform width. Below a depth of about 1.5 km the feeders have to represent quite negligible volumes and masses, or else they would show up differently in the observed anomalies.

Gravity profiles. The resolution of the Vestby study is estimated not better than ± .03 km 'pipe length,' and underlying feeders with diameters of less than 0.3–0.4 km would hardly be identified. Also, the general value of the Vestby results needs to be checked since the Vestby occurrence contains rocks of more extreme ultramafic composition than most of the other occurrences. Fig. 39 shows gravity profiles through eight of the other major occurrences. The roughly E–W profiles through the four largest of the Hadeland necks (Fig. 39a) are taken from a detailed Bouguer map constructed from that area. After removal of the 'regional' (e.g., the combined effect of the downward dip of the Moho toward the Caledonides to the west and the effect of the large syenite batholith to the east), the four occurrences all exhibit residual gravity anomalies in the range 1.5–2.6 mgal.

A similar anomaly (2.8 mgal) is found associated with the largest of the two subcircular Huseby necks (Fig. 39b), while a much smaller value (0.65 mgal) is revealed above the largest of the three Ullern-Husebyåsen necks. This profile shows that virtually no anomaly exists above the southern part of the neck which is occupied by akeritic rocks with hardly any density contrast with respect to country rock. The small positive anomaly to the north of the neck

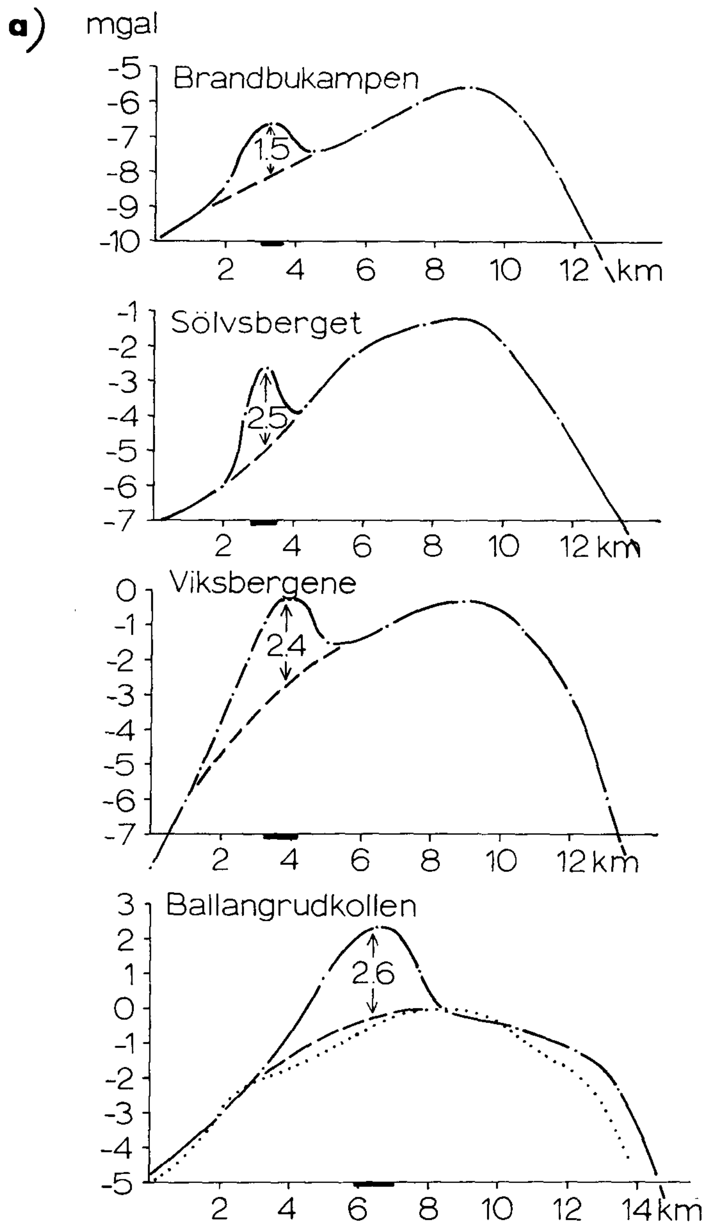


Fig. 39a. E–W Gravity profiles across the four major volcanic necks at Hadeland. Observed gravity: dot-dash lines; Assumed 'regional': dashed lines. Observed profile between the Viksbergene and Ballangrudkollen occurrences is drawn on the lowermost figure and shows only the 'regional' effect.

is caused by interference from body no. 2 situated just to the NE of the investigated body.

Larger positive anomalies of about 5.1 and 7.5 mgal mark the Sønstebyflakene and Tofteholmene occurrences, respectively. The former occurrence is a large and complex body of predominantly modumite (Barth 1945) which

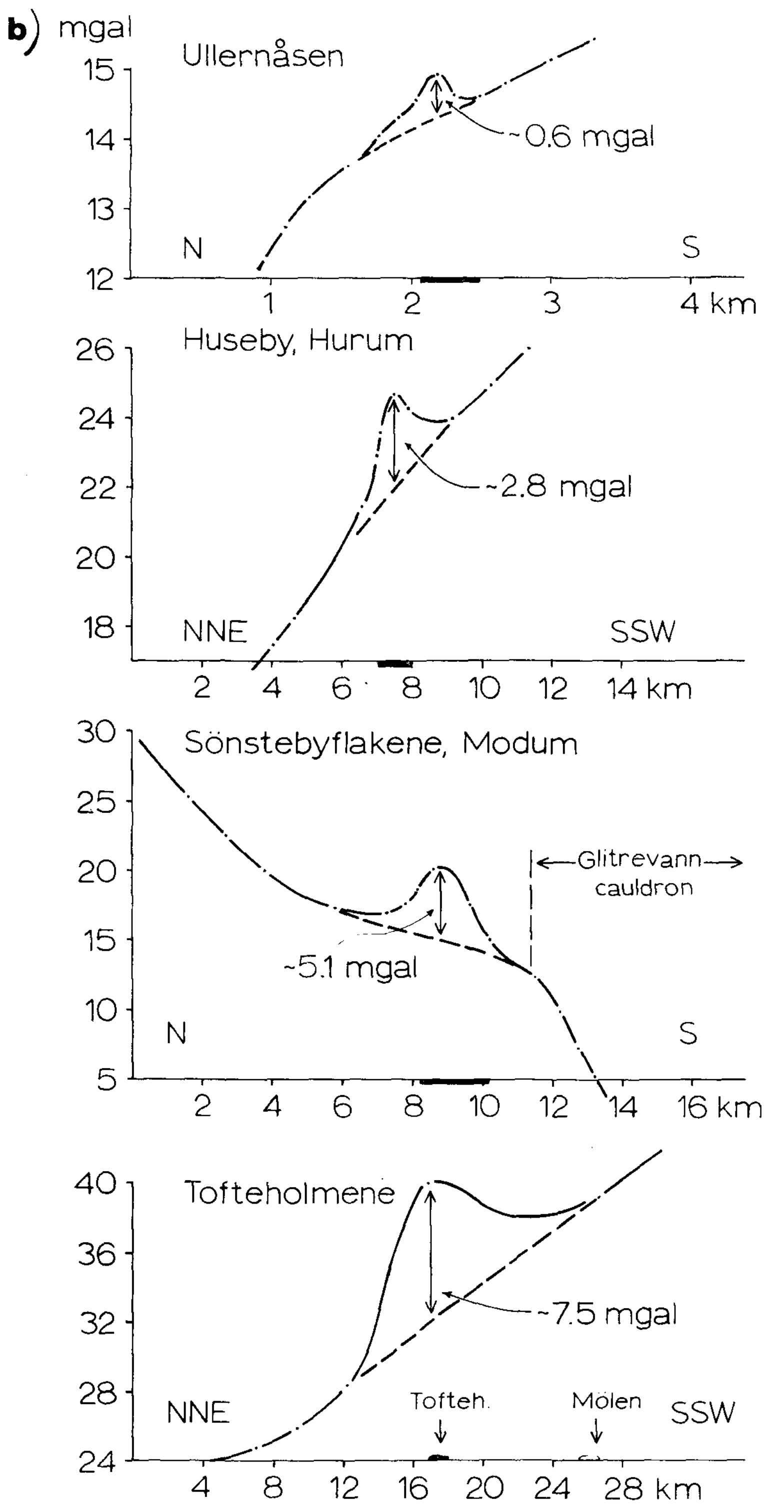


Fig. 39b. Gravity profiles across the Ullernåsen, Husebyåsen (Hurum), Sønstebyflakene and Tofteholmene volcanic necks.

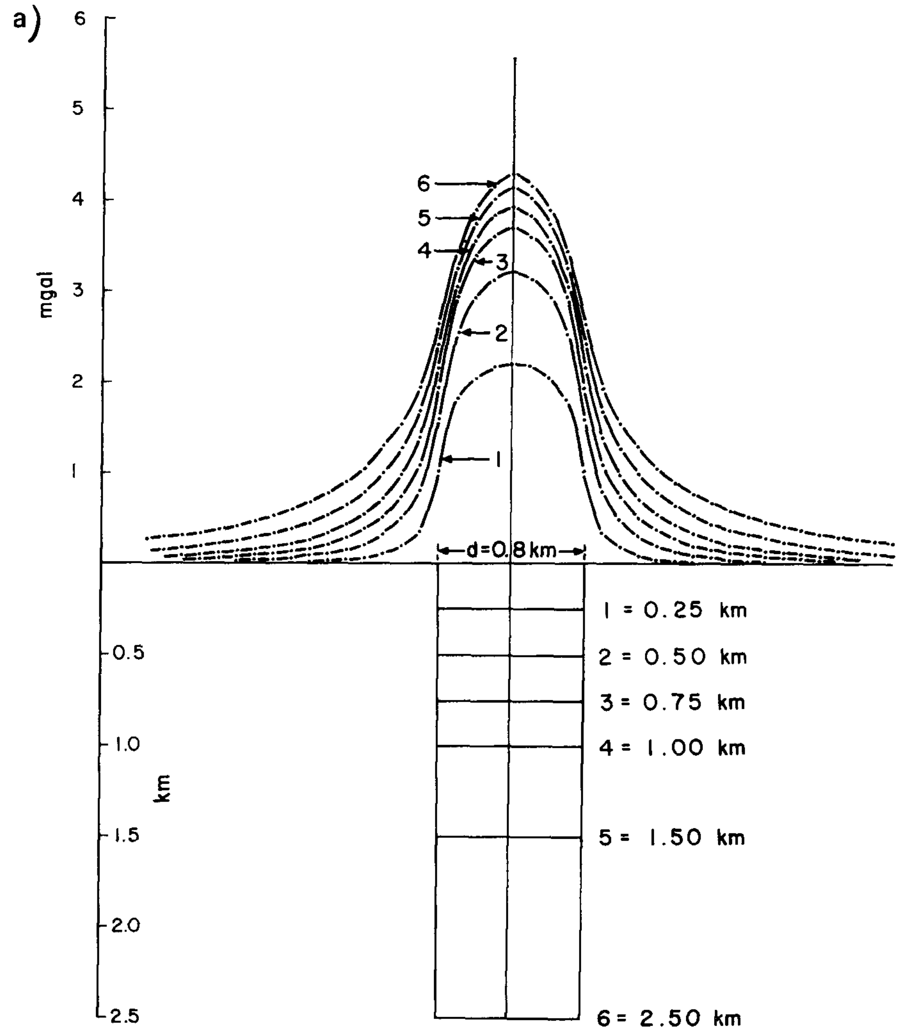


Fig. 40. Theoretical models of volcanic pipes varying the diameter (d) and thickness (z), compared to the observed peak values of the 'Oslo-essexites'. Density contrast kept constant, $\Delta\varrho = + 0.3$ g/cm^3. (b) and (c) shown on p. 90 and p. 91, respectively.

occurs associated with a body of eucrite-gabbro and transitional rock-types. The eucrite-gabbro was originally termed 'basic akerite' on the Brøgger & Schetelig 1 : 100 000 Kristiania and Hønefoss maps, and it is not yet clear whether or not this gabbroic rock is part of the outer marginal ring of the Finnemarka complex. The Tofteholmene occurrence is found as scattered intrusives on a number of islands and skerries in the Oslofjord ('Der Grosse Hurumvulkan', Brøgger 1931a). Its exact outline is unknown. As in Vestby, the rocks range from ultramafic to monzonitic in composition. The residual anomaly is highly uncertain since the central and southern part of the profile is based on just a few measurements on the islands. It may imply, however, that in harmony with the petrographical similarities between the two occurrences, the Tofteholmene high is of the same order of magnitude as the Vestby high.

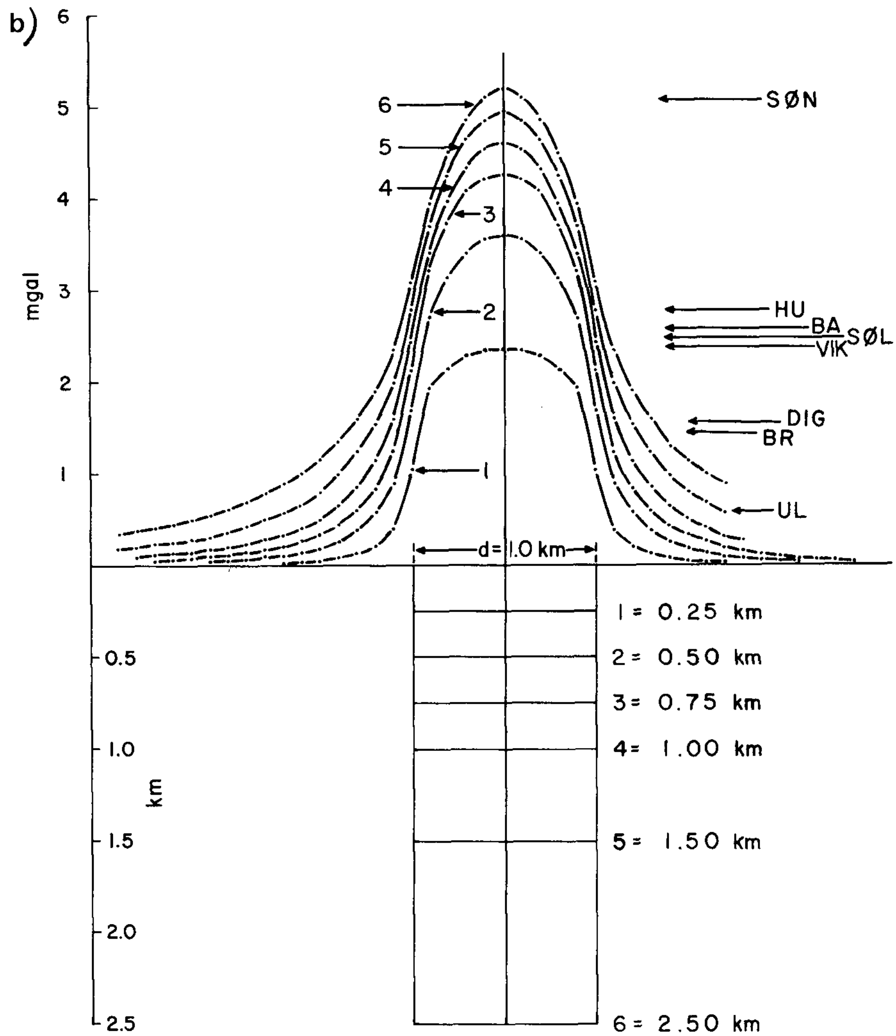


Fig. 40b. Explanation on p. 89. Abbreviation (SØN, HU, etc.), see Fig. 5.

Gravity measurements from the small, elongate neck at Tingelstad (Fig. 5) give only a very small or negligible anomaly, that is, less than 0.2 mgal. The same applies to the Snaukollen occurrences within the Glitrevann cauldron subsidence.

Interpretation. Most of the gravity anomalies connected with the volcanic necks seem surprisingly small and they certainly represent but minor anomalous masses. Detailed structural interpretations of every one of the many occurrences are not presented, but some general information is obtained by comparison of the observed profiles with theoretical curves calculated from simple three-dimensional models (Figs. 40 & 41).

Except for the Vestby and Ullernåsen necks, and perhaps also the Tofteholmene occurrence, the 'Oslo-essexites' exhibit mean density contrasts of

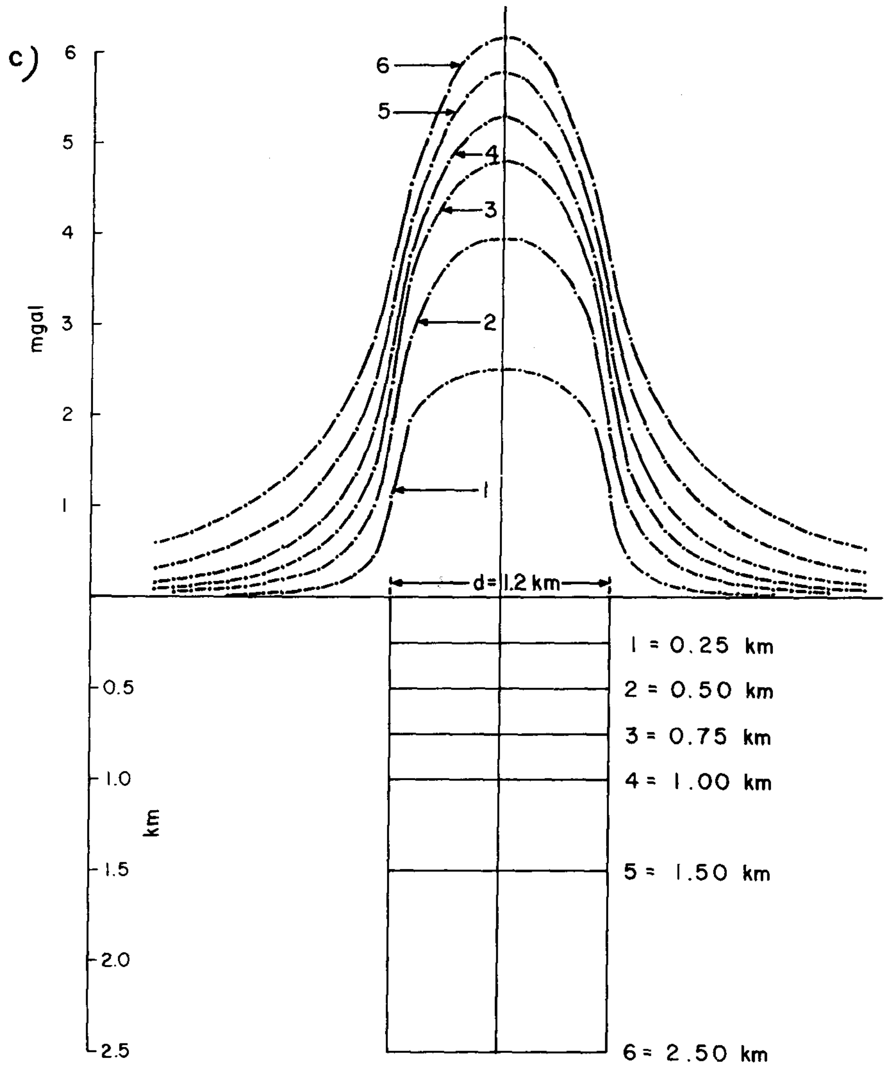


Fig. 40c. Explanation on p. 89.

about 0.2–0.3 g/cm³ with respect to the Precambrian gneisses. If we consider the data in Table 10 along with the families of curves in Figs. 40 & 41 and assume simple cylindrical downward extensions of the various necks, the four largest Hadeland necks, and the Ullernåsen, the Dignes and the Huseby necks all have plausible thicknesses of about 0.3–0.6 km. The Sønstebyflakene occurrence may have a thickness of about 0.8 km. In all cases *the thicknesses of the bodies appear to be less than the average diameters.* Since the Bouguer anomalies were calculated with reference to sea-level, and as only a constant Bouguer density of 2.67 g/cm³ has been applied, the gravity anomaly of the mass surplus between surface and sea-level has been included in the residuals. The estimated thicknesses of the necks should therefore be measured downward from the present-day surface (and not from sea-level). The possibility that the

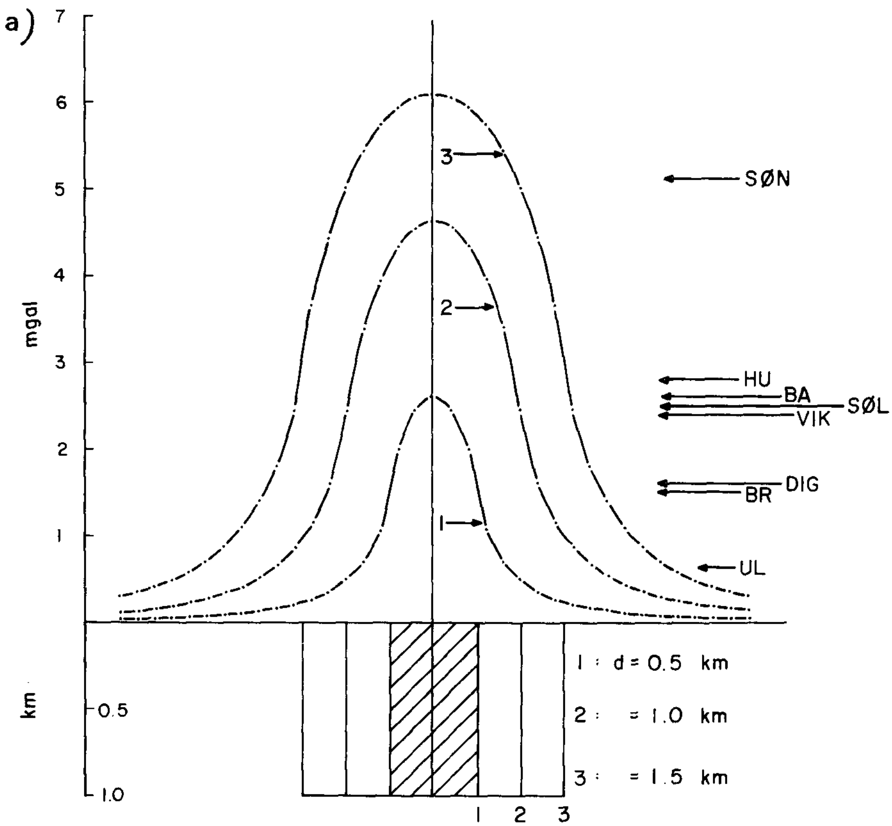


Fig. 41. Theoretical models of volcanic pipes, (a) varying the diameter, d, and keeping the density contrast ($\Delta\varrho = 0.3$ g/cm³) and thickness ($z = 1.0$ km) constant.

Table 10: Estimates* of long and short axes of 'Oslo-essexite' outcrops, assumed density contrasts, and residual gravity anomalies.

Neck	Approx. outcrop	Assumed $\Delta\rho$	Residual g
Brandbukampen	0.35 x 0.9 km	0.3 g/cm³	1.5 mgal
Tingelstad	0.1 x 0.5 »	0.25 »	0.15 »
Sølvsberget	0.7 x 0.9 »	0.25 »	2.5 »
Viksbergene	0.8 x 0.9 »	0.25 »	2.4 »
Ballangrudkollen	0.8 x 1.25 »	0.3 »	2.6 »
Dignes	0.3 x 0.6 »	0.25 »	1.6 »
Sønstebyflakene (incl. gabbroic body)	1.3 x 1.8 »	0.25 »	5.1 »
Ullernåsen I	0.3 x 0.7 »	0.15 »	0.65 »
Huseby I	0.7 x 0.8 »	0.3 »	2.8 »
Vestby I	0.7 x 0.8 »	0.5 »	7.5 »
Vestby II	0.35 x 0.8 »	0.5 »	6.0 »
Tofteholmene		0.4? »	7.5? »

* Some necks have irregular or boomerang-shaped outline (e.g Sølvsberget, Ballangrudkollen) and only rough estimates of the axes are possible.

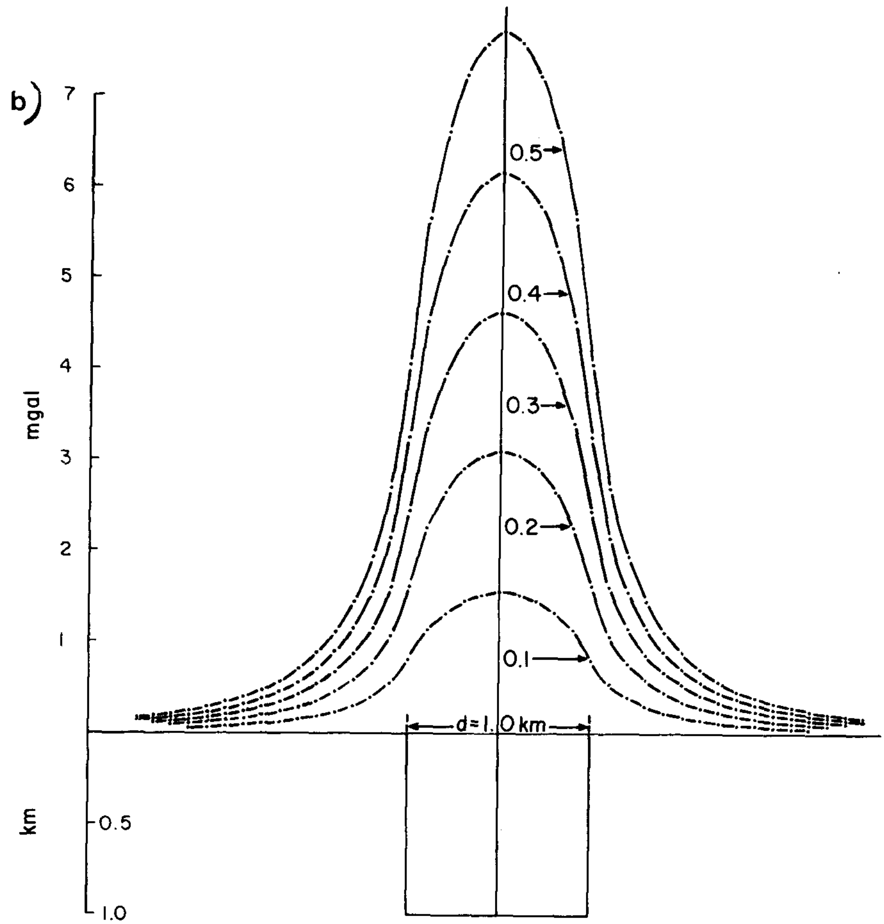


Fig. 41b. Varying the density contrast, $\Delta\rho$, and keeping the diameter (d = 1.0 km) and thickness (z = 1.0 km) constant.

'Oslo-essexites' have only a very restricted vertical extent seems to accord with some field evidence as exemplified by Brøgger's (1933b) geological section across the Brandbukampen (Brandberget) neck (Fig. 42). Other field observations indicate contacts which are close to vertical or even outward dipping (cfr. Vestby) contacts, implying that the necks are but thin barrel-shaped or pear-shaped bodies.

Discussion and conclusion. Though not all of the 'Oslo-essexites' have been studied, the remarkable consistency of the gravity data obtained lends credence to the conclusion that the main mass of the bodies is concentrated close to the present surface. Below a depth of about 0.5 km, sometimes 1.0 or 1.5 km, no traceable density contrast exists.

The most reasonable geological interpretation seems to be that the 'Oslo-essexites' represent subvolcanic magma chambers. Alternatively, but less likely, they may form the upper, expanded portions of volcanic pipes (e.g., like the Kimberley pipe, Holmes 1965). In any case, below the shallow depths

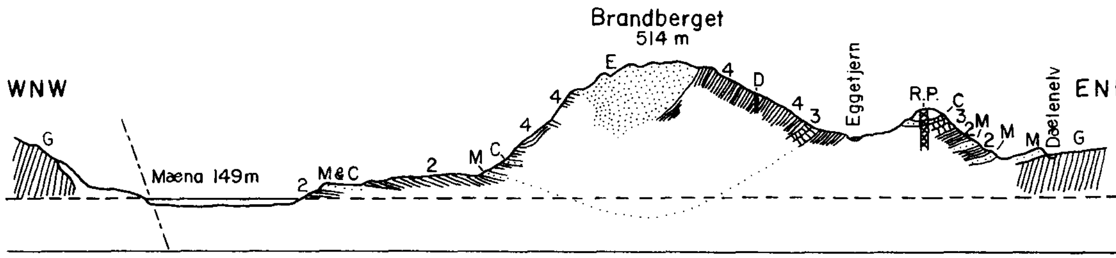


Fig. 42. Geological section across the Brandbukampen (earlier: Brandberget) occurrence (*after* Brøgger 1933b).

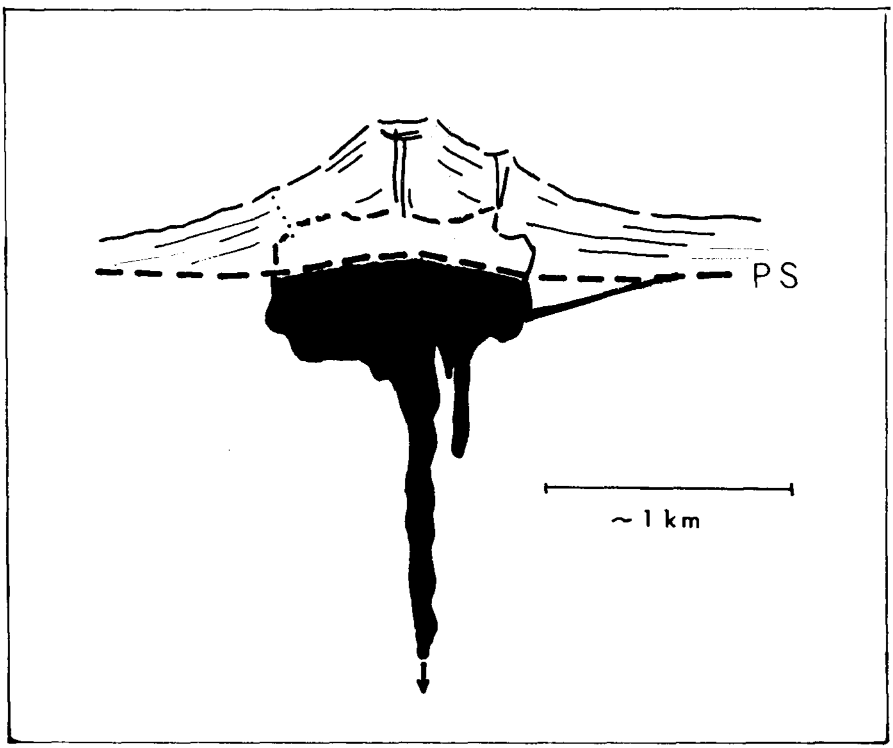


Fig. 43. Generalized cross-section of the Oslo Region volcanic necks ('Oslo-essexites') based on gravity interpretation. PS = present-day surface.

indicated by the gravity anomalies, only thin dikes or feeders may extend downwards along fracture zones (Fig. 43). From the mafic intrusions of the Monteregion Province, Canada, similarly small gravity anomalies are encountered over intrusions comparable in size to the 'Oslo-essexites' e.g., Mt. Johnson (Kumarapeli et al. 1968), again indicating floored magma chambers.

The possible magma chambers seem to be surprisingly small. Very little is known about the actual shape and size of such reservoirs in general (Macdonald 1972), but from the active Avachinsky volcano in Kamchatka, gravity and magnetic studies (Steinberg & Rivosh 1965) have localized an ellipsoid magma chamber with diameters of about 1.5 by 4 km at 2.5 to 8 km depth below the base of the volcano.

Studies of earthquakes and volcanic tremor have led to the conclusion that the Sakurajima volcano in Japan is underlain by a magma reservoir at the depth of about 3.5 km (Sugimoto & Namba 1958). Tiltmeter studies in the area of Kilauea, Hawaii, have indicated a reservoir at about 3.5–4.5 km below the volcano, while an analysis of data from different volcanic regions of the earth implies that extinguished and active volcanoes generally rest on top of magma chambers at 2 to 7 km depth (Eaton & Murata 1960, Macdonald 1961). Thus, the phenomenon that so many of the 'Oslo-essexites' are cut by the present-day surface in the middle of their expanded, pear-shaped bodies, may be due to the fact that they represent a specific depth level relative to the Permian surface, the present erosional surface being up to 3 km below the Permian one (McCulloh 1952). The postulated, shallow magma chambers evidently cannot mark the region of magma generation; they may therefore represent storage chambers fed by basaltic magma through narrow vents or feeders from considerably greater depths. Minor dike-like occurrences such as Tingelstad, Gåsøya etc. may represent such underlying feeders and constitute the true volcanic pipes of the Oslo Region.

In the Oslo Province, Brøgger (1894a) thought that the closely associated camptonite and maenaite dikes were comagmatic, originating by differentiation of an original basaltic magma. Brøgger assumed that this differentiation took place within the dike and sill intrusions themselves, as seems to be the case, e.g., in the huge Jarenvann sill east of Sølvsberget. Here, in new road-cuts, the camptonite in places grades continuously upward into maenaite, implying a separation by gravitative differentiation. Commonly, however, the maenaite and camptonite dikes are more separated in space, and Oftedahl (1957c) suggested that the differentiation took place within subcrustal magma chambers close to the Precambrian peneplain. The existence of such magma chambers seems supported by the present gravity study. Also, a detailed gravity study of the Jarenvann camptonite (Lehne 1972) shows that it is a plate-like intrusion, or sill, dipping gently towards the south-west and probably extending along the Precambrian peneplain towards the row of 'Oslo-essexites' to the west.

While many plugs, necks and volcanic pipes are largely or wholly composed of fragmental material, the 'Oslo-essexites' contain rather homogeneous, coarse-grained and virtually non-porphyritic rocks forming a differentiated series of their own (Fig. 6a). Zonal arrangement of the petrographic types and gradational contacts are commonly observed, e.g., in the Vestby and Huseby occurrences. Rhythmic layering shows that many of the basic rocks are cumulates and indicates that the intrusives solidified by depositing successive layers on the walls of the magma chambers. All evidence taken together, the conclusion would appear to be that the 'Oslo-essexites' were formed in some sort of subsurface storage chambers.

From the detailed gravity survey of the Vestby twin necks and the more scattered gravity readings and other relevant data referring to the other necks (Table 10), the total mass and volume involved in the 'Oslo-essexite' intrusions

Table 11. Anomalous mass, total mass, and volume of the 'Oslo-essexites' chiefly based on the analogy with the Vestby necks and data in Table 10

Rock complex	Anomalous Mass, ΔM	Total Mass, M	Volume, V
Vestby, I and II	0.31×10^{15}g	2×10^{15}g	0.63×10^{15} cm^3
All 'Oslo-essexites'	$\sim 1.50 \times 10^{15}$g	$\sim 12 \times 10^{15}$g	$\sim 4.7 \times 10^{15}$ cm^3

have been roughly estimated (Table 11). Except for the Vestby necks, an average density contrast of 0.3 g/cm^3 has been employed for the computations.

6.4 INTRUSIVES OF INTERMEDIATE COMPOSITION

Rocks of intermediate composition are found as: 1) monzonitic plutons in the southern Oslo Region, e.g., the Larvik and Skrim batholiths, 2) syenodiorites within some of the cauldron subsidences, and 3) syenodiorites within the Nordmarka – Hurdalen batholith. Altogether they occupy about 2000 km^2 of the surface area.

Gravity measurements are in general unable to unveil much information about the syenodioritic rocks because their average density is not significantly different from that of the surrounding gneisses. Nevertheless, some ideas about the total volume of syenodiorite involved in the Oslo Province are important when considering the tectonomagmatic development of the region. Some points of relevance to this matter are summarized below:

a) A large proportion (approx. 60%) of the Oslo volcanic sequence is of intermediate composition (rhomb-porphyries, etc.). The alternate occurrence of basaltic and latitic extrusives suggests a tapping either from coexisting magma reservoirs of gabbroic and monzonitic composition, respectively, or from different levels of a single, stratified chamber. The chemical composition of the rhomb-porphyries seems remarkably constant with time and place (Brøgger 1933a) even though a tendency for a differentiation trend from intermediate to acidic composition has been shown (Oftedahl 1967).
b) The Skrim and especially the Larvik batholith exhibit large areas of relatively homogeneous monzonitic rocks, and it seems reasonable that these complexes extend downward to some depth. Oftedahl (1952) argues along these lines of evidence and arrives at an estimate of about 10,000 km^3 of monzonite.
c) Regional aeromagnetic surveying (Haines et al. 1970) flown at about 10,000 feet altitude gave anomalies of about 600–800γ over the large syenitic and monzonitic massifs. These anomalies are of a similar magnitude to those of the NGU data obtained from much lower altitudes (100–150 m), a fact which is interpreted (Åm, pers. comm. 1972) as being due

to a large vertical extent for both the syenitic and the monzonitic batholiths.

d) Aeromagnetic anomalies (Haines et al. 1970, Åm, pers. comm. 1972) not accompanied by any notable gravity anomalies might imply the existence of monzonitic rocks in the following areas: 1) the Baerum area south of the cauldron subsidence and north of the Asker gravity high; the Paleozoic sedimentary rocks show contact metamorphic phenomena similar to those close to outcropping batholiths (Goldschmidt 1911); 2) the Skagerrak area at about 8° 30′ E/58° N; 3) the Brumunddal area.

e) Possible subcircular syenodioritic batholiths in the Nordmarka area and the Hurdalen–Toten area seems partly removed by the subsequent syenitic intrusives by stoping. Individual 'inclusions' like the Gjerdingen-massif may be at least 1 km thick (Grønlie 1971, Kristofferson 1973). Larger amounts of syenodiorite and/or Precambrian gneisses are likely to occur along the flanks and below the Nordmarka–Hurdalen batholith and other felsic intrusions (Section 6.2).

f) Rocks of monzonitic composition would seem to be more common at depth than is revealed by surface observation in some of the cauldrons (Section 6.5). These smaller, central intrusions may be representative of the Oslo plutonic activity in general and indicate the existence of and possible transition into more syenodioritic rocks with depth.

g) The Siljan syenitic massif seems to grade with depth into rocks of intermediate composition.

h) The existence of major rhomb-porphyry dikes in a wide area outside the Oslo Region proper (Fig. 2) indicates the presence of a widespread monzonitic source magma in Permian time.

Conclusion. There seems to be enough evidence to suggest that the Oslo igneous province contains at least the same amount of intermediate rocks as felsic rocks, and possibly a good deal more. The ratio intermediate/felsic rocks seems to increase downward. One might speculate as to whether the entire Oslo Region is underlain by a syenodioritic substratum, or more likely by a complex mixture of fractured Precambrian rocks, syenodioritic intrusives, dikes and stoped blocks.

6.5 CAULDRON SUBSIDENCES*

Cauldron subsidences were formed throughout the Oslo Region, as revealed from Fig. 7. A thorough treatment of the complexes has been given by Oftedahl (1953), and later also by Oftedahl (1960, 1967) and Naterstad (1971). A gravity study of the Baerum cauldron suggests that the cauldron block of mostly effusive rocks subsided into a nordmarkite pluton with a possible thick-

* 'Cauldron subsidence' and 'cauldron' may be applied as synonymous expressions according to the Glossary of Geology, Amer. Geol. Inst. 1972, although the former term commonly relates to the *process* of formation, and the latter to the *structure*.

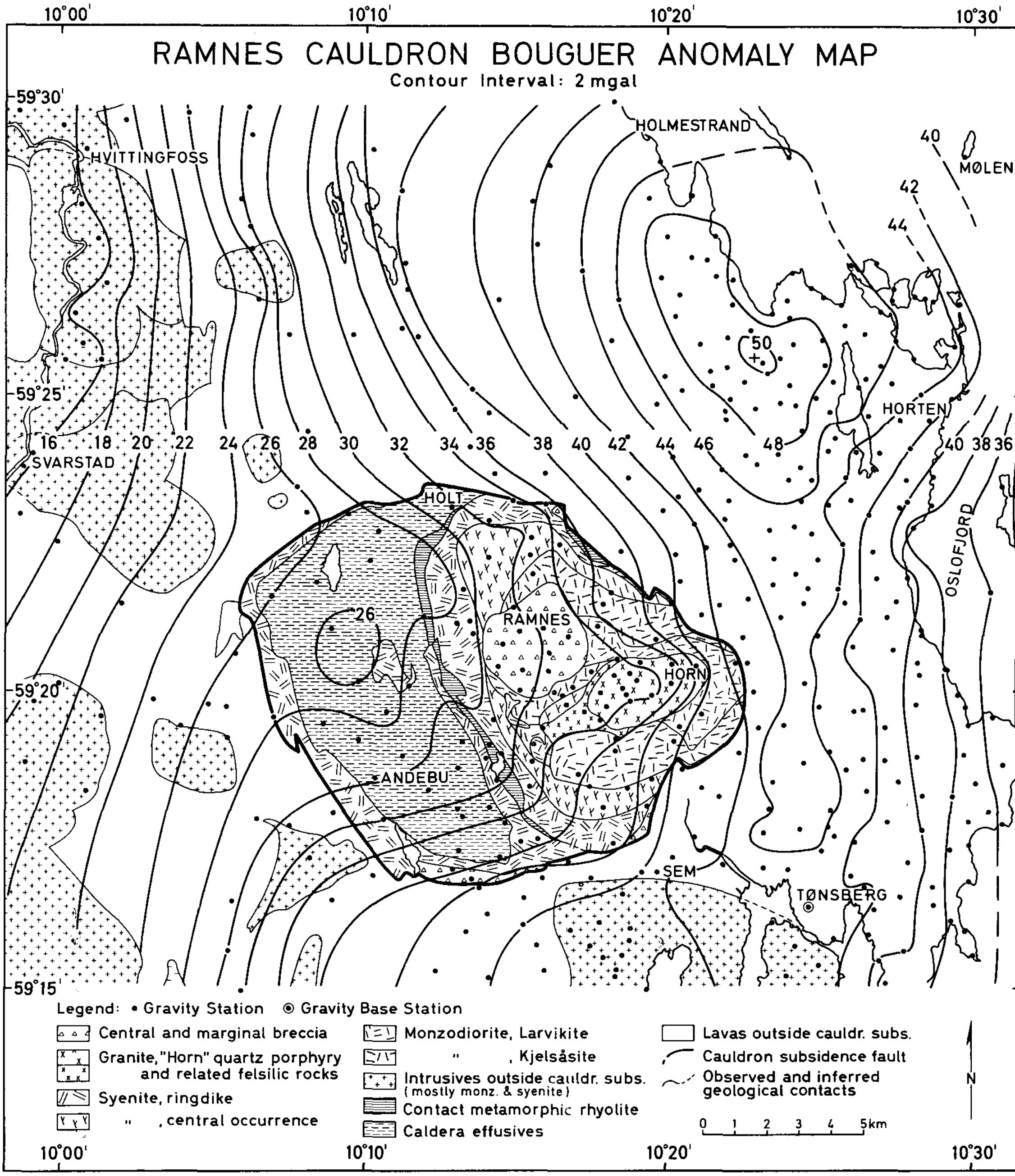


Fig. 44. Detailed Bouguer anomaly map of the Ramnes cauldron and surrounding area. Geology *after* R. Sørensen (pers. comm. 1972).

ness of about 9 km (Smithson 1961). A number of other 'ring structures' within the igneous complex may be genetically related to the cauldrons, but have not been dealt with as such in the present account.

Central intrusions seem almost ubiquitous in cauldrons, and from the Oslo

Region, Oftedahl (1952) has attributed the cauldron subsidence to withdrawal of magmatic support. The rather common diameter of the cauldrons and other ring complexes of about 10–15 km indicates that the circular fractures originated as a result of axial stresses at a certain depth level. If we assume the inward angle of the cone-shaped fracture systems (Durrance 1967) to be about 60–65°, the depth is about 16–18 km. This could represent the initial depth of the ascending magma. Gravity data may throw light on the present configurations of these rock masses.

a. Ramnes cauldron

The Ramnes cauldron is the southernmost of the presently described Oslo Region subsidences (Fig. 7). It consists of a western part of latite, trachyte, and mostly rhyolite flows (Oftedahl 1967) and an eastern part dominated by a monzonitic to syenitic central pluton (Sørensen, pers. comm. 1972); see Fig. 44. Oftedahl (1967) suggested that the central pluton represents a huge lava-dome with the cryptocrystalline porphyric varieties as true volcanics, while Rutten & Everdingen (1961) look upon the whole complex as nothing but volcanic rocks transformed by pneumatolitic activity. However, field occurrence and chemical variation suggest that the kjelsåsite-larvikite-syenite form a comagmatic intrusive series, while the granitic intrusion has a Th/U-ratio different from that of any other Oslo Region pluton (Raade 1973). Broad syenitic ring-dikes encircle both the eastern and western parts.

Two gravity profiles have been drawn on the anomaly map (Fig. 44), one roughly N–S (Fig. 45a) and one E–W (Fig. 45e). The overall impression is that the cauldron is associated with only a very weak gravity low. Both profiles show close relationships between small-scale gravity anomalies and local geological features.

No drastic change in the gravity field occurs at the transition between the chiefly effusive western part and the plutonic eastern part. The predominantly rhyolitic cauldron effusives are looked upon as representing the upper portion of the Vestfold lava sequence (Oftedahl 1967). The total thickness of the basaltic flows (B1–B5) is about 250 m in Vestfold; the gravity field, however, shows no sign of possible subsided basalts (or other dense rocks) within the Ramnes complex. Unless balanced by subjacent mass deficiencies in the western part, it is therefore uncertain as to whether the lower lava series was originally present in this area at all. Alternatively, the dense rocks may have sunk into and been dispersed within a large, hypothetical, underlying magma.

In Fig. 45b a residual anomaly is obtained for the N–S profile by subtracting the regional anomaly of Plate 4 from the observed gravity. The straight (dashed) line with a slight positive gradient is referred to as the Horten–Tønsberg high. The remaining residual (below dashed line in *b*) is ascribed to the rocks of the cauldron subsidence. A simple two-dimensional approach (Fig. 45c) shows that the anomaly may be simulated by rather small masses of felsic rocks along the margin and at the center. The geological interpretation of the

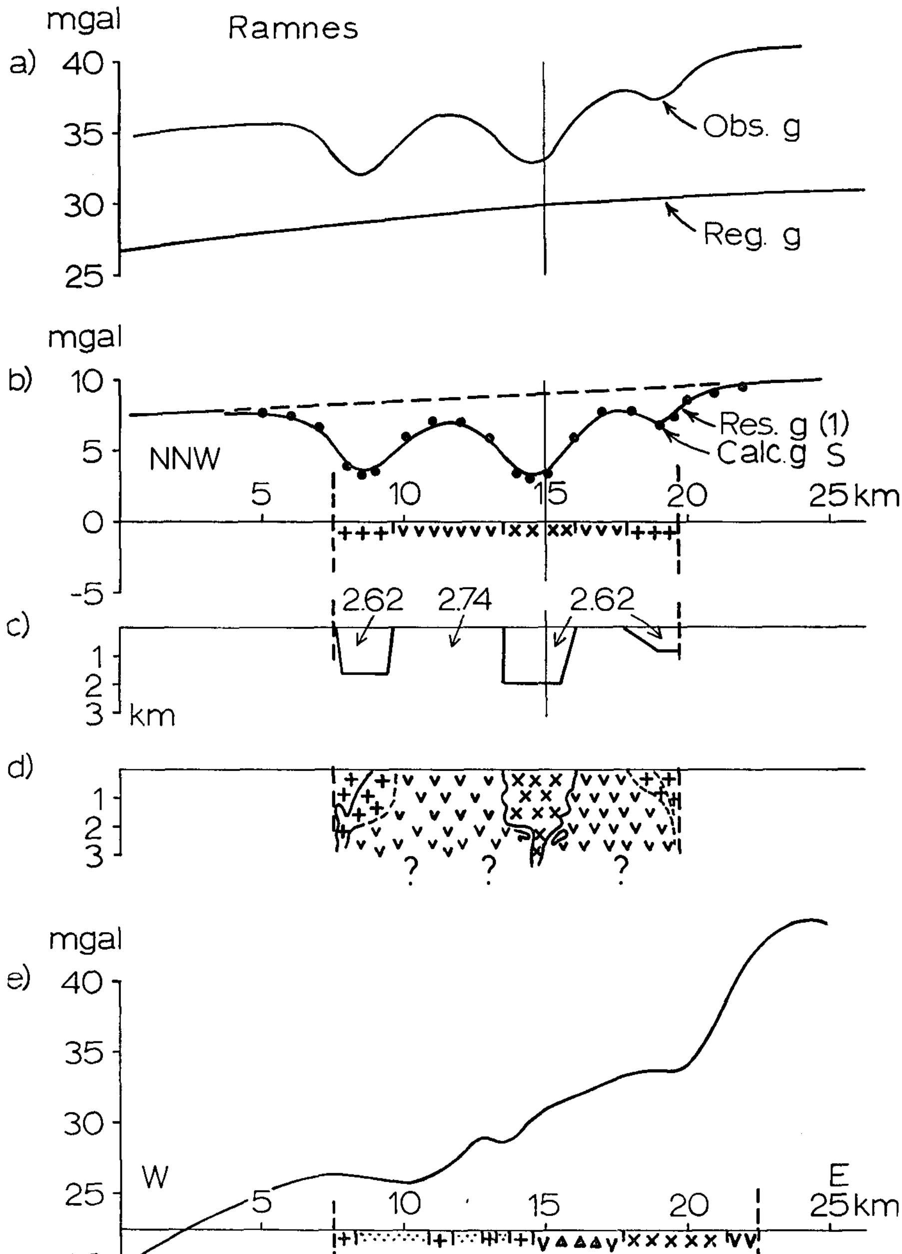


Fig. 45. Ramnes cauldron, (a) observed gravity profile (approx. N–S) and regional anomaly (Plate 4,) (b) residual and calculated anomaly, (c) geophysical model, and (d) geological interpretation. (e) E–W gravity profile Rock symbols as in Fig. 44, except for syenites (crosses), syenodiorites which here is grouped together (V's), and felsic volcanics (dots).

model is that the main central pluton of kjelsåsite-larvikite-syenite may have risen, cutting the syenite ring-dike, and itself being cut by a subsequent intrusion of granite and quartz-porphyry. The depth of the postulated central pluton

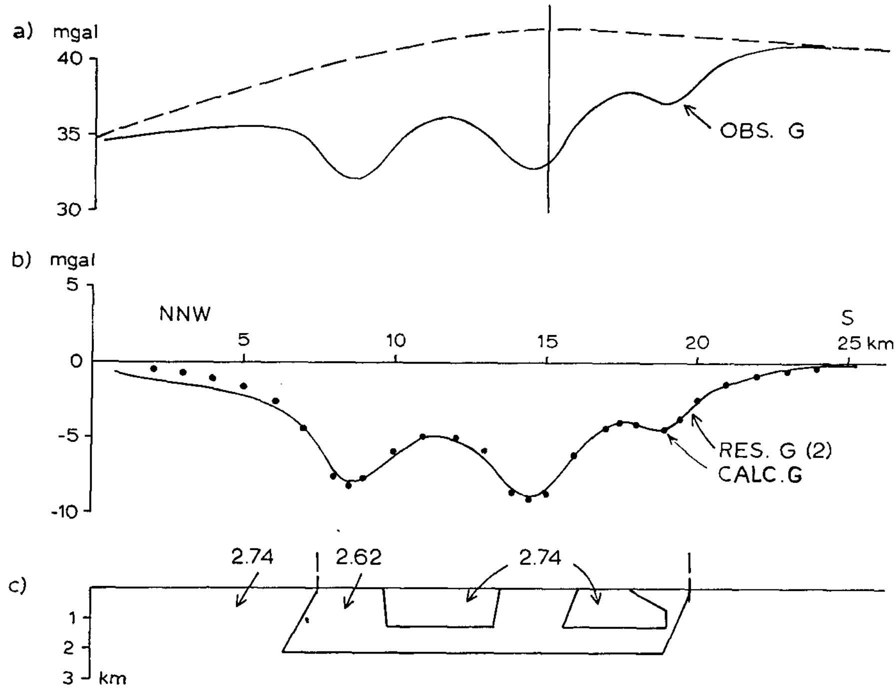


Fig. 46. Ramnes cauldron, (a) the N–S gravity profile with alternative 'regional' anomaly (see text), (b) and (c) alternative residual anomaly and geophysical model, respectively. Black dots: calculated anomaly.

cannot be revealed by means of gravity. A foundered block may or may not be present within the pluton.

An alternative gravity interpretation arises from the choice of a different Horten-Tønsberg anomaly. If we assume that the body which gives rise to the Horten-Tønsberg high was formed first, then it is a fair assumption that the gravity high attained a smooth elliptic configuration before it was deflected by the cauldron anomaly (Fig. 44). The resulting convex 'regional' and residual profiles are shown in Fig. 46a & b, respectively. The residual is satisfied by a simple two-dimensional model (Fig. 46c) which differs from the first alternative in that a thin layer of felsic rocks has had to be introduced below the entire eastern part of the cauldron subsidence. This could either represent compositional variations in the central pluton, or mean that a simple pluton of largely syenitic composition does not extend deeper than about 2–2.5 km, or, again, that a foundered block is present at that depth. Since the larvikites seem to represent the latest intrusive even in the Ramnes complex (Oftedahl 1967) and as the cause of the Horten–Tønsberg anomaly is believed to post-date the emplacement of the larvikites (Ramberg & Smithson 1971), the first alternative (Fig. 45c, d) is regarded as the most likely interpretation.

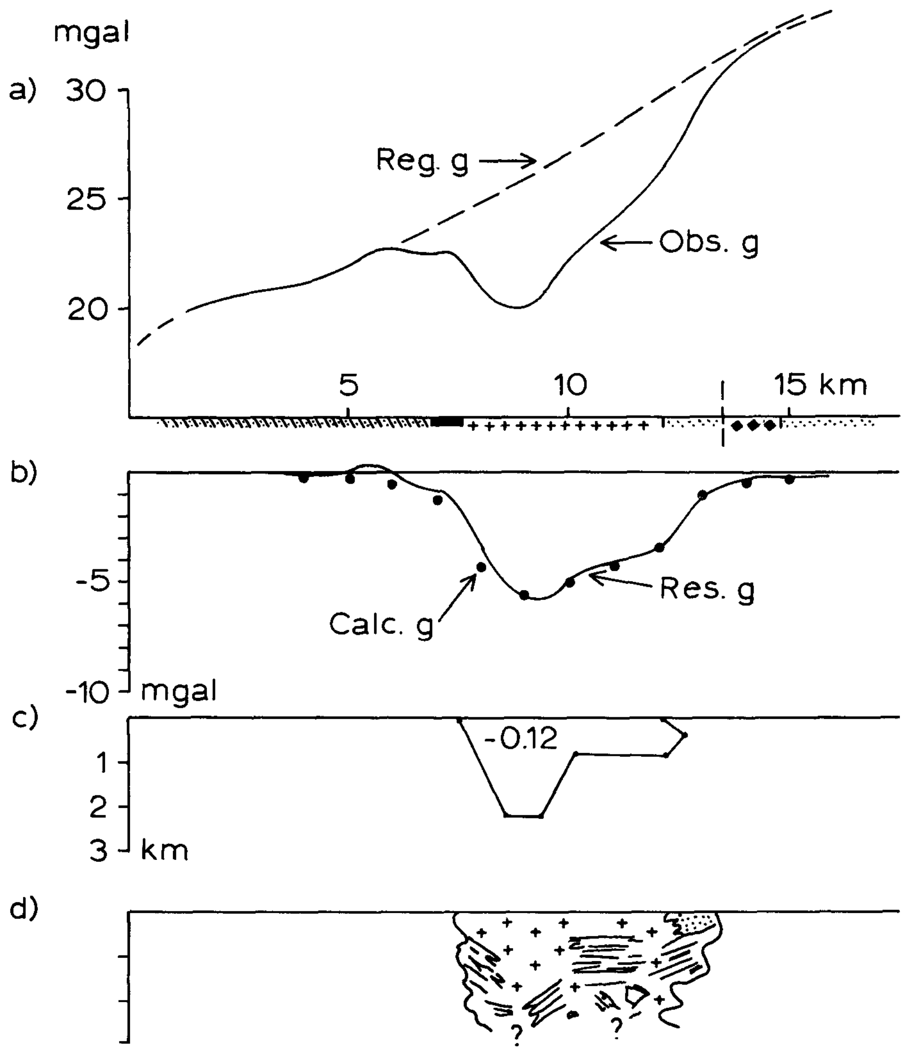


Fig. 47. Hillestad cauldron, (a) observed gravity profile and assumed 'regional' anomaly (sharp rise because of proximity to the Horten-Tønsberg high), (b) residual anomaly, (c) geophysical model, and (d) geological interpretation. Rock symbols: Downtonian sandstone (obliquely hatched and dotted), basaltic rocks (black), syenite (crosses), felsic volcanics (dots), rhomb-porphyry (black diamonds). For location of profile, see Fig. 34 (H1).

b. The Hillestad cauldron

This small complex was termed the Hillestad laccolith by Brøgger (1933a) but, like several similar complexes, is now acknowledged to have formed by a cauldron subsidence. It consists of a subsided block of felsic effusives and a younger, post-cauldron pluton of nordmarkitic composition in the NE central part (Everdingen 1960, Oftedahl 1967).

A gravity low occurs above the central part of the nordmarkite intrusive (Fig. 34), but the gravity coverage is not very good and only a simple two-

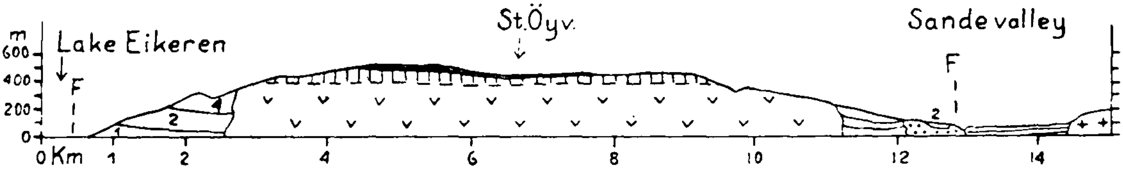


Fig. 48. Earlier interpretational section (WSW–ENE) through the Sande cauldron (*after* Oftedahl 1953). Symbols: Alkali granite or ekerite (V's), syenitic rocks (vertically hatched), syenodioritic rocks (black), rhomb-porphyry (white with numerals), syenitic ring dike (dots).

dimensional model has been attempted (Fig. 47). The somewhat irregular gravity profile with a maximum anomaly of about 6–7 mgal may be accounted for by a wedge-shaped model with density contrast of -0.12 g/cm^3 and a maximum depth of about 2 km. This situation is very much like the one found for the large Nordmarka-Hurdalen batholith and other felsic intrusions, and calls for an alternative explanation. One such solution is sketched in the bottom part of Fig. 47 and assumes that thick blocks of the stratigraphical middle and lower parts of the lava series have descended into a nordmarkitic magma having a present thickness of at least 2–3 km.

c. *The Sande cauldron*

This cauldron (Fig. 34) is rather regularly shaped with a central intrusion surrounded by a ring-shaped area of subsided lavas and marginal, porphyritic intrusions along ring faults (Oftedahl 1953). The central intrusion is itself subdivided into three units in concentric arrangement: an outer zone of ekerite grades into nordmarkite which again shows gradational contacts toward a monzodioritic core.

There have been different opinions regarding the spatial distribution of the plutonic rock units. Based on field evidence and mineralogical studies, Oftedahl (1953) concluded that the kjelsåsite-larvikite core formed by assimilation of overlying lavas in the upper part of a granitic (ekeritic) magma. The resulting rock distribution is demonstrated in the interpretational section in Fig. 48. Geochemical studies (Raade 1973) have revealed that the basic core is characterized by the same uniform Th/U-ratio as in other kjelsåsite-larvikite occurrences in the Oslo Region. Raade, therefore, concluded that the core is composed of genuine kjelsåsite-larvikite rocks that were intruded from deeper levels.

Fig. 49 shows an E–W gravity profile across the Sande complex. The 'cauldron anomaly' clearly stands up as a high between the gravity lows associated with the adjacent ekerite batholith to the west and the Drammen granite to the east. This feature must be caused by a slight mass *surplus* below the cauldron area; no extensive masses of felsic rocks can possibly exist. From the profile it is further seen how gravity highs and lows are closely connected with the outcrops of lavas and ekerite-nordmarkite, respectively.

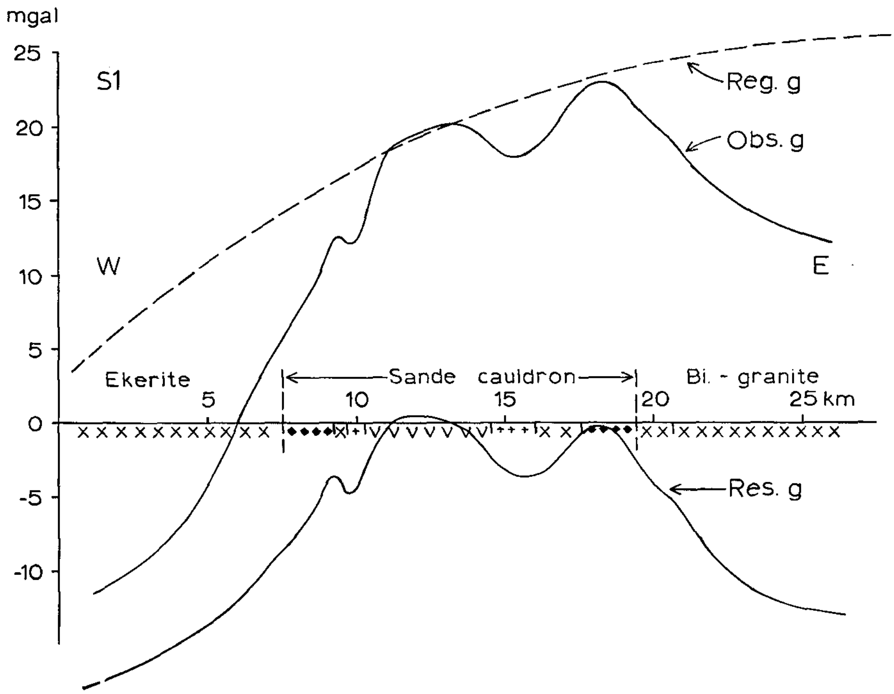


Fig. 49. E–W gravity profile across the Sande cauldron. Regional anomaly from Plate 4. Note how the complex is marked with a relative gravity high with respect to the lows associated with the alkali granite (ekerite) and biotite granite complexes on both sides. Granitic rocks (X'es), syenitic rocks (crosses), syenodiorite (V's), rhomb-porphyry (diamonds). For location of profile, see Fig. 34 (S1).

These features are even better seen in Fig. 50, a NE–SW profile along the most densely measured parts of the complex (see Fig. 34). The southern end of the profile is slightly negative because of the flank effect of the ekerite massif. To the north the effect of the rather shallow Drammen granite has been included. Applying the observed densities (Table 5, Fig. 16) and the geological information available, the felsic rocks (nordmarkite, ekerite) clearly have a rather limited vertical extent. The density contrast represented by the monzodioritic core widens downward and apparently extends to at least 3.5 km depth. However, the small density contrast between the monzodioritic rocks and Precambrian gneisses makes any realistic depth estimates impossible. On the other hand, if any plutonic rocks underlie the cauldron subsidence to some depth, as generally assumed, the gravity data would signify that their overall density is close to that of the intermediate rocks.

Relatively large mass surpluses underlie the exposed lava ring. At the southern end of the profile (Fig. 50) these lavas are B1 and RP1. Here the vertical throw is estimated to amount only to about 100 m (Oftedahl 1953). Thus, it is evident that the local gravity high cannot be caused by the lavas alone. This conclusion applies to the 'marginal high' all the way around, the maximum subsidence (at the northern end) being estimated as about 900 m (Oftedahl

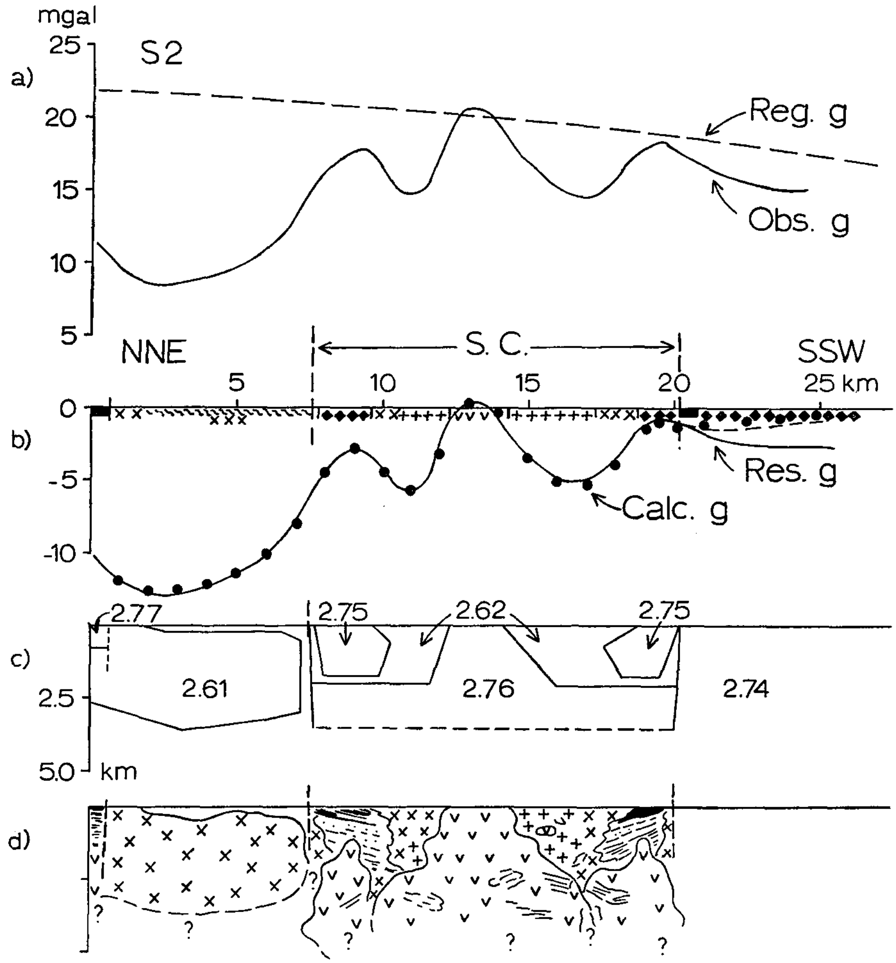


Fig. 50. Gravity and geological interpretations along a NNE–SSW profile across the Sande cauldron. Regional anomaly from Plate 4. Rock symbols as in Fig. 49 with the addition of Downtonian sandstone (obliquely hatched) and basalt (black). S.C. = Sande Cauldron. For location of profile, see Fif. 34 (S2).

1953). To form the observed gravity highs the lavas are consequently underlain either by subsided Cambro-Silurian rocks or intermediate to basic intrusives. The latter is possibly the case, especially below the lavas in the eastern part of the complex where the gravity high forms a northern extension of the Horten-Tønsberg high.

To summarize, the Sande complex does not involve large quantities of felsic rocks. If underlain by a large pluton, its overall composition must be close to that of the monzodioritic rocks in the core. This is in accordance with the results of Raade (1973, p. 72) who divides the central intrusion into two separate magmatic rock groups, "one differentiation series of kjelsåsite-larvikite-syenite porphyry, probably connected with the subsidence of the cauldron, and a younger ekerite magma giving rise to the quartz porphyry ring-dikes and to nordmarkite by assimilation".

d. The Drammen cauldron

The somewhat smaller Drammen cauldron (diameter about 7 km) is completely surrounded by the Drammen granite complex (Fig. 33). Its western part is made up of a subsided block of heavily faulted lava flows (B1 – RP13), while the eastern part is covered by the flat floor of the Lier valley and the inner part of the Drammensfjord. The map of Fig. 33 is largely based on unpublished excursion maps from the Drammen cauldron by Mr. Odd Halsen.

The cauldron subsidence has generally been connected with the emplacement of the Drammen granite batholith (Oftedahl 1953, 1960). Current studies (Gaut, pers. comm. 1972), imply, however, that the formational sequence may be quite complicated.

As seen from Plate 5 and Fig. 33 the Drammen cauldron is clearly marked as a gravity high relative to the larger low associated with the granite complex. The 'cauldron high' amounts to about 8–10 mgal in the west and slopes off to the east. It is overprinted by a local low that can be traced along most of the Lier valley. The low is correlated with the Quaternary deposits and may rise to about 3 mgal.

Close to the town of Drammen is a sharp aero-magnetic anomaly that has been interpreted as being caused by a concealed volcanic neck (Oftedahl & Åm 1971). This anomaly is not reflected in the gravity field despite a rather dense station coverage in the area. The hypothetical volcanic neck consequently has to consist of rocks of intermediate composition and negligible density contrast to the country rocks, or alternatively the magnetic anomaly is associated with some kind of explosion breccia with a relatively high content of magnetite.

An E-W profile across the Drammen granite and the Drammen cauldron has been interpreted (Fig. 51). Observed densities were applied for the granite (2.61 g/cm^3) and the country rock (2.74 g/cm^3), while a density of about 2.15 g/cm^3 was assumed for the Quaternary sedimentary deposits. Assuming the extreme possibility that, except for the exposed lava sequence, the 'cauldron high' may be caused by basaltic rocks only (Case I), a slab of more than 1 km thickness had to be applied to fit the calculated anomaly with the residual. Compared with the possible thickness of B1 + B3 of 200–250 m in the Drammen cauldron, and B1 + B3 of about 200 m in the Vestfold area (Oftedahl 1960, 1967), this value clearly seems too high. Also, since the *average* density of the subsided block is certainly distinctly lower than 2.90 g/cm^3, the thickness of the block would have to be several kilometers to match the observed gravity data. If Cambro-Silurian rocks are part of the subsided block, the implied thickness is of course possible. In any case this model suggests that the amount of subsidence is larger than earlier believed 500 m.

Another possibility (Case II) is that the subsided block of effusives (with an observed average density of about 2.77 g/cm^3) has a thickness of only 0.5–0.6 km. In this case, the block has to be underlain by a body of intermediate density and with at least the same thickness as the Drammen granite.

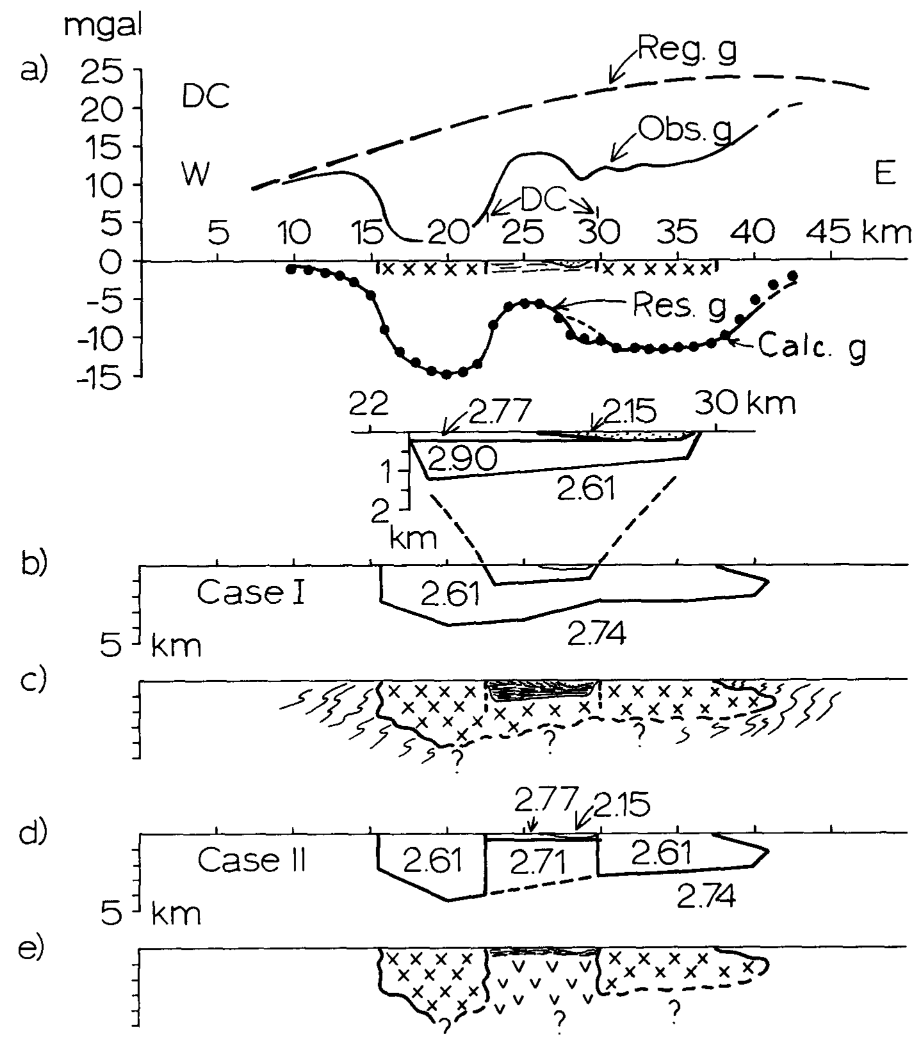


Fig. 51. Drammen cauldron, (a) observed gravity profile, regional anomaly (Plate 4), residual and calculated anomaly, (b) and (c) geophysical and geological interpretations, Case I, (d) and (e) Case II. Drammen granite (crosses), rocks of intermediate compositions (V's), volcanics (horizontal lines), Lier Valley Quaternary fill (dots). For location of profile, see Fig. 33 (DC).

This could represent brecciated and partly intruded Cambro-Silurian and Precambrian rocks, or more likely plutonic rocks of intermediate composition. Combinations of Cases I and II would certainly also fit the data.

The gravity data has revealed that the Drammen cauldron is underlain by a relative mass excess. Whether this surplus is partly due to monzonitic intrusives, as seems to be the case in the Sande and Ramnes cauldrons, cannot be decided. If, on the other hand, the entire gravity high is caused by a subsided block of alternating rhomb-porphyry and basaltic flows and Cambro-Silurian rocks, this block will be very thick and leave little room for granitic rocks below. It seems unreasonable that a huge cauldron subsidence should be

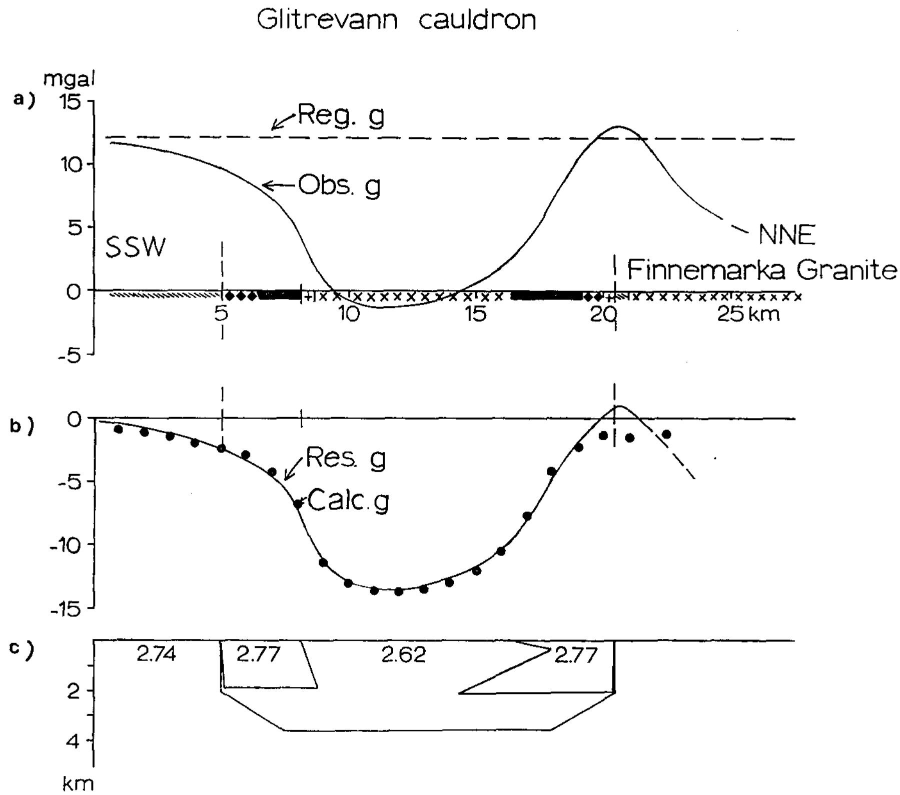


Fig. 52. Glitrevann cauldron, observed gravity profile (see Fig. 33) assumed regional (from Plate 4), residual anomaly, and geophysical model. Rock symbols as in previous profiles (Figs. 50–51). Note gravity high at the contact between the Finnemarka and the Glitrevann complexes.

floored by a negligible volume of intrusives. Hence, in both cases plutonic rocks of higher density than the granite will have to occur below the granite in the cauldron area.

e. The Glitrevann cauldron

The large Glitrevann cauldron (10 x 16 km) has been extensively studied by Oftedahl (1953). It consists of a central pluton of aplitic biotite granite and quartz porphyry surrounded by an incomplete ring of syenite porphyry (Fig. 33). An outer belt of subsided lavas (B1–RP13) shows a general inward dip (see Oftedahl 1953, fig. 9). Ring dikes consist of quartz porphyry and syenite porphyry. The Glitrevann complex may possibly subdivide into an older and a younger area of subsidence (Gaut & Naterstad, pers. comm. 1972) as indicated in Fig. 33.

Oftedahl (1953) assumed that the cauldron subsidence was caused by the granitic magma being continuous from the Drammen complex to Tyrifjorden.

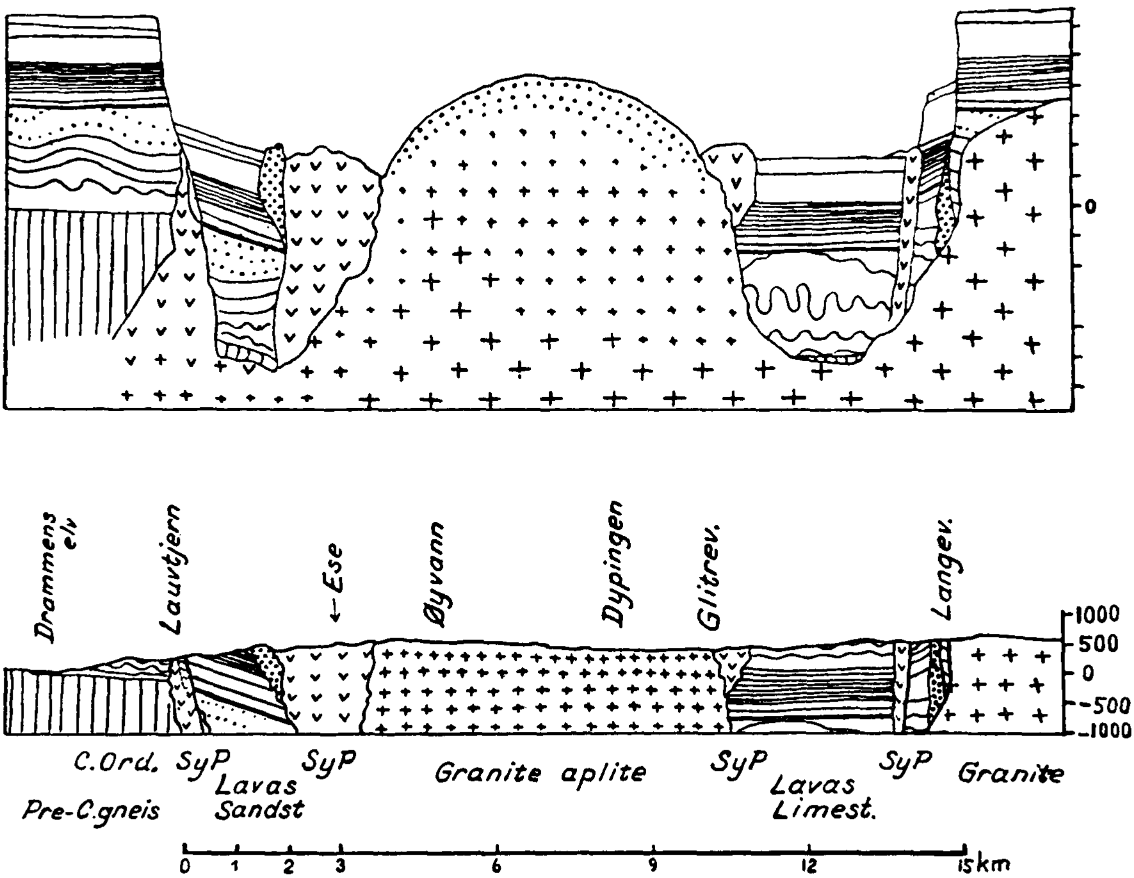


Fig. 53. Interpretational N–S section through the Glitrevann cauldron (*after* Oftedahl 1953) that accords with the gravity model.

Raade (1973) concluded that the syenite porphyry – aplitic granite form a comagmatic series with regularly increasing U, Th and Th/U-ratio towards the more acid members, and that the central intrusion seems to represent the end product of a local differentiation series not genetically related to either the Drammen or the Finnemarka granite complexes.

The cauldron is marked by an almost circular gravity low centered on top of the central pluton (Plate 5, Fig. 33). It marks a substantial mass deficiency of roughly cylindrical or cone-shaped structure. The individual gravity lows associated with the felsic occurrences in the Glitrevann, Finnemarka and Drammen granite complexes, and the high gravity readings in the Lier area east of Glitrevann, indicate that no continuous subsurface layer of felsic material is present.

A two-dimensional model along a NNE-SSW profile (Fig. 52) shows that the density contrast may extend to 3.6 km depth. For the subsided blocks of rhomb-porhyry and basalt flows plus underlying Cambro-Silurian rocks, an average density of about 2.77 g/cm^3 has been assumed (Table 5). The model indicates some widening of the subsided blocks with depth. Still the high gravity values at the border region between the Glitrevann and Finnemarka granite complexes have not been fully accounted for; it is suggested that high density basement rocks follow a narrow belt from the Asker-Lier high to the Tyrifjord high (see Section 6.6).

A geological interpretation of the model in Fig. 52 is almost perfectly illustrated by the geological N–S sections presented by Oftedahl (1953, fig. 19), here reproduced as Fig. 53. The only principal difference is that gravity data do not prove any connection between the central pluton and the Finnemarka and Drammen granites to the north and south, respectively. With regard to the depth of the pluton, the two-dimensional model indicates that the densiy contrast extends *at least* to a depth of 3.5–4.0 km (see also Fig. 30). No feeder or wedge-shaped continuation has been indicated; if these exist they have to represent rather minor volumes. The flat-bottomed gravity signature across the subsidence shows some resemblance to the theoretical configuration displayed in Fig. 22a & b, and hence the felsic rocks may sit atop a central foundered block or above intrusives of an overall intermediate composition. By analogy with, for instance, the Ramnes cauldron, the latter interpretation is favored.

f. The Nittedal cauldron

Having a diameter of about 12 km, the Nittedal cauldron (Naterstad 1971) consists of faulted blocks of Precambrian and Cambro-Silurian rocks, Permian lavas (B1–RP 13) and overlying sediments in an irregular marginal arc. Syenitic intrusions follow the original ring fractures. Post-cauldron nordmarkite and alkali-granite intrusions (parts of the large Nordmarka-Hurdalen batholith) occupy the central and western parts. The amount of subsidence is considerable, possibly about 1000 m. Detailed accounts from various subregions have been given by Holtedahl (1943), Sæther (1946, 1962) and McCulloh (1952).

The gravity maps (Plate 5, Fig. 54) show that the gravity low over the cauldron is part of the larger gravity trough correlated with the Nordmarka-Hurdalen syenitic batholith. No models of the felsic rocks will be presented other than those given in Figs. 26–28. From the age relations indicated by Naterstad (1971) – (1) Holterkollen granite and monzonite, (2) ring dikes, and (3) syenitic central pluton – it seems logical to associate the cauldron subsidence with movements in the underlying, extensive syenitic pluton. Furthermore, the system of ring fractures in the Precambrian to the east of the complex (Fig. 8) is possibly related to the same rock masses and not to the granite. This is supported by the aeromagnetic anomalies over the ring-fractured gneiss terrain. Such anomalies seem to correlate with the syenitic (and monzonitic) batholiths, and not with the granites which in the Oslo Region are characterized by a lower content of magnetite (Åm & Oftedahl, in prep.). Thus, the felsic rocks in the gravity models from the Nittedal area (Figs. 26–28) may partly be syenitic rocks and partly older granitic (as in Holterkollen) intrusions.

The subsided blocks of lavas and Cambro-Silurian rocks in the north and

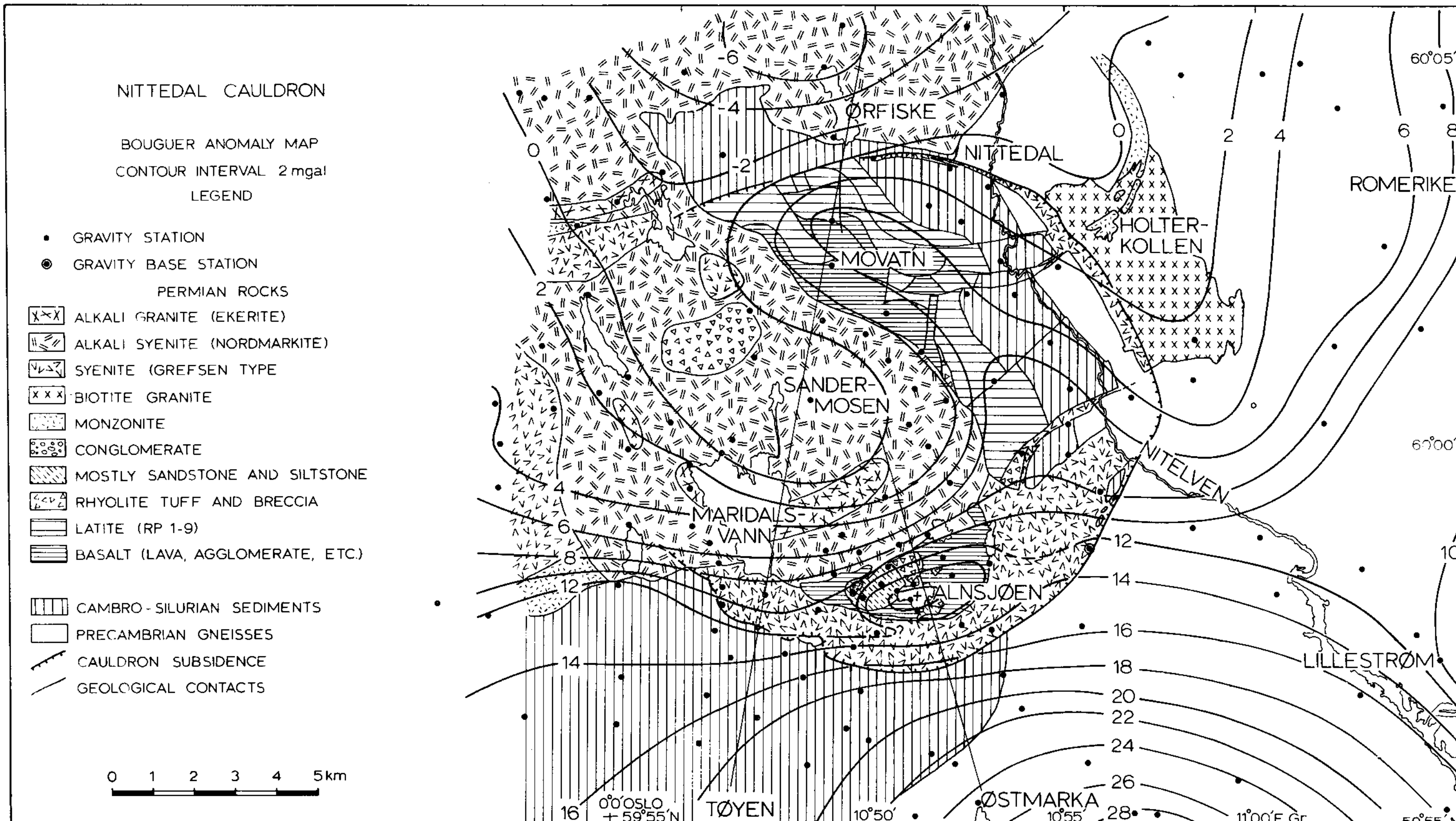


Fig. 54. Detailed Bouguer anomaly map of the Nittedal cauldron. Geology *after* Naterstad (1971) and Sæther (1962).

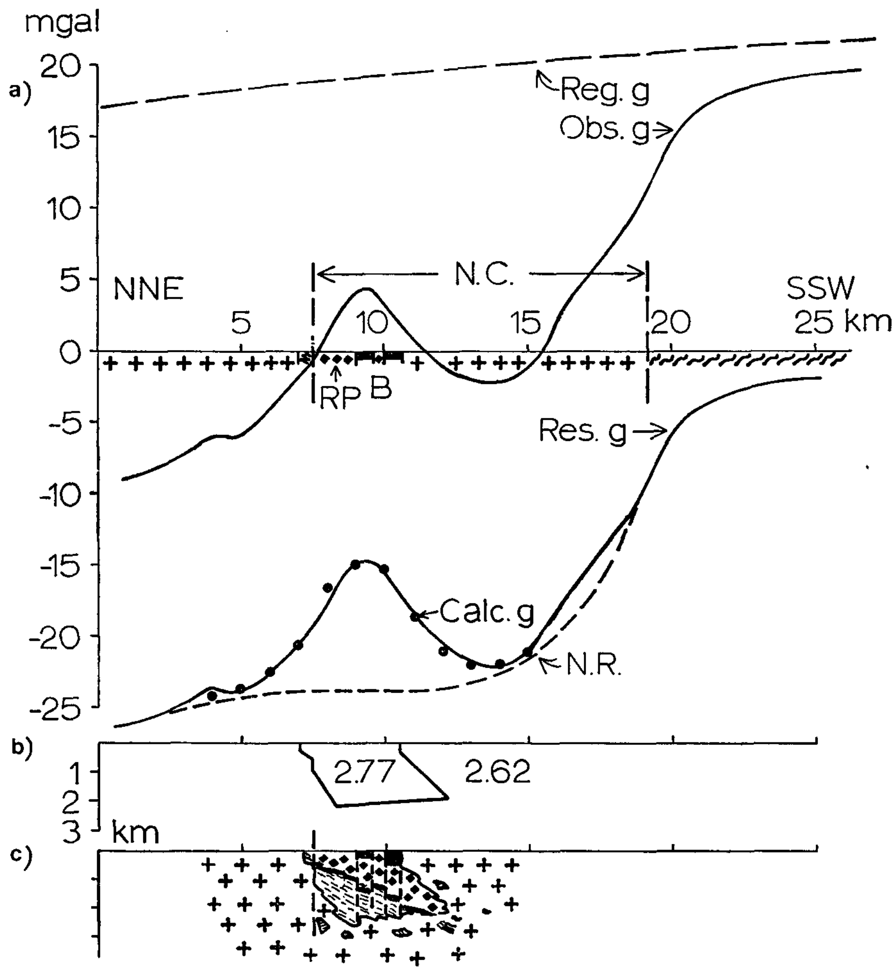


Fig. 55. Nittedal cauldron, (a) observed gravity profile (Fig. 54), assumed regional (Plate 4), residual anomaly, geophysical and geological interpretation of subsided supracrustal rocks. N.R. = Nordmarka-Hurdalen (batholith) residual. Rock symbols as in Fig. 50.

east are marked by a gravity high superimposed upon the larger gravity low. In a roughly N–S profile (see Fig. 54), a two-dimensional model shows that the sharp high may be explained by a subsided block of mainly rhomb-porphyry and basaltic flows (Fig. 55). With a somewhat gentler average dip of the layers, as indicated on Naterstad's (1971) profile somewhat more to the east, less basaltic material will be part of the postulated block which therefore has to be even thicker (2–3 km) to balance the observed profile. The block will then have to include more Cambro-Silurian rocks in its deeper parts. The gravity high is most marked in the northern part, but is apparently continuous towards the east and south indicating that the amount of subsided material close to the surface is about the same in all these areas. To the west no trace of shallow, dense material is found. If originally present, the supracrustal rocks must have sunk to great depths as stoped blocks in the syenitic magma.

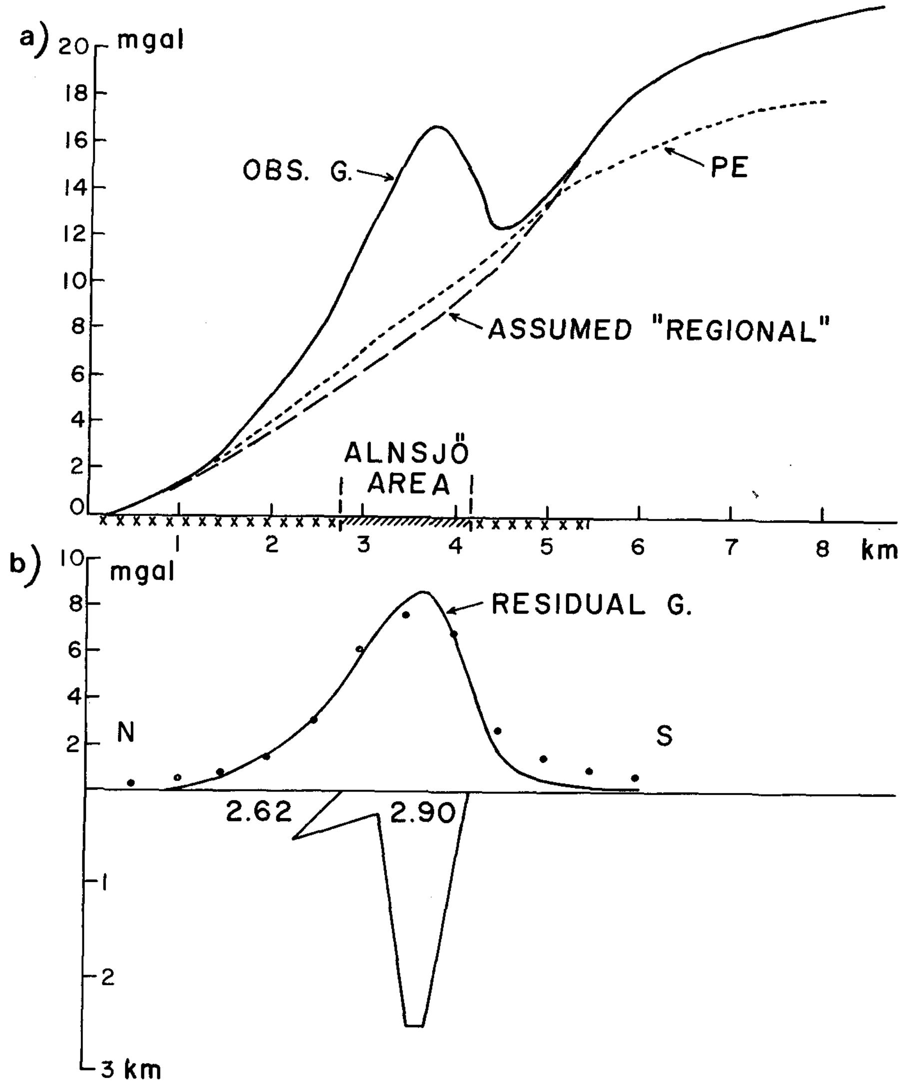


Fig. 56. Alnsjø area; gravity profiles and geophysical interpretation. Location of profile, Fig. 54. Rock symbols: Syenitic rocks (X'es), the Alnsjø supracrustals, obliquely hatched. PE = profile to the east of the Alnsjø area. Gravity model simulates extreme case with basaltic pipe overlain by a very thin screen of conglomerates, sand and siltstone. Difference between residual and calculated anomaly (block dots) on the south flank may be due to low density syenitic ring dike not included in the present model.

Especially high gravity readings occur in a restricted area near lake Alnsjøen in the southern part of the complex (Fig. 54). The gravity high coincides with a strong aeromagnetic anomaly interpreted as being caused by a volcanic neck below possible crater deposits (Oftedahl & Åm 1971). The gravity measurements support the assumption of having dense rocks below the supracrustals in the Alnsjøen–Storhaug area, while low gravity anomalies prevail over the basaltic rocks to the east and west of Alnsjøen and indicate rather small thicknesses of basalt, etc., in these regions, consistent with geological observations (Holtedahl 1943, Naterstad 1971). On the assumption that the gravity high is caused by rocks of basaltic composition, Fig. 56 shows one possible interpretation. The two-dimensional approach represents an under-estimate of the thickness by about 10% (using Nettleton's (1940) end correction), implying that the structure might actually represent a volcanic center as mentioned earlier by Holtedahl (1943). Alternatively, if the anomaly is caused by supracrustals of somewhat lower average density, the thickness would be correspondingly greater. However, the rather narrow, pipe-like extension required to fit the flank gravity values, favors the first alternative. In order to better match the residual and calculated peak values, the 'pipe' would have to extend even deeper.

g. Concluding remarks

The cauldrons of the Oslo Region represent root or subsurface expressions of calderas or volcanic ring complexes. From field observations most of them are characterized by having central plutons or domes. The gravity data has revealed that three of the complexes studied here (Hillestad, Glitrevann, Nittedal) are associated with significant negative anomalies correlated with felsic, partly post-subsidence intrusives, but the vertical extent of the density contrasts is rather limited (3–4 km) in the cases of Hillestad and Glitrevann. These depths would be expected if crustal blocks had traded place with the magmas (Figs. 21b, 22a), but it is regarded equally likely that the felsic complexes are floored by intermediate intrusives. Two of the complexes (Ramnes and Sande) seem to ride on central intrusions of an overall intermediate composition, while only minor volumes of felsic rocks are found. The same conclusion may apply to the Drammen cauldron, while an alternative interpretation involving thick blocks of subsided supracrustal rocks is equally satisfactory from a gravity point of view. From the Bærum complex the large akerite porphyry in the central area may prove to represent an intrusive dome. Similar latite or trachyte domes exist within the Øyangen cauldron remnant in the Ringkollen area (Ramberg & Larsen, in prep.) In summary, the gravity data restrict the extent of the felsic rocks and show that denser rocks occur at shallow levels within most of the cauldrons. These rocks may be either foundered blocks or basaltic or syenodioritic intrusives.

On the basis of the distribution of radiometric elements, Raade (1973) concluded that the central plutons in the cauldrons studied by him (Ramnes, Sande, Glitrevann) are composed of a differentiated comagmatic series. Compared with the gravity interpretations and the diversity of intrusive rocks found, this implies that the various central plutons represent individual intrusive centers. Nested into each other, they may form larger composite batholiths, as seen in the northern part of the Oslo Region. The horizontal section from the margin towards the center in complexes like the Sande cauldron may be tentatively looked upon as a model of a vertical section through the Oslo Region

Table 12: Anomalous mass, total mass, and volume of felsic intrusives in some of the cauldron subsidences

Cauldron Subsidence	Anomalous Mass, ΔM	Total Mass, M	Volume, V
Ramnes	0.18×10^{17}g	3.5×10^{17}g	1.39×10^{17} cm^3
Hillestad	0.01 x »	0.3 x »	.11 x »
Sande	0.11 x »	2.2 x »	.83 x »
Glitrevann	0.31 x »	6.1 x »	2.34 x »
Drammen	?	–	–
Total	0.61×10^{17}g	12.1×10^{17}g	4.57×10^{17} cm^3

in general, the mafic center representing the deeper part. This picture is in agreement with the common belief (Macdonald 1972) that differentiation takes place within the underlying magmatic bodies of the ring complexes, producing a portion of magma at the top of the chamber that is more acidic than that deeper down. In the present case, gravity measurements may have revealed just this acidic top portion in some of the complexes (e.g., Hillestad, Glitrevann, Nittedal), while in other complexes (Ramnes, Sande, Drammen(?)) the syenodioritic deep-seated rocks evidently rose to or remained at higher levels.

Cauldron formation usually takes place late in the history of a volcano (Macdonald 1972). Possible occurrences of volcanic necks or subvolcanic chambers in the Drammen and Nittedal cauldrons, evidence of early crustal warping and basaltic volcanism in the Øyangen complex (B. T. Larsen pers. comm. 1974), as well as inclusions of 'Oslo-essexites' within the effusives of the Glitrevann complex, support the idea that the Oslo Region cauldron subsidences represent the continued development of fractionating volcanic centers, concealed by cauldron fill and partly digested by rising central plutons. Thus, we can recognize the following general development: fault-controlled, early, central (basalt) volcanoes, differentiation, rise of central plutons and cauldron subsidence, and continued post-cauldron (felsic) intrusion.

Estimates of the mass and volume of the felsic rocks assumed to occur within the cauldrons studied are given in Table 12. The anomalous masses, etc., of the Nittedal complex as well as the Bærum and Øyangen complexes have been included earlier with the estimates of the Nordmarka-Hurdalen complex. The values listed in Table 12 have been added in Table 9 to obtain the total estimate for the Oslo Region felsic intrusives.

6.6 OTHER PROMINENT GRAVITY ANOMALIES

Six local gravity highs on the residual map (Plate 5) will be considered in some detail. Only one of the anomalies is clearly associated with outcropping dense rocks, this being the Skien high situated above a partly subsided slab of basaltic rocks in the southwest corner of the region. The Kongsberg and Tyrifjord gravity highs are located close to the western border of the Oslo Region. Both

Table 13: Estimates of maximum possible depths to the top surface of the gravitating dense bodies

Gravity high	Depth (2D) km	Depth (3D) km
Narrefjell	3.5	4.7
Kongsberg	1.9	2.5
Tyrifjord	3.9	5.1
Asker-Lier	1.4	1.9
Horten-Tønsberg	3.9	5.1

these anomalies are elongated in the NNE direction parallel to the border; the anomalies, however, seem to be caused by Precambrian, not by Permian rocks. Within the Oslo Region proper, the Asker–Lier, the Horten–Tønsberg and the Narrefjell highs are not correlated with any known occurrences of dense rocks. On the contrary, they show cross-cutting relations to surface rocks of contrasting densities, testifying to the local existence of denser rocks at shallow depths.

Maximum depth calculations for both two-dimensional and three-dimensional cases have been carried out according to the method outlined by Bott & Smith (1958), based on the maximum gravity anomaly and the maximum gravity gradient from residual maps (Table 13).

Considering the likely possibility that the causal masses of the Kongsberg and Tyrifjord anomalies may be exposed (see below), with the fact that maximum depth calculations commonly greatly overestimate the depths, the tops of the anomaly-producing bodies have to be situated at quite shallow levels. Since some of the anomalies are caused by material whose nature and density is largely unknown, any detailed interpretations are unwarranted in these cases.

a. The Skien high

This 25 km-long gravity high is defined by only three gravity road profiles. Although no gravity details are revealed, the anomaly shows a clear correlation with the outcropping, elongate Skien–Porsgrunn basalts (B1) (Plate 5).

The basalts, which rest on a relatively thin sequence of basal Permian sediments on top of Downtonian sandstones, range up to a maximum width of ca. 5 km in map projection. Although variable, the dip is generally 30°–60° towards the nordmarkites and larvikites to the east (Oftedahl 1952). The total thickness of the basalts has been variously estimated to be about 100–200 m. From his schematic cross section along the highway from Skien to Siljan (Fig. 57), Oftedahl (1952) tentatively concluded that the thickness may possibly be as much as 500 m or more.

A gravity profile parallel to the same Skien-Siljan section (Fig. 58), shows a sharp gravity high of about 12 mgal centered on the basalts. For the calculations, an average density of 2.72 g/cm^3 has been applied for the Cambro-

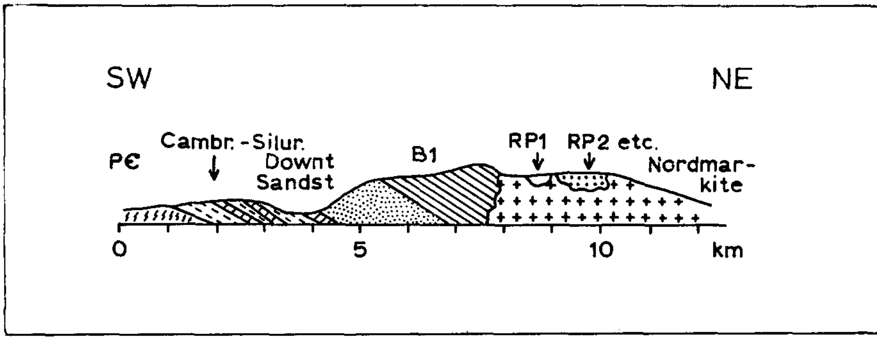


Fig. 57. Schematic cross section along the highway from Skien to Siljan, southwestern part of the Oslo Region (*after* Oftedahl 1952).

Silurian sequence and overlying sandstones (Table 5). A body of density 2.90 g/cm^3, about 2.5 km thick and about 5 km wide at its lower level, will match the residual profile. Various modifications of the model are of course possible. The shape of the profile, however, seems to exclude the existence of any subsurface continuation of the dense slab far to the east. Also, the peak value is unreasonably high to have resulted from basalt thicknesses of about 200 m or even 500 m, even if stoped blocks and fragments of basaltic rocks may have been piled up within the intrusive rocks beneath the outcropping basalts.

The geological interpretations of the above model can be separated into two principal types (Fig. 58c & d). In the first case the basaltic rocks form a slab with an apparent stratigraphic thickness of about 2 km, partly broken up as fragments within the nordmarkite-larvikite mass and extending down to at least 2.5 km depth. Contact metamorphic phenomena in the sediments to the west of the basalts show that intrusive rocks occur at shallow levels. If the intrusive rocks below the subsided basaltic block are of chiefly felsitic composition, this rock has to contain a number of basaltic xenoliths also below the 2.5 km level in order not to produce an interfering gravity low. Thus, a hypothetical, deep body of felsic rocks beneath the basalts could mean that the basalt residual is actually still higher than assumed. The computed depth of the basalt of about 2.5 km will in both cases represent a minimum value. Brøgger (1911, see Oftedahl 1952) suggested that the apparent huge thickness of the basalts was due to tectonic repetition of B1. No indications of faults parallel to the strike have been observed (Oftedahl 1952); the apparent thickness therefore seems to be realistic.

The alternative interpretation (Fig. 58d) is to assume that the main basalt block is underlain by an extensive body of dense rocks, possibly in the Precambrian rocks. The mass could also be a kind of subvolcanic reservoir or some completely unrelated rocks. Whatever the nature of the mass surplus, it has to underlie the basalts all along their strike direction for about 25 km and never penetrate much outside the general outline of the basalts. As an elongate reservoir of these dimensions and with such constraints seems unreasonable, the first alternative is preferred.

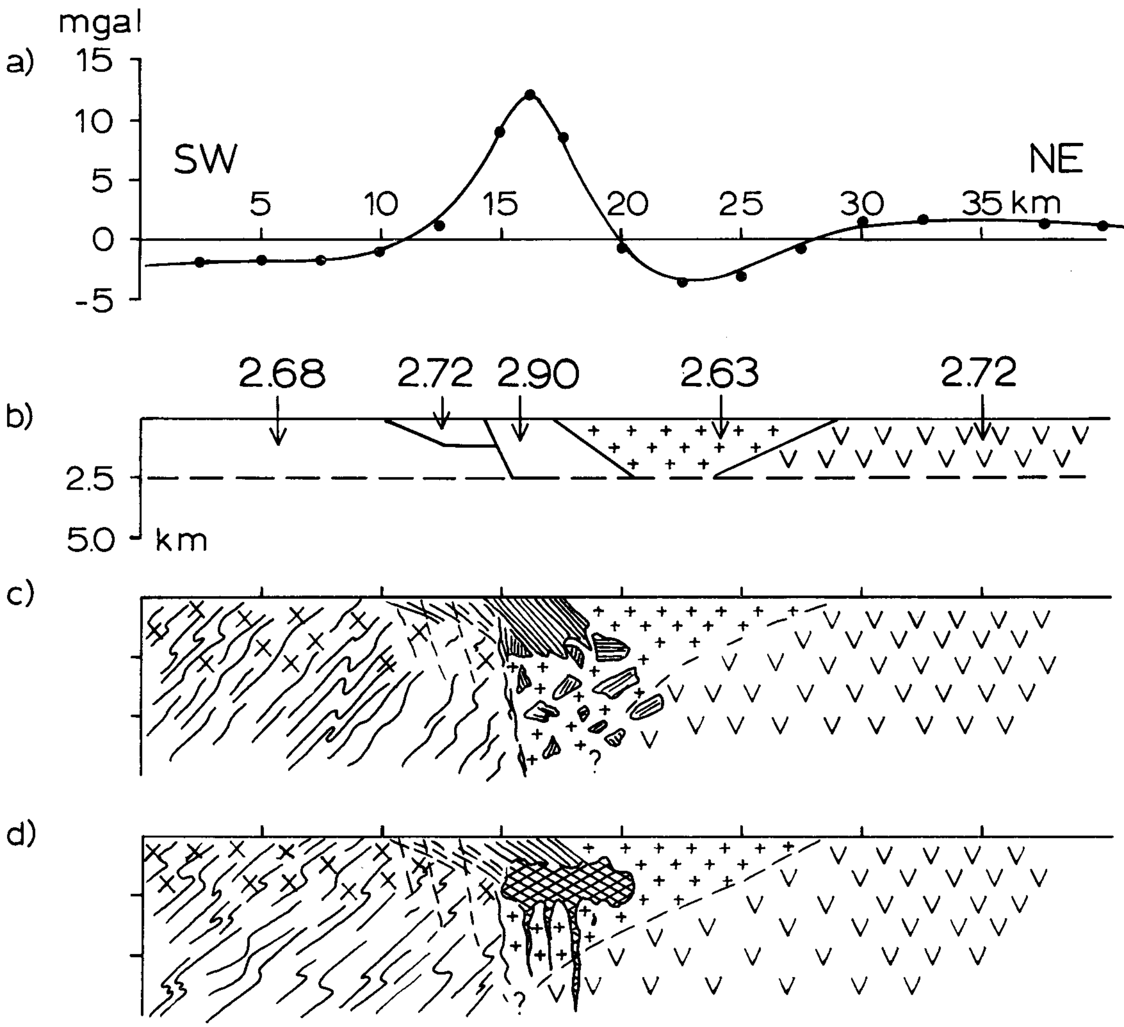


Fig. 58. The Skien basalts, (a) residual gravity profile (Plate 5) and calculated gravity (black dots), (b) geophysical model, (c) and (d) alternative geological interpretations. Profile follows roughly the highway Skien-Siljan. Rock symbols: Syenitic rocks (crosses), monzonitic rocks (V's), mafic rocks in general (cross-hatched), basaltic rocks (parallel ruled), Precambrian gneisses (irregularly lined).

It seems possible that the Skien–Porsgrunn B1 basalts have a total thickness of the order of kilometers, while further north, e.g., in the Bærum area it is less than 20 m. It is not a single flow but consists of alternating aphyric basalt, pyroxene- and plagioclase-basalt. This, together with contrasting chemical characteristics found at various B1 localities (Weigand 1975), makes it reasonable to suggest that the appearance of basaltic centers was initiated in the Skien district, and that the B1 stratigraphical unit does not represent regionally continuous flows. The possible large thickness of the Skien basalts, if confirmed by detailed stratigraphical studies, might indicate a position close to or on the flank of a basalt volcano for these flows. It further implies

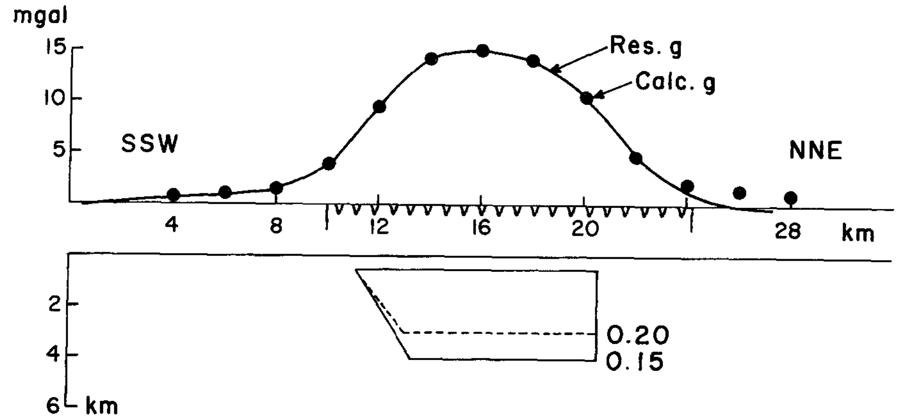


Fig. 59. The Narrefjell high. Gravity profile and models across the monzonite terrain.

that the lava thicknesses for stratigraphical lower units might have been much greater in parts of the Oslo Region than earlier conceived.

b. The Narrefjell high

A gravity high of about 15 x 35 km occurs within the Skrim monzonite massif (Plate 5). The peak value of about 15 mgal occurs close to the western border of the massif and at the NNE extension of the Skien high.

As can be seen from Fig. 16 there seems to be no correlation between density distribution of surface rocks and gravity anomaly. However, numerous xenoliths of basalt are known to occur in the outer parts of the monzonite pluton (Rohr-Torp 1969, 1973), probably remnants of basalt flows similar to those in the Skien district. Aeromagnetic data reveal that a possibly younger, subcircular, central portion of the Skrim massif is less magnetic than the marginal part (Åm & Oftedahl, in prep.), and a ring-dike has been observed in places between the inner and outer parts of this ring complex. The variation can be due to large-scale differences in content of magnetite, as can be observed elsewhere in the monzonites (e.g. Ula southeast of Larvik), or to the distribution of basalt inclusions.

If we assume that the Narrefjell high, which apparently cuts the contact between the inner and outer part of the complex, is caused by a subsurface accumulation of basaltic rocks, this hypothesis has been tested in the cross-section of Fig. 59. The two-dimensional model shows that with the observed density contrast between basalt and monzonite (2.90–2.71 g/m^3), a slab of almost 3 km thickness is required to fit the residual anomaly. This model is valid for a solid block of basalt. With just a little monzonitic material intermixed with hypothetical basalt fragments, the thickness of the bulky body will increase as exemplified by the density contrast 0.15 g/cm^3 on the figure. Basalt thickness is about 3–400 m in the Skreifjella area just to the north

(Rohr-Torp 1973). Even when considering the possible thickness of the Skien–Porsgrunn basalts, the 'available' basalt volume in any neighboring area does not justify these models. The Narrefjell high apparently must have other causes.

Segregation of magnetite will lead to density variations, but certainly not to overall contrasts exceeding 0.05 g/cm^3. However, models applying density contrasts less than 0.1 g/cm^3 have to be made unreasonably thick to reach the peak value but will still not fit the residual flank values The minimum plausible density contrast is in the range 0.1–0.15 g/cm^3 Thus it is suggested that the Narrefjell high is caused by basic intrusive rocks, or by the combined effect of subsided basaltic xenoliths, as is suggestive from the NGU aeromagnetic map, and some dense intrusives beneath the subcircular Skrim batholith in general. While the gravity data have definitely excluded any possibility of extensive subsurface masses of alkali-granite (as appears likely from geological evidence), it now seems plausible that the monzonites are instead associated with masses of more mafic rocks. It cannot be resolved for sure whether the mafic rocks are Permian or Precambrian in age; however, the presence of magnetic anomalies and the association with ringshaped, possible subsidence structures, suggest Permian rocks since the mafic rocks of the neighboring Precambrian terrain usually have but negligible magnetic anomalies.

c. The Kongsberg high

The Bouguer map (Fig. 60) shows that the center of the gravity high is situated directly above the erosional contact between the inward (eastward) tilting Cambro-Silurian sediments of the Oslo Region and the Precambrian rocks of the Kongsberg area. The position and apparent extension subparallel to the contact (Plate 5) initiated a closer study of this occurrence.

The center of the anomaly is found in an area partly covered by extensive Quaternary deposits. To the north-east, the Precambrian rocks in the triangle Kongsberg–Krekling–Skollenborg have been studied by Bugge (1917) and S. Jacobsen (pers. comm. 1974). Dioritic gneiss and amphibolite occur interfolded with granitic and various quartz-dioritic gneisses. The dioritic gneisses and amphibolites (layered complex?), together with later, small intrusions of Vinor diabase and gabbro, cover the area close to the gravity anomaly center, but also occur in wide areas outside this region (Fig. 61). Field checks in the anomaly region revealed that the commonly occurring plagioclase-hornblende-(quartz-garnet) gneisses gradually change into gabbroic rocks with remnants of igneous texture and with heavily uralitized pyroxene towards the central parts. Density measurements of rocks from the core and the margin failed to prove the existence of any significant density differences, although some amphibolitic rocks in peripheral zones show somewhat lower density (average of two samples: $\varrho = 2.89$ g/cm^3). The dioritic gneisses have an average density of about $\varrho = 3.00$ g/cm^3 (st. dev. = 0.05, range: 2.916 – 3.087, 18 samples),

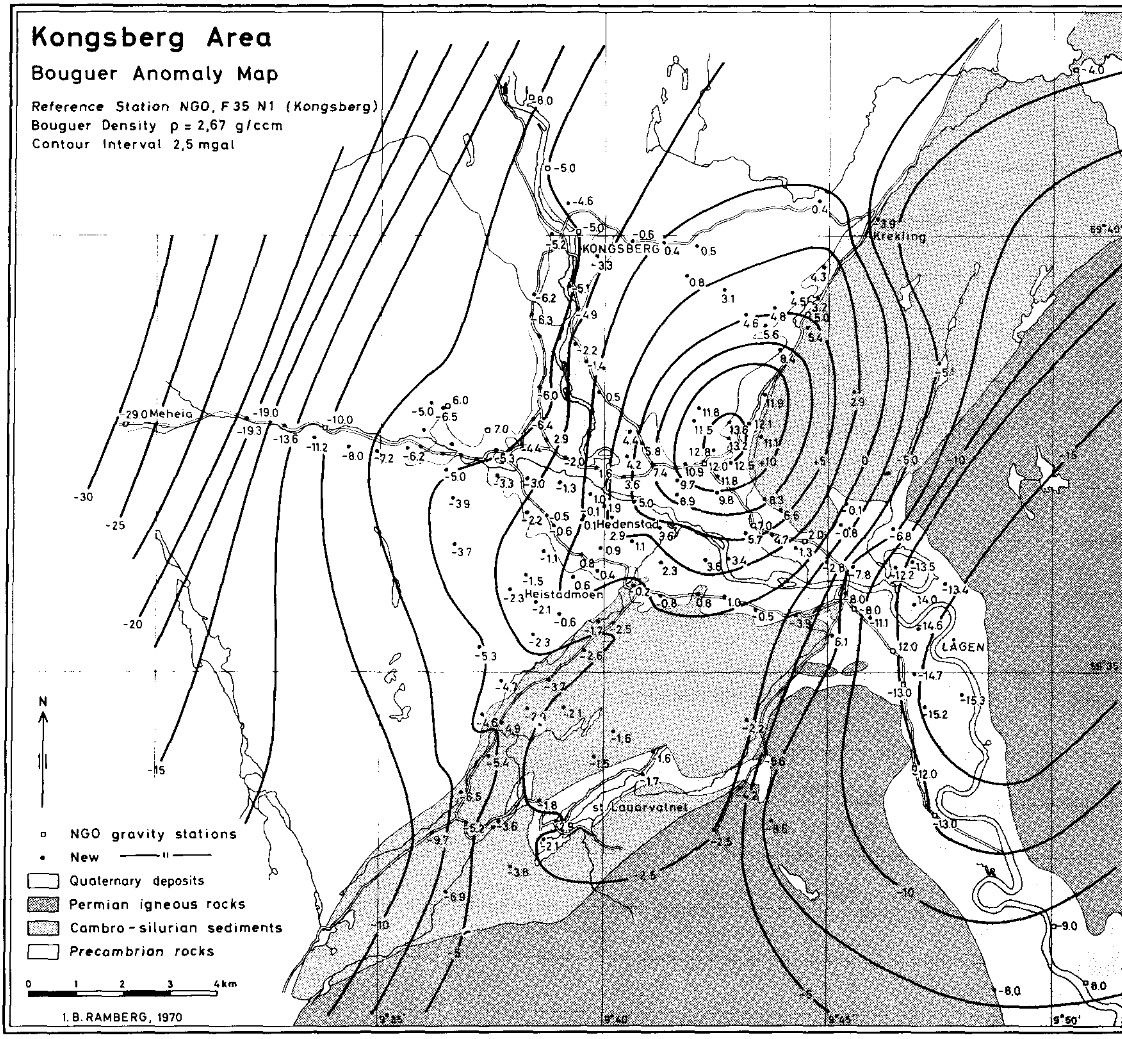


Fig. 60. Detailed Bouguer anomaly map of the Kongsberg area.

Thus an increased thickness of the uralite-gabbro in the central part seems a logical cause of the Kongsberg anomaly.

Fig. 62 presents a simple three-dimensional model. Assuming a slightly cone-shaped body and a density contrast of 0.25 g/cm^3, the body has to extend to a depth of about 7.5 km. If we consider the overall high-density country rock, a smaller density contrast of 0.2 g/cm^3 seems reasonable and will lead to a vertical extension of about 15 km to match the residual peak anomaly. The local occurrence and the sub-circular anomaly pattern suggest the existence of some kind of feeder zone for the originally intrusive amphibolites and dioritic gneisses, the uralite-gabbro probably representing the less deformed central part of the postulated pipe.

Although presumably caused by Precambrian rather than Permian intrusive

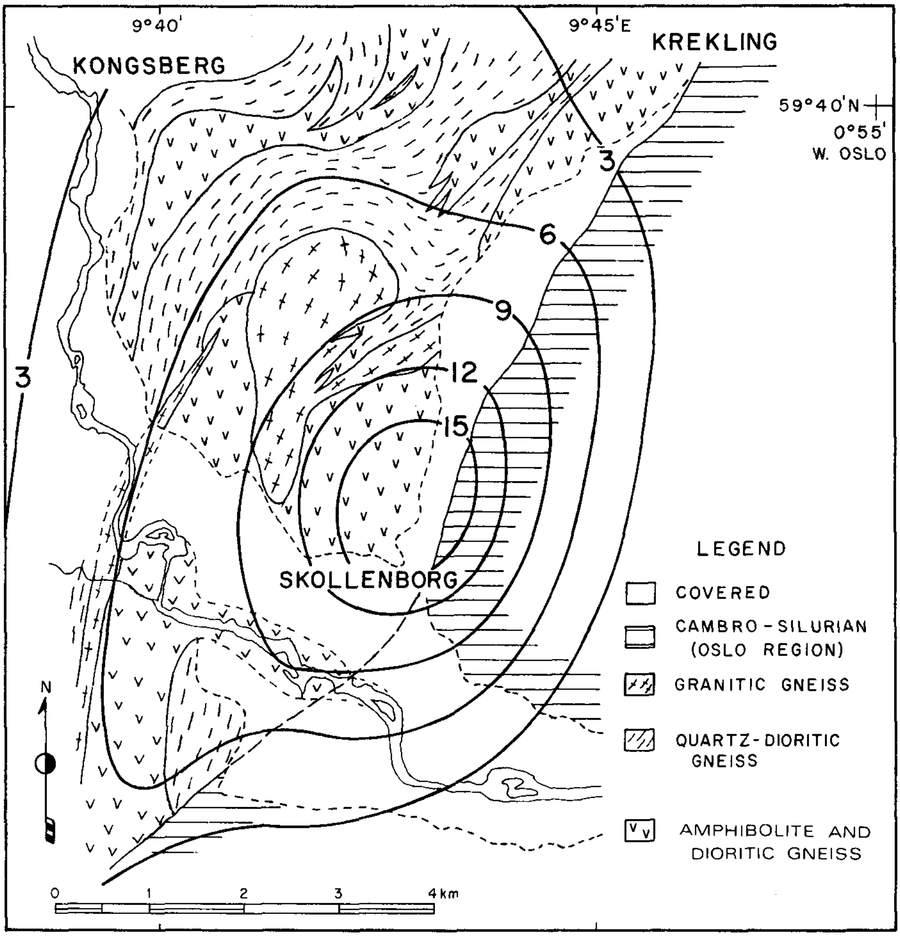


Fig. 61. Residual anomaly of the Kongsberg area after removal of the regional anomaly (Plate 4) from the observed field (Fig. 60) and bedrock map of the Precambrian terrain west of the Oslo Region.

rocks, the existence of the Kongsberg occurrence is of interest in the present context. Because of its location exactly on a straight line through the 'Oslo-essexites' Aurenhaugen, Dignes, Sønstebyflakene, Snaukollen and Eiangen, it is implied that the Precambrian zone of weakness also served as an avenue of ascent for the Permian intrusives. By extrapolating the zone further towards the southwest, it lines up with the alkaline Fen complex of assumed Eocambrian age (Ramberg & Barth 1966), late Carboniferous (?), ultrabasic intrusions along the 'Great friction breccia' (Barth 1970, Touret 1970) and the possible volcanic centers (Permian and Tertiary) in the Skagerrak outside Kristiansand (Åm 1973).

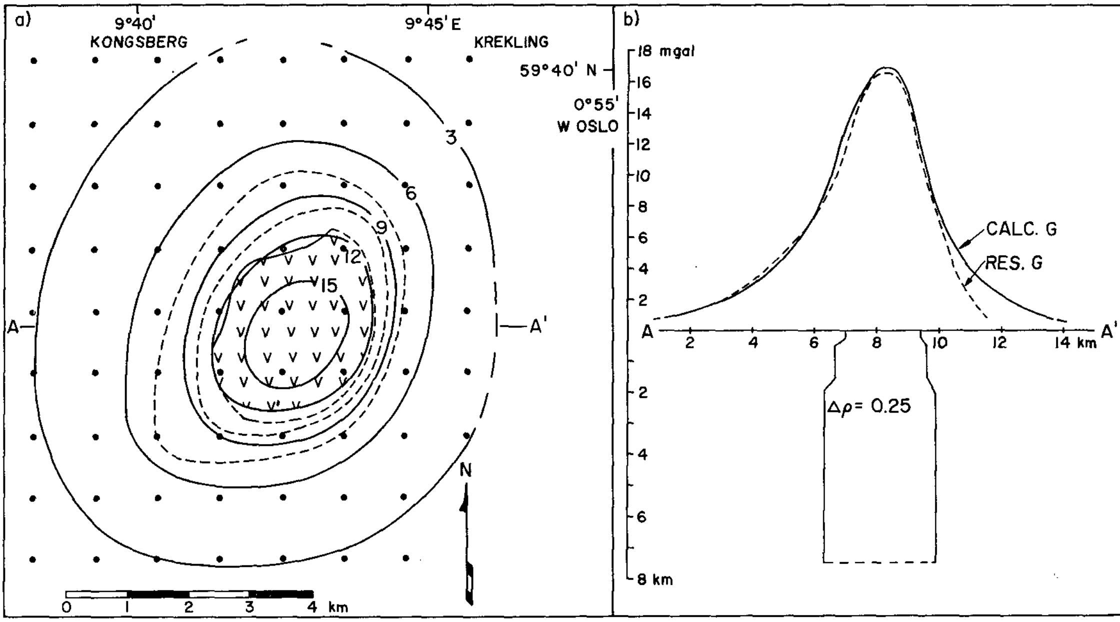


Fig. 62. Kongsberg gravity high, three-dimensional model and calculated gravity, (a) Map projection. Calculated g (solid lines), depth contours (dashed lines), black dots (positions of calc. g. values), (b) E–W profile.

d. The Tyrifjord high

Like the Kongsberg gravity high, the Tyrifjord high seems to be located exactly along the western border of the Oslo Region. It is elongated NNE–SSW along the border and rises to a residual high of about 31 mgal. In its south-eastern part it is superimposed by the subcircular Finnemarka low (see p. 74).

The Precambrian area west of Tyrifjorden (the Holleia area) contains no Permian rocks except some N–S striking rhomb-porphyry dikes. The area has long attracted some interest (Lassen 1876, Vogt 1893) because of occurrences of nickeliferous ore, often associated with mafic to ultramafic rocks. The area is dominated by amphibolites and a variety of banded gneisses and migmatites (Hofseth 1942). Aker-Johannesen (pers. comm. 1972) shows that the mafic to ultramafic bodies occur throughout the area in a number of separate outcrops, sometimes down to a few hundred meters across or less. One is left with the impression that the entire gneissic terrain might be underlain by a larger mafic body of which only parts of the irregular roof are exposed on the present-day surface. The variety of gabbroic rocks in the Holleia area is considered a differentiation series from a common parental source (Hofseth 1942).

The large gravity anomaly in the area supports the idea of the presence of substantial amounts of subsurface, dense rocks. None of the gravity stations included in the Bouguer anomaly map is located on any of the gabbroic outcrops, so there are no strictly local anomalies contributing to the extensive

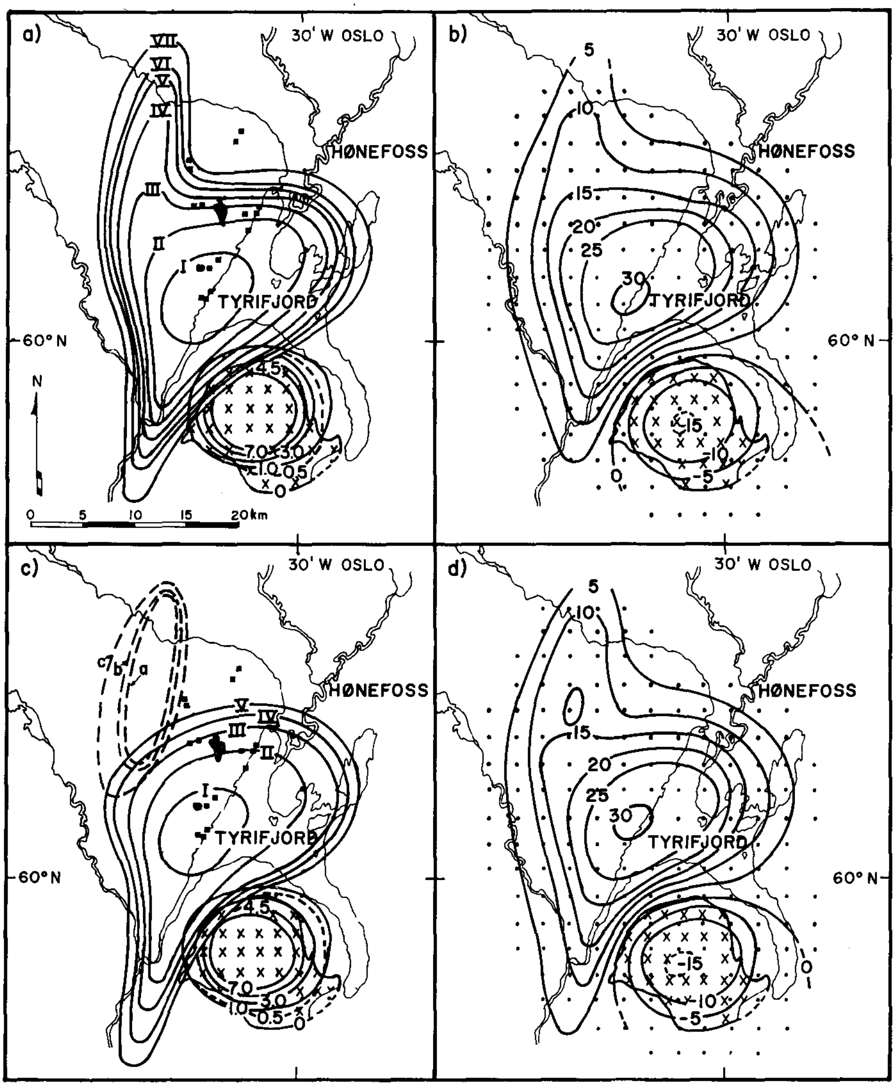


Fig. 63. Three-dimensional models cf the Tyrifjord gravity high. (a) Interpreting the high as caused by a single dense body. For a density contrast of $\Delta\varrho = +0.2$ g/cm^3, the contour lines represent the following depth levels (in km.): I = 0.5, II = 1.0, III = 1.5, IV = 3.0, V = 4,0, VI = 5.5, VII = 6.0. For $\Delta\varrho = 0.3$ g/cm^3 : I = 0.7 and VII = 4.0. For $\Delta\varrho = 0.1$: I = 0.2, and VII = 11.0; however, in the last case the gradients are too gentle compared with the residual. (b) Calculated gravity anomalous for $\Delta\varrho$ = 0.2 g/cm^3. (c) Alternative model with two separated anomalous bodies. Depth contours for $\Delta\varrho = 0.2$ g/cm^3 : I = 0.5, II = 1.0, III = 3.0, IV = 4.5, V = 6.0 km (main body), and a = 0.5, b = 1.5, c = 2.5 km (for second body). (d) Calculated gravity for the alternative model. In both cases the mass deficiency of the nearby Finnemarka granite ($\Delta\varrho = -0.13$ g/cm^3) was included. Exposed granite is shown by crosses; syenodioritic rim is not shown. Outcrops of mafic to ultramafic intrusives in the Holleia area indicated with irregular black spots (sizes of small occurrences exaggerated). Small black dots indicate positions of calculated gravity values (b & d). Number of lamellae used when calculating theoretical anomalies: (a) Tyrifjord dense body (12), Finnemarka granite (14); (c) Main dense body (13), second dense body (4), Finnemarka granite (22).

Tyrifjord residual high (Plate 5). Fig. 63 shows some simple three-dimensional models of the hypothetical body. By assuming a depth of 0.5 km to the top surface (for the sake of simplicity taken to be a horizontal plane), a roughly 15 x 20 km wide body will have a thickness of about 6 or 4 km with a density contrast of 0.2 or 0.3 g/cm^3, respectively. Since the average density of the gneisses and amphibolites in this part of the Precambrian terrain is expected to be higher than 2.80 g/cm^3 a largely gabbroic rock will probably exhibit a density contrast of 0.2–0.25 g/cm^3. If the anomalous mass is not represented by a compact, dense (gabbroic) body but by a lighter, bulky body composed of a number of basic intrusives, a smaller density contrast ($\sim$0.1 g/cm^3) might be more appropriate. The thickness of such a body will be accordingly greater.

A variety of model solutions are possible, but the anomalous mass estimate is not ambiguous (Table 14). The numerous outcropping gabbroic bodies and the maximum depth estimates (Table 13) suggest that the depth to the top of the anomaly-causing body is not very great.

No detailed measurements have been carried out in this area, but considering the economic potentials of the rock types involved this seems a worthwhile project. Also, more detailed gravity measurements are likely to result in a subdivision of the anomalous masses into at least one elongate western part associated with an area of exposed hyperitic rocks and a more oval eastern part beneath the gabbros and ultrabasics of Holleia (Fig. 63c).

The present importance of the Tyrifjord occurrence seems to be the same as for the Kongsberg high; Precambrian intrusive rocks occur right at the western border zone of the Oslo Region. The Precambrian age of the dense masses is indirectly supported by the aeromagnetic measurements of NGU. Neither the Tyrifjord nor the Kongsberg gravity high shows up in the aeromagnetic maps in contrast to, e.g., the 'Oslo-essexites' and the Permian basic extrusives.

e. The Asker-Lier high

This high is found above the Cambro-Silurian rocks and Permian lavas in the area between the Drammen and Finnemarka granite complexes. On the map (Plate 2) it is divided into two highs, the Lier high to the west and the Asker high to the east, both reaching residual peak values of about 12 mgal. More recent measurements (Ramberg 1972a, Table XIX) show that the two highs in effect form one continuous high divided only by a relatively narrow gravity trough of about -2 to -4 mgal along the Lier valley.

The nature of the Asker–Lier anomalous mass is not revealed at the surface. Except for the thin B1 basalt flow and the commonly occurring diabase dikes (and other rarer basic dikes), no rocks denser than the Cambro-Silurian are known in this area. No subsurface intrusive centers are indicated by the general course of the contact metamorphic zones as shown, e.g., on the geological map of Kristiania by Brøgger & Schetelig (1917). To the south, west, northwest, and partly also to the east, the gravity high is flanked by gravity lows. This

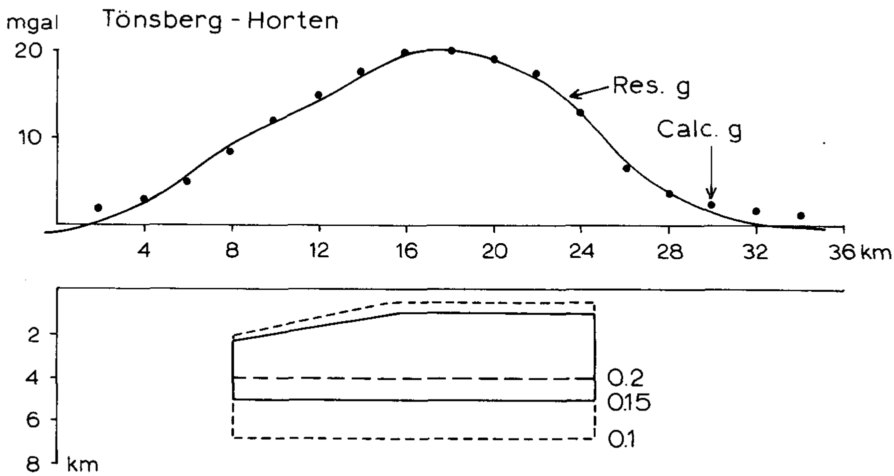


Fig. 64. Horten-Tønsberg gravity high, residual gravity profile (Plate 5), and geophysical models.

makes the real outline of the anomaly and the gradients due to the mass surplus hard to identify.

On the basis of general geological evidence, it seems reasonable to assume that the dense body is some kind of gabbroic rock. If we assume a density contrast of about 0.2 g/cm^3 and a depth to the top surface of about 0.5 km, a two-dimensional body with a width of about 5 km has to be about 2.5 km thick to fit the residual anomaly of Plate 5. This example shows that a considerable subsurface mass excess has to be located in the Asker–Lier area. Estimates are given in Table 14.

Despite the uncertainty, two principally different suggestions as to the nature of the anomaly are presented here. The mass anomaly may be due to: (1) local intrusions of Permian basic rocks, or (2) dense rocks of Precambrian age in the underlying basement. As with the Tyrifjord and Kongsberg gravity highs, no accompanying aeromagnetic anomaly is apparent. The Tyrifjord-type rocks can be continuous south-eastward into the Asker–Lier area, forming a high density basement belt. However, since Permian intrusives characterized by low-intensity aeromagnetic anomalies do occasionally occur in the Oslo Region (Åm & Oftedahl, in prep.) and as no extensive high-density rocks have been recognized in the Precambrian outcropping SE of Asker, the first alternative is tentatively held as the most likely one.

f. The Horten–Tønsberg high

This gravity high runs along the west side of the Oslofjord from south of the Drammen granite to beyond Tønsberg and possibly to Ferder and further southwestward (Plate 2). The anomaly covers a number of contrasted surface rocks such as Permian extrusives and intrusives (larvikite-tønsbergite), Downtonian and Silurian sedimentary rocks, as well as Precambrian gneisses. It seems that the anomaly must inevitably be caused by subsurface rocks, these being situated at relatively shallow depths (Table 13).

A WSW–ENE profile across the center of the residual anomaly reveals a peak value of about 20 mgal (Fig. 64). The various two-dimensional models displayed indicate that with the top surface at 1 km depth, the minimum plausible density contrast has to be found within the range 0.1–0.15 g/cm^3. With a thicker overburden the density contrast has to be still higher.

Regarding the nature of the anomaly-causing mass surplus, the same two alternatives exist as discussed in connection with the Asker-Lier high: dense Precambrian basement rocks or Permian basic intrusives. In addition, the possible occurrence of subsided, dense surface rocks such as basalt has been discussed but rejected (Ramberg & Smithson 1971). One reason is that the bottom (B1) basalt flow and underlying sandstones are exposed in wide areas, and no large-scale subsidence of these rocks seems to have occurred. Another reason is that aeromagnetic anomalies indicate a subcircular outline for the Larvik batholith (Sellevoll & Aalstad, 1971) and that the Vestfold lava region is probably not underlain by extensive masses of larvikite, as commonly believed, but by Precambrian basement rocks and largely non-magnetic Permian igneous rocks (*Åm & Oftedahl, in prep.*).

The metasomatic transformation of larvikite into tønsbergite has been discussed by Raade (1973, p. 64) who suggests as one possible alternative that "the existence of a hidden, granitic rock beneath the tønsbergite might well explain the exsolution of iron oxides in the tønsbergite feldspars as a result of heating." The tønsbergite is found directly below the southern part of the Horten–Tønsberg gravity high. It seems that both the metasomatism and the gravity high have a related cause; the hot intrusives would then necessarily have to be denser than granitic (or monzonitic) rocks.

The preferred interpretation is that the Horten–Tønsberg gravity high is caused by Permian basic intrusives forming a thick sill-like body at relatively shallow levels and at least 50 km in length along the Oslofjord region. The anomaly, as possibly the case for the Narrefjell and Asker–Lier highs, may therefore be indicative of the existence of large, subsurface basic intrusives not only at deeper levels in the Oslo Region (Chapt. 7) but also locally at relatively shallow depths. If so, the 'basaltic' body must be older than the Drammen granite against which the anomaly is abruptly terminated, but younger than the larvikite that surrounds it. It must also be later than the basalts around Horten since these are cut by the larvikite.

g. Concluding remarks

Mass and volume estimates of the different dense rock bodies considered above are given in Table 14. Since the density contrasts of the dense bodies may be assumed always to be greater than zero, it is believed that the anomalous mass estimates are dependable. All of the gravity residuals were easily resolved except perhaps the Asker–Lier high which overlaps with several gravity lows.

Table 14: Anomalous mass, total mass, and volume estimates of dense rock bodies

Occurrence	Anom. Mass, ΔM 10^{17}g	Applied dens. contr. g/cm^3	Total Mass, M 10^{17}g	Volume, V 10^{17} cm^3
Skien basalts	0.42	0.16	7.5	2.59
Narrefjell	0.69	0.15	13.1	4.57
Kongsberg	0.16	0.20	1.9	.64
Tyrifjord	2.85	0.20	41.9	14.26
Asker-Lier	0.38	0.15	7.2	2.50
Horten-Tønsberg	3.62	0.15	69.8	24.15

1 km^3 = 10^{15} cm^3

It seems likely that the Kongsberg and the Tyrifjord highs are caused by intrusive rocks of Precambrian age. Nevertheless, the markedly NNE-elongated gravity anomaly pattern and their location along the western border of the Oslo Region along with Permian and other intrusive rocks are considered to have a tectonic significance. The western border of the Oslo Region and its extension along the 'Great Friction Breccia' down the coast seems to have been a zone of tectono-magmatic activity a long time before the graben formation took place.

The nature of the Narrefjell mass surplus is uncertain. It may in part represent a NNW continuation of the subsided Skien–Porsgrunn basalts and in part denser intrusives (?) in a wide area below the Skrim monzonites. Similarly, the Asker–Lier and the Horten–Tønsberg highs are tentatively ascribed to the presence of shallow, basic, Permian intrusives, but the existence of dense Precambrian basement rocks cannot be excluded.

7. Gravity models of the regional anomalies

7.1 POSSIBLE STRUCTURAL SOLUTIONS

The regional (II), as shown in Plate 4, does not reflect the sometimes rather irregular and accidental borders of the province, but marks the Oslo Region only in a general way. This, together with the smooth appearance of the isogals, is in harmony with a deep-seated source for the regional variations, but may computationally just as well be due to a wide, thin body at more shallow depths. Since no anomalously high upper mantle velocities have been reported, the regional variations will include gravity effects by Moho undulations and by possible density anomalies in the crust below the level of the more superficial felsic rocks.

In theory, a whole series of different structures may all satisfy a gravity anomaly like the broad Oslo high (Fig. 65). Alternatives (a) and (b) in Fig. 65 are not very likely on geological grounds: the surface geology is well-exposed, and no extensive field of basic extrusives, as shown in alternative (a), is known. Such a basin would have to cover an area more than twice as wide as the

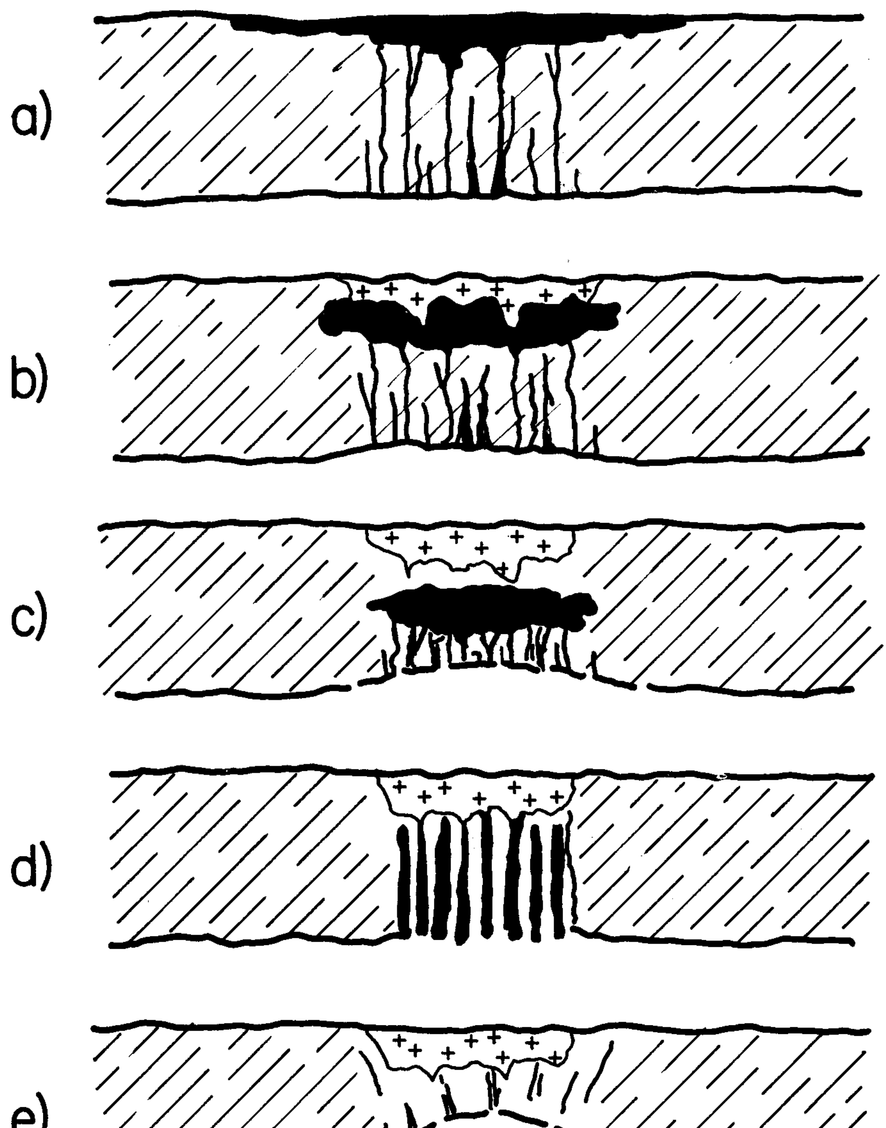


Fig. 65. Possible causes of the Oslo Region gravity high. Mafic igneous rocks (black), felsic igneous rocks (crosses), Precambrian crustal rocks (obliquely hatched) on top of upper mantle rocks (no symbol). (a) mafic extrusives or subvolcanic rocks, (b) shallow mafic intrusives, (c) mafic intrusives at intermediate depths, (d) axial dikes, (e) Moho upward, (f) deep crustal intrusives on top of moderate Moho upwarp. Discussion, see text.

graben (100–120 km). Further, the felsic rock series is almost exclusively younger than the basic intrusives of the area. A situation like (b), with a wide layer of subsurface basic intrusives more or less completely roofed by the felsic rocks, would most likely be reflected in the types of inclusions found. If anything, the inclusions in the felsic rocks, and especially in the northern half of the region, indicate either a monzonite (or syenodiorite) substratum or preexisting monzonite intrusions at the same depth level as now occupied by the felsic rocks.

The remaining alternatives, (c) to (f) of Fig. 65, are all geologically feasible, but no surface geological criteria exist that can discriminate between them. From the Trøndelag–Oslofjord seismic refraction profile (Kanestrøm & Haugland 1971) and crust-mantle models based on detailed seismic studies in the NORSAR area (Kanestrøm 1969, 1973), crustal thinning is indicated for the parts of the Oslo Region covered by these studies. Preliminary reports on detailed refraction studies in the central part of the graben (Sellevoll 1972) confirm earlier suggestions (Ramberg & Smithson 1971, Ramberg 1971, 1972b) of a dense crustal body with $V_p = 7.5–7.6$ km/sec at and below a depth of approximately 22–23 km.

In the gravity studies, estimates of depth to the top of the assumed, dense body were based on maximum depth formulae. For two-dimensional mass distributions and compact masses the formula $H \leq 0.65 \cdot \frac{A_{max}}{\Delta G_{max}}$ (Bott & Smith 1958) can be applied (h = depth to top of body; A_{max} = max. grav. anom.; ΔG_{max} = max. grav. gradient). ΔG can be read from the difference between the regional I and regional II in Figs. 66–73 or from the flanks of the constructed residual in Fig. 76. The maximum gradient is in the range 0.7–0.9 mgal/km equivalent to a maximum depth estimate in the range of approximately 22 to 26 km. Since maximum depth formulae generally give too deep estimates, the smaller value (22 km, or less) is regarded as the most realistic estimate.

Hence, as a first approximation, a hypothetical structure like the one depicted in Fig. 65f is favored. It appears to satisfy the various geophysical and geological data at hand, and has consequently formed the basis for most of the model computations; however, the alternatives (c) or (d) should also be considered.

7.2 GRAVITY PROFILES

The 15 profiles (nos. 0 to XIV, Figs. 66–73) shown in Fig. 20 are drawn perpendicular to a generalized graben axis. They are spaced at regular intervals 25 km apart, except for profiles 0 and XIV which are 50 km to the north and to the south of the nearest profiles, respectively (Plate 1). The three northernmost profiles (0–II) cross north of the exposed graben. The three southernmost (XII–XIV) cross the southeastern part of the Norwegian Chan-

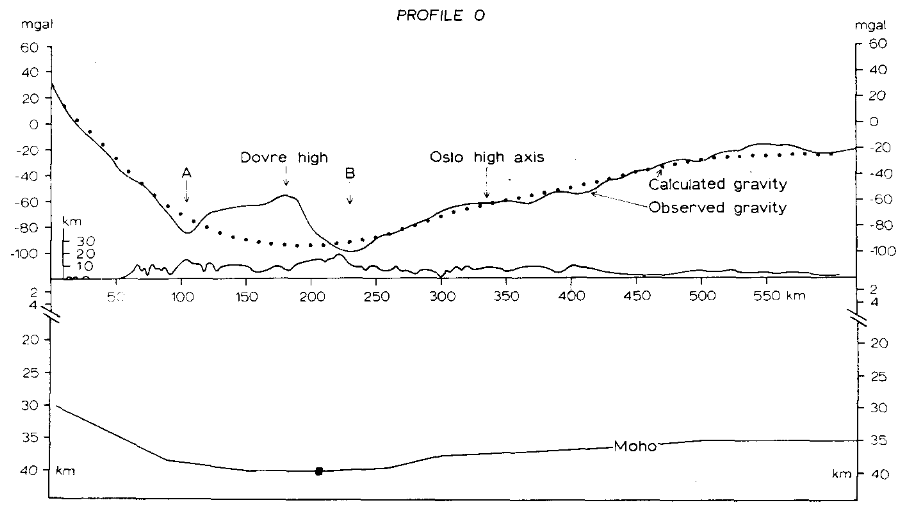


Fig. 66. Gravity profile 0 (for location see Plate 1) and crustal model. A and B: local lows on both sides of the Dovre high. ■: Seismic Moho depth along the Trøndelag – Oslofjord profile (Kanestrøm & Haugland 1971). Note that at this distance from the Oslo Region (about 100 km) hardly any gravity high is detected; the gravity field may be explained by variation in crustal thickness only. Crust/mantle density contrast applied: 0.45 g/cm^3.

nel, a pronounced submarine trough (Holtedahl 1960, 1963) in the SE extension of the Oslo Graben, and will be discussed in more detail at the end of this chapter.

For each profile, gravity models have been computed to satisfy the two regionals (I and II) which were drawn according to the methods previously described (Chapter 5). To the north of the Oslo Region (Figs. 66–67), both methods coincide, implying that the Oslo gravity high in profiles I and II is primarily due to an axial Moho upwarp, while in profile 0, some 100 km NNE of the graben, the Oslo gravity high is no longer visible. Here also, the regional anomalies closely coincide with the observed Bouguer anomaly for most of the profile lengths. The one exception, the 'Dovre high', occurs at the SW tip of the Trondheim Region in the extension of the 'Faltungsgraben'. With the regional chosen, the gravity lows at A and B (Fig. 66) suggest that the anomalous dense masses causing the 'Dovre high' are in part regionally isostatically compensated. Similarly, the regionals are drawn as smooth curves below the local Jotunheimen and Slidre highs (Figs. 67–69) and the Hardanger and Telemark highs in other profiles, etc. In profiles I and II Fig. 67, the broad gravity low (of about –20 to 25 mgal) over the Dala sandstone and Filipstad granite, consistent with a depth of about 5 km for a body with a density contrast of about –0.1 g/cm^3, has been compensated for. To the south (Figs. 70–71) similar lows are associated with the Østfold granite close to the graben (Ramberg & Smithson 1971) and above gneissic granites in Sweden (Åmål–Kroppefjäll granites) just east of the international border (Magnusson 1962).

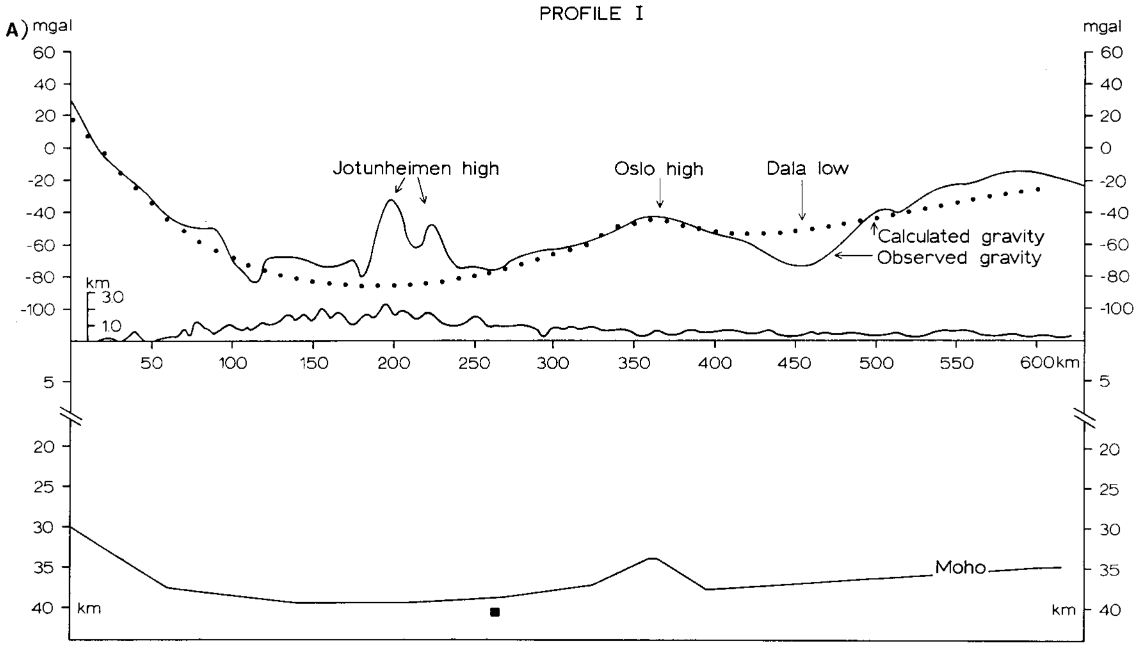
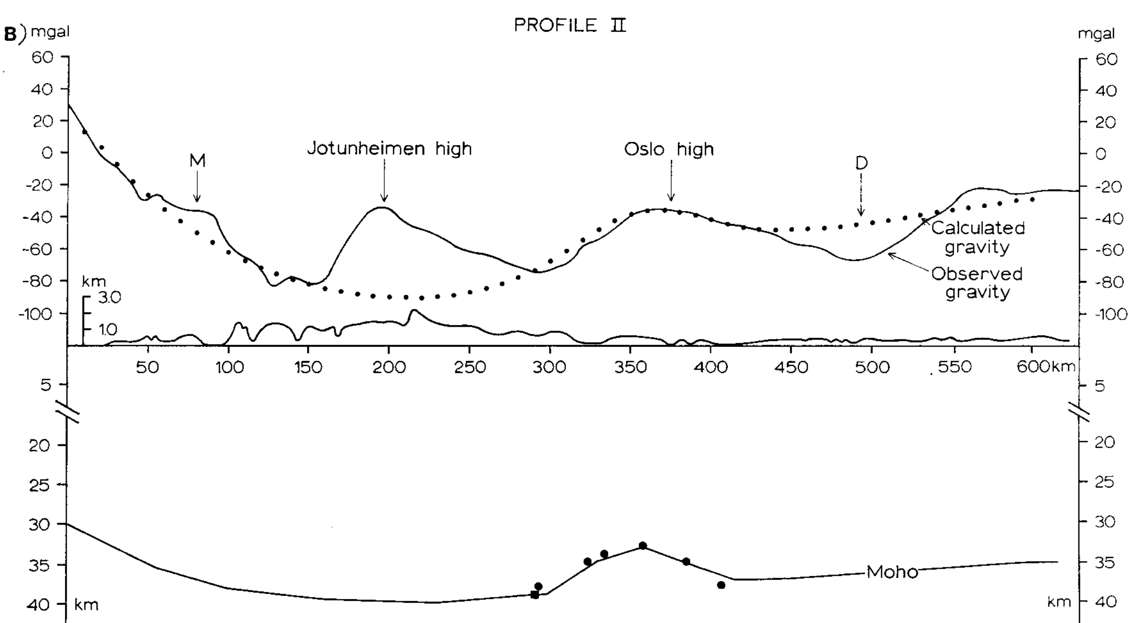


Fig. 67. (a) Gravity profile I and crustal model. Note that seismic Moho depth, ■, is below the chosen depth since the gravity profile crosses the Trøndelag – Oslofjord seismic profile in an apparent local increase of crustal depth. (b) Gravity profile II and crustal model. M = Møre gravity high (ultramafics?), D = Dala low (granite and sandstones). ●: Seismic Moho depth under the NORSAR (Kanestrøm 1973). Crust/mantle density contrast, $\Delta\varrho = 0.45$ g/cm³.



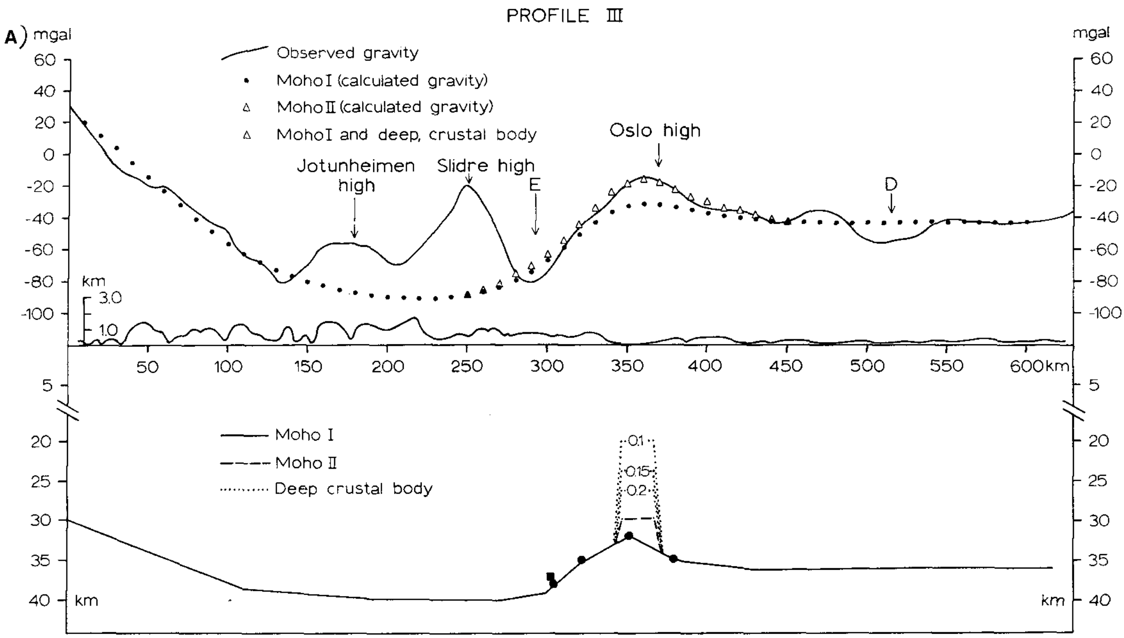
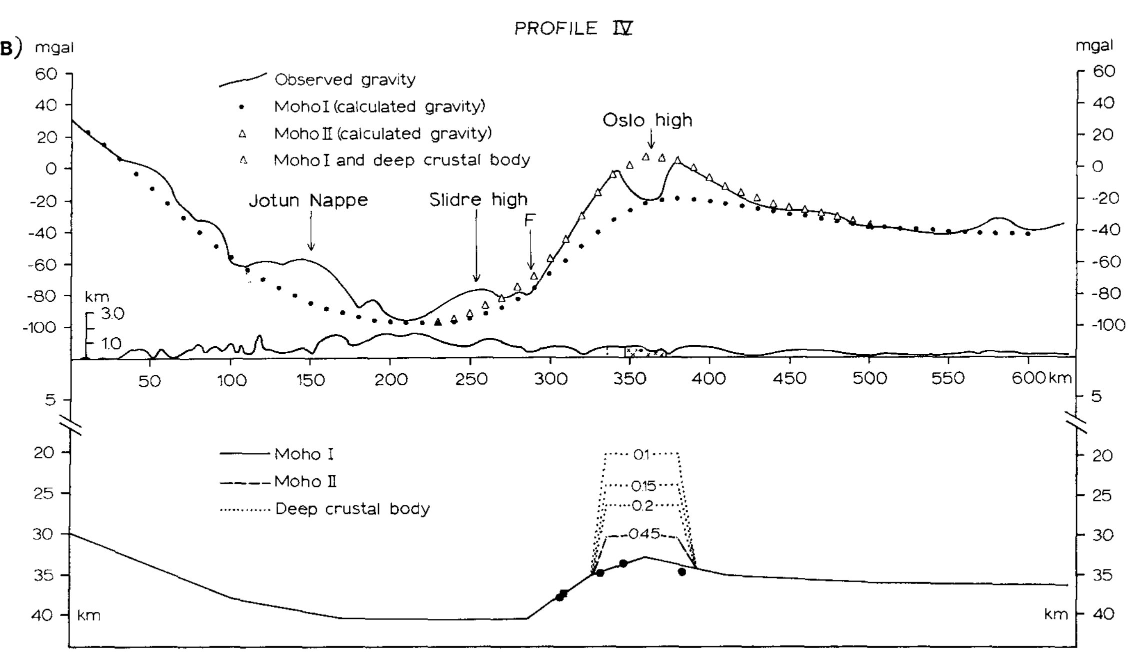


Fig. 68. (a) Gravity profile III and crustal models. E = Etnedal gravity low. Other symbols as in Fig. 67. (b) Gravity profile IV and crustal models. F = Flå gravity low (granite). Other symbols as in Fig. 67. Oslo Region felsic intrusives (X'es).



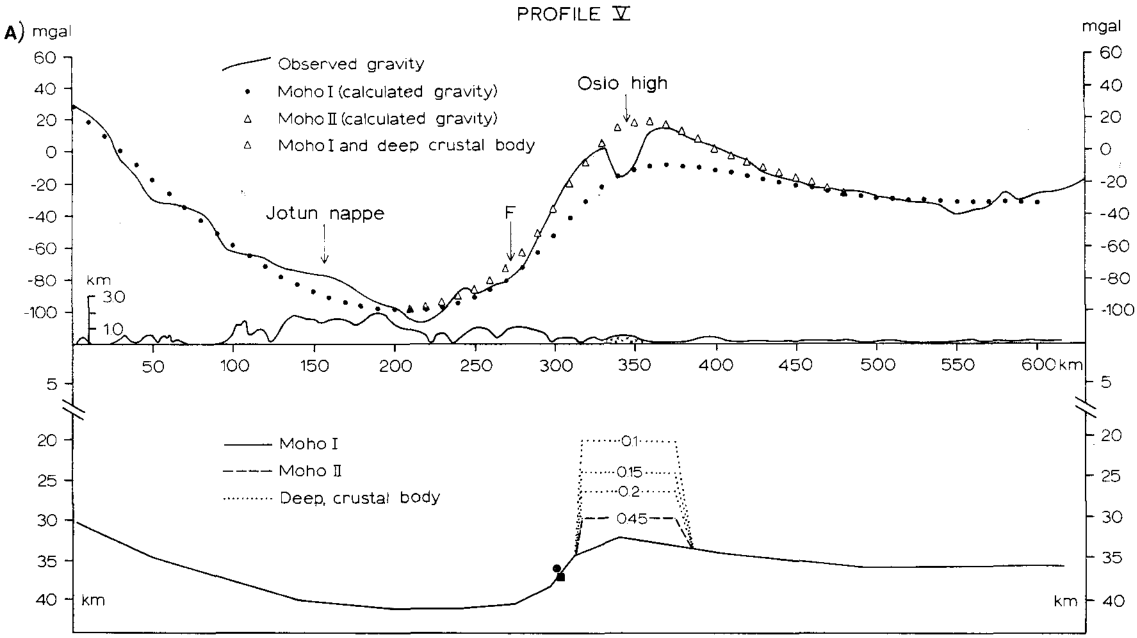
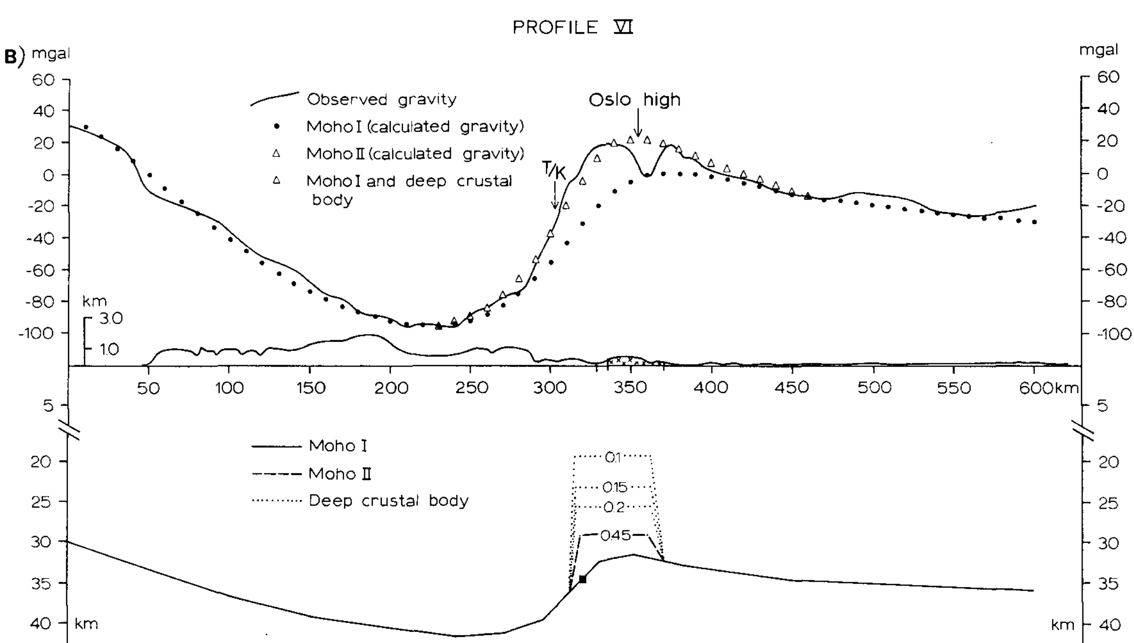


Fig. 69. (a) Gravity profile V and crustal models. (b) Gravity profile VI and crustal models. T/K = Telemark/Kongsberg contact zone. Other symbols, etc. see Figs. 67–68.



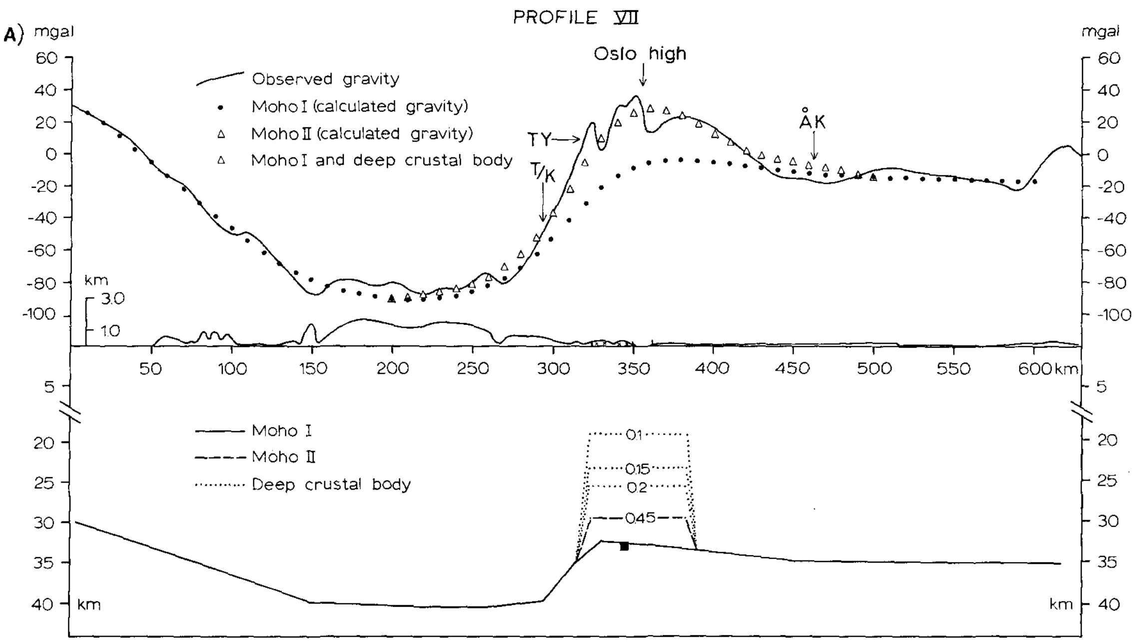
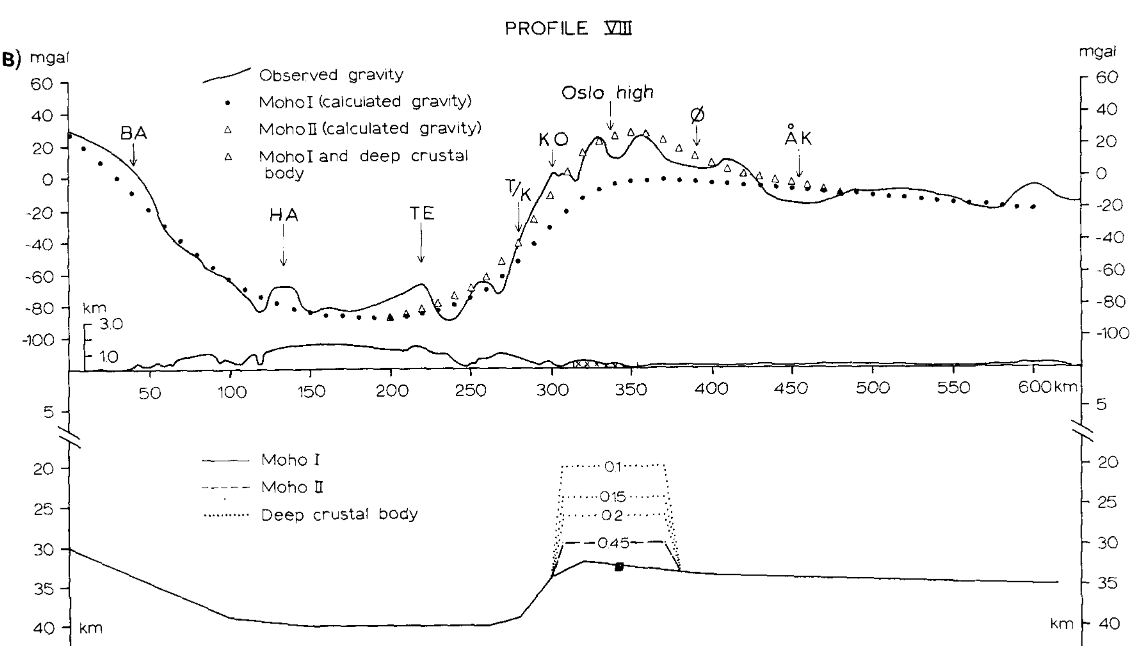


Fig. 70. (a) Gravity profile VII and crustal models. TY = Tyrifjord gravity high (mafic rocks), ÅK = Åmål-Kroppefjell low (granitic rocks). (b) Gravity profile VII and crustal models. BA = Bergen arcs, HA = Hardanger high (Precambrian gabbroic rocks), TE = Telemark high (basic volcanics?) KO = Kongsberg high (mafic rocks), Ø = Østfold low (granite). Other symbols, etc., see Figs. 67–69.



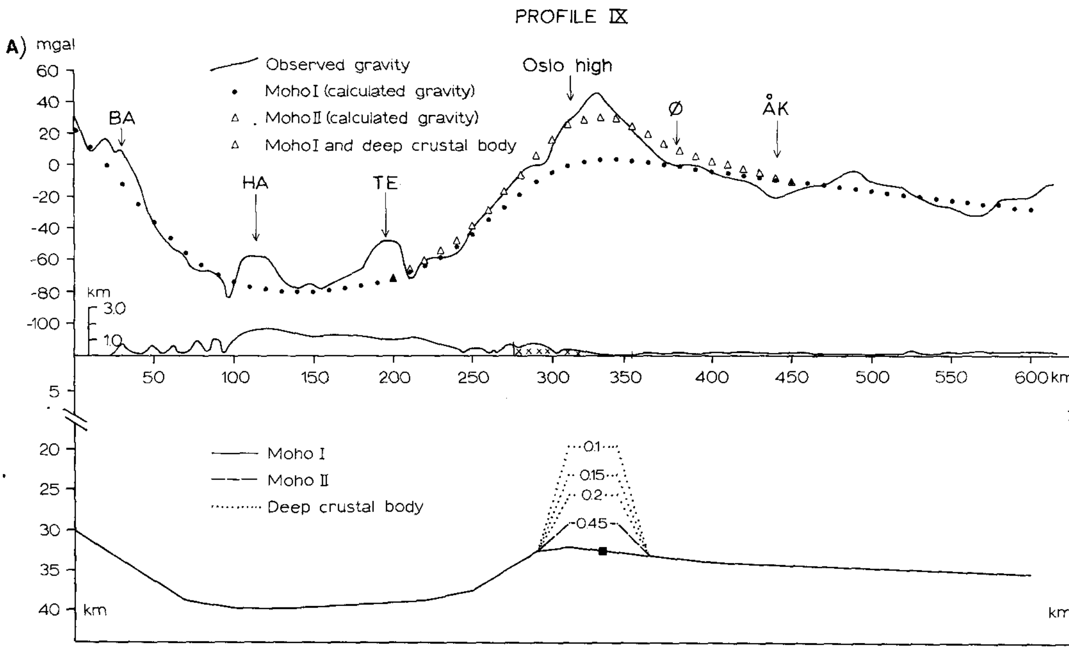
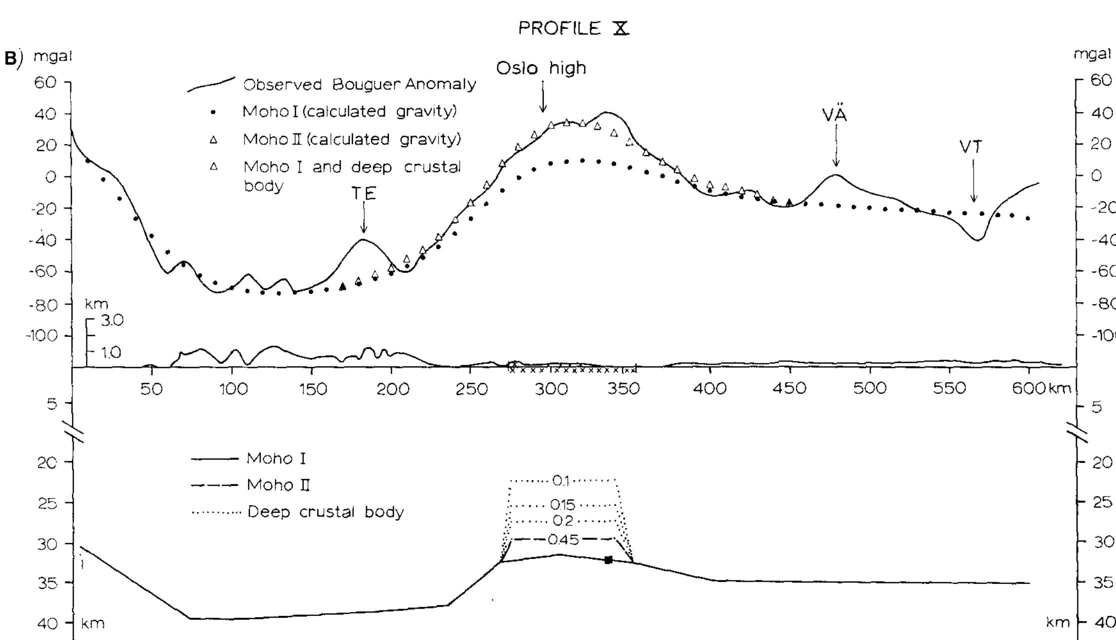


Fig. 71. (a) Gravity profile IX and crustal models. (b) Gravity profile X and crustal models. VÄ = Vänern high (mafic rocks), VT = Vättern low (graben). Other symbols, etc. see Figs. 67–70.



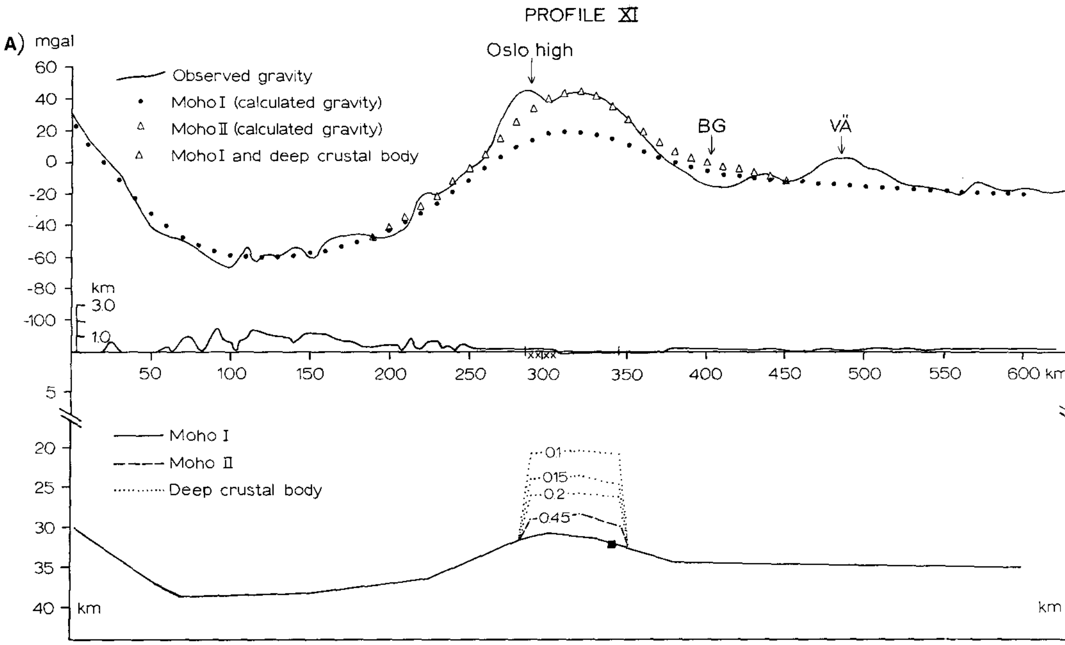
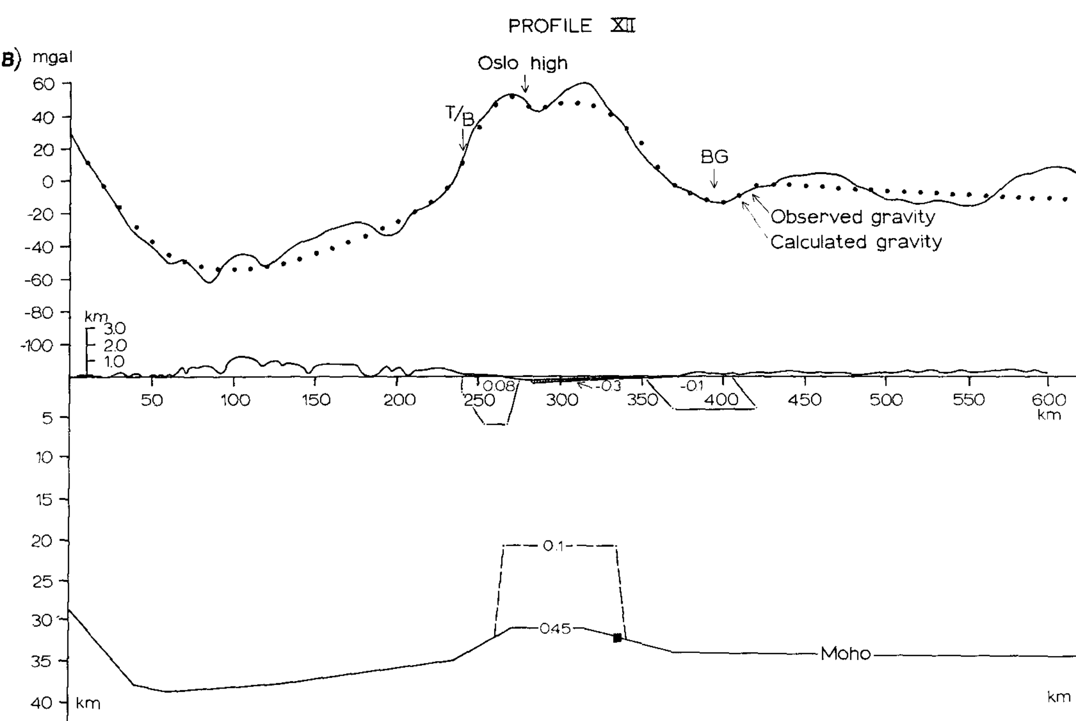


Fig. 72. (a) Gravity profile XI and crustal models. BG = Bohus low (granite), VA = Vänern high. (b) Gravity profile (XII) across the Skagerrak, and crustal model. T/B = Telemark/Bamble contact zone, BG = Bohus low. Density contrasts as indicated on figure. Central gravity peak not accounted for may be due to shallow mafic intrusion.



A)

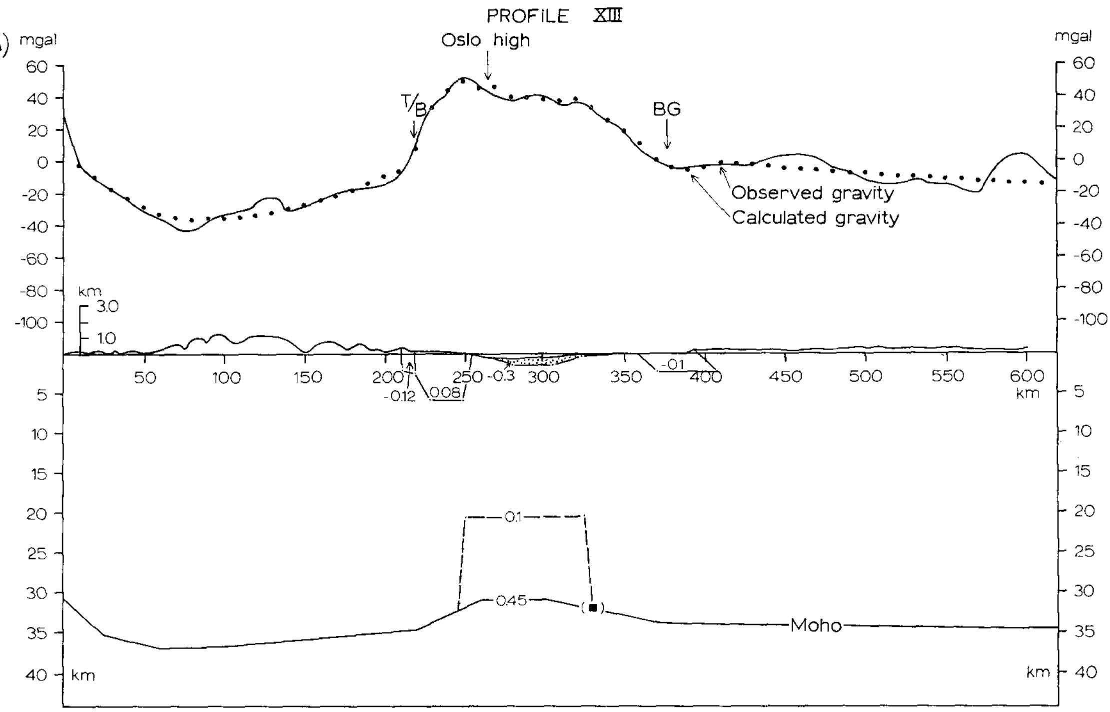
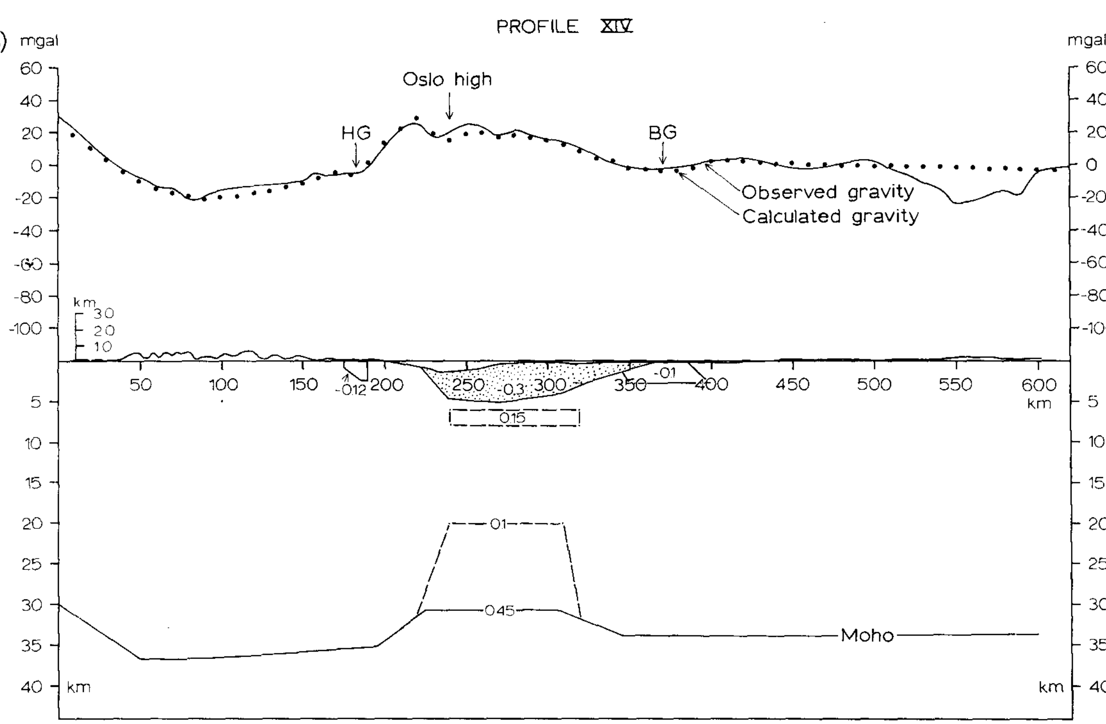


Fig. 73. (a) Gravity profile (XIII) across the Skagerrak, and crustal model. (b) Gravity profile (XIV) across the Skagerrak, and crustal model. HG = Herefoss granite low. Other symbols, see Fig. 72. Density contrasts as indicated on figures.

B)



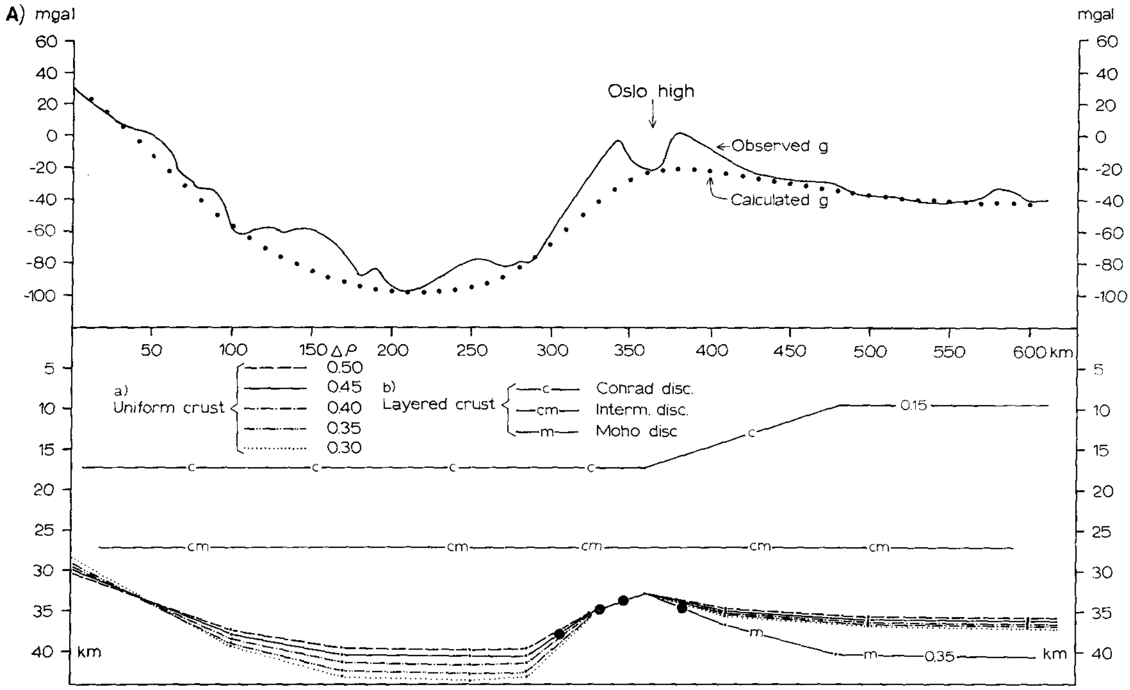
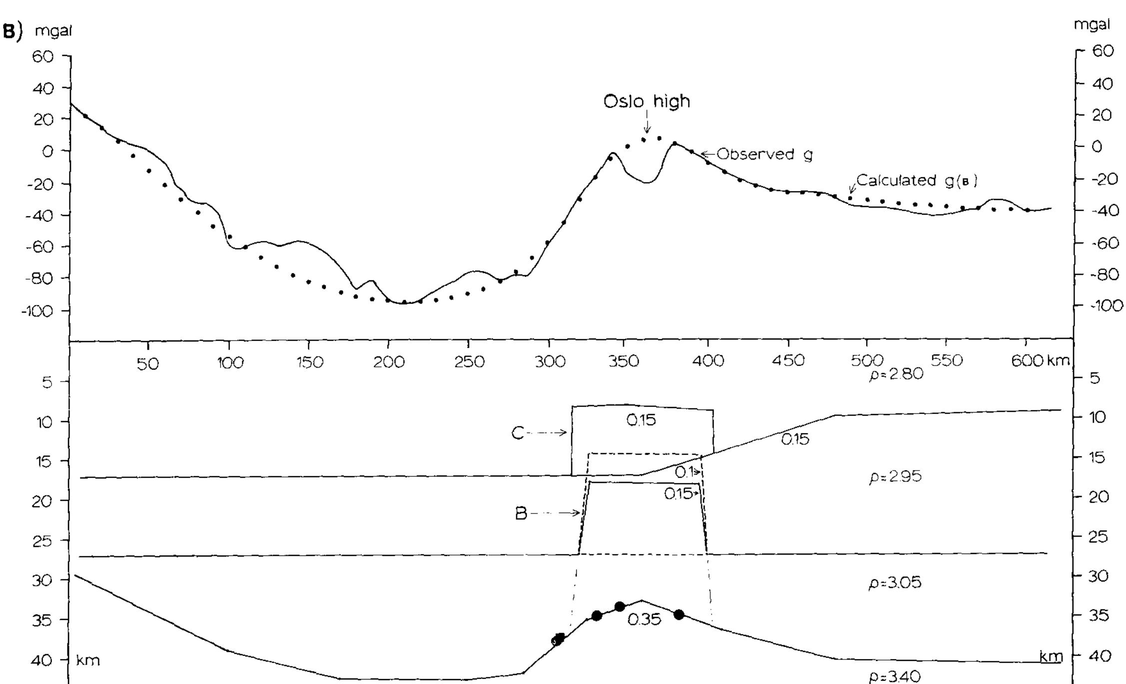


Fig. 74. (a) Alternative crustal models, profile IV. a) Effect of various crust/mantle density contrasts, b) shallowing of depth to Conrad towards east into Sweden according to seismic data (Dahlman 1971) leads to corresponding increase in Moho depths as exemplified by the $\Delta\varrho = 0.35$ g/cm^3 deep crustal/upper mantle density contrast.
(b) Alternative solutions of dense crustal body, profile IV. (A: At deep crustal levels, see Fig. 68), B: At intermediate crustal levels, C: At upper crustal levels (but below felsic intrusives).



In profile V (Fig. 69) the disc-shaped Flå granite (Smithson 1963b) creates a local, steep gradient on the western flank of the Oslo high. Since this feature is not caused solely by the Oslo rocks, the regional (II) is drawn smoothly as shown. The same reasoning applies to profiles in Figs. 69–70, where the contact (marked with an arrow) between the granitic rocks (Flesberg granite) of the Telemark region to the west and the relatively dense rocks of the Kongsberg region to the east forms an abrupt increase in the gravity gradient. An improved gravity coverage in this area would better define the sharpness of the gradient and help clarify the mutual relationships between the Telemark and the Kongsberg rocks. Nevertheless, the general features can be inferred from the existing gravity data and from the apparently rather similar conditions met with along the Bamble/Telemark contact zone further to the southwest (profiles XI to XIV, Figs. 72–73). There the gravity coverage is better and the overall density distribution better known (Smithson 1963a), and the supposed effect of the more superficial density variations has been added in the profiles XII to XIV.

7.3 DEPTH TO MOHO

A crust/mantle density contrast of $\Delta\varrho = 0.45$ g/ cm^3 (Section 4.4) and a two-dimensional density distribution was first assumed for all the models (Figs. 66–73). Starting with the profiles north of the Oslo Region where regional I and II coincide, emphasis was put on profiles II and III where most seismic control exists (below the NORSAR) (Fig. 75). The seismic Moho depth in profile I is regarded as atypical, occurring in the bottom of a local depression in the intersecting seismic refraction profile (Kanestrøm & Haugland 1971). By applying the same adjustment (same depth correction), similar computations were performed for the remaining profiles (IV–XIV). With the applied density contrast (0.45) and the depths fixed by the seismic data beneath the Oslo Region, the curves all show a close similarity to those obtained from the preliminary Moho contouring by Sellevoll & Warrick (1971) and those based on isostatic considerations (Fig. 19a).

Recalling that the regional I presumably reflects variations in Moho depth while regional II also comprises effects of mass anomalies above Moho but below the felsic intrusives in the Oslo Region, it is apparent from the profiles (Figs. 66–73) that Moho I (solid line on the figures), as reflected by the regional variation I, shows a marked thinning of the crust below the Oslo Region. The apparent asymmetric configuration, with a deeper Moho under the Caledonian mountain belt to the west than under the Precambrian shield to the east, is also clearly seen in the profiles.

For comparison, the profiles show what variations in crustal thickness would be necessary to account for the regional II. With the applied depth correction and the density contrast used (0.45 g/cm^3) this would bring the Moho from 2 to 4 km higher up below the Oslo Region (Moho II is shown in all the profiles as a heavy stippled line beneath the Oslo Region). This result is not confirmed by the seismic interpretations (Kanestrøm & Haugland 1971, Sellevoll 1972, Kanestrøm 1973). Then, to match the seismic depths below the Oslo Region, some other depth correction has to be applied. This, however, would not affect the Moho relief, and consequently lead to Moho depths down to 44–45 km below central southern Norway west of Oslo. These values are much too deep as compared with the preliminary Moho contouring (Sellevoll & Warrick 1971) or with general isostatic considerations. As inferred, therefore, the regional II cannot be explained by Moho undulations alone but by the presence of additional dense masses.

Variations in the depth to the Moho I in the models are also dependent on the density contrast chosen. Larger density contrasts would lead to less relief in the Moho curve, and would agree better with the Moho configurations obtained on the basis of the empirical relationships displayed in Fig. 19a. On the other hand, lower density contrasts which would be more consistent with the supposed density contrast at the base of the crust (Chapter 4), leads to a sharper Moho relief. Fig. 74a shows the Moho calculated on the basis of various density contrasts, all fixed by the NORSAR seismic depths below the Oslo Region.

Future seismic studies may help decide which of the density contrasts is the better choice in southern Norway. The density contrast applied for all the profiles (0.45 g/cm^3) is to be regarded just as an interpretational example; exact Moho depths are not to be read from these figures. The principal features of crustal thinning towards the west coast and beneath the Oslo graben axis are, however, consistent, regardless of which density contrast is used.

Another assumption invoked was the lateral constancy of crustal density. This requires constant depths to crustal interfaces or discontinuities. Reported seismic depths to the 'Conrad' interface in southern Norway (Kanestrøm & Haugland 1971, Sellevoll & Warrick 1971) are all in the range of 16–18 km, while in central and southern Sweden the depth may be about 10–12 km (Dahlman 1971, Gregersen 1971). These features have been included in Fig. 74a. To fit the gravity regional, the gradual change in depth to the 'Conrad' will affect the calculated depths to the Moho east of Oslo, bringing the crustal model more in harmony with the seismic interpretations from the adjacent parts of Sweden (Dahlman 1971) and inferring a more symmetrical configuration of the Moho upwarp below the graben than that indicated by the cruder models in the profiles, Figs. 67–73.

7.4 DENSE CRUSTAL BODY

The dense crustal body may be located at the base of the crust above the Moho discontinuity and in direct continuation with the mantle upwarp. Assuming various possible density contrasts (0.1, 0.15, and 0.2 g/cm^3), the

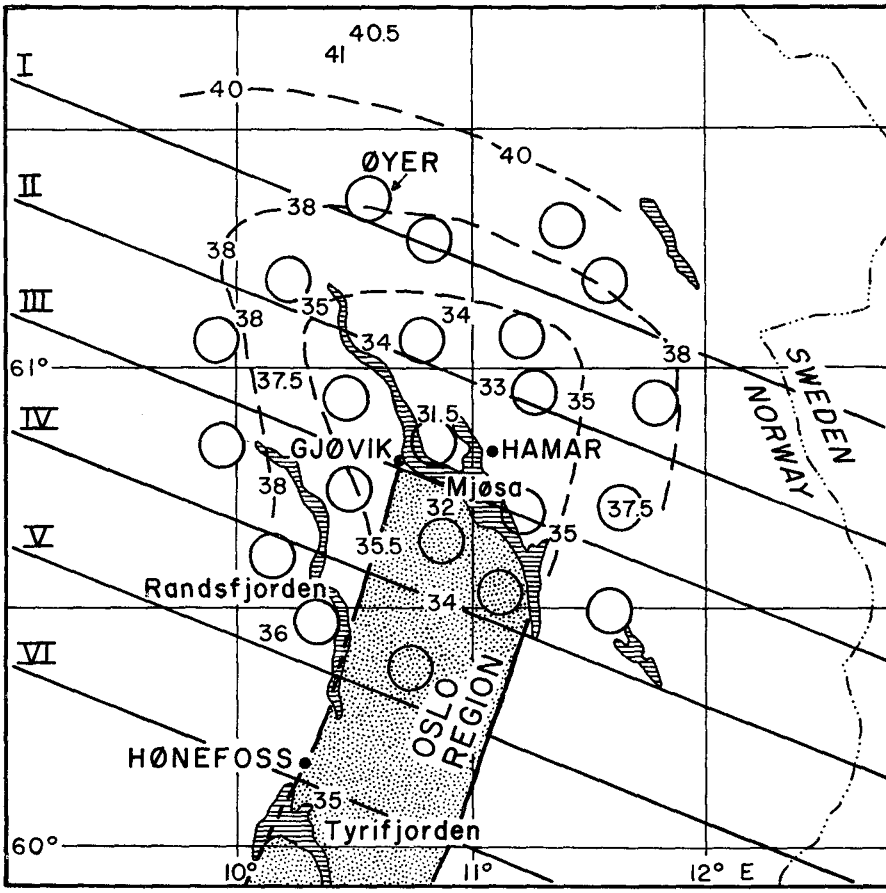


Fig. 75. Depth to Moho under the Norwegian Seismic Array (NORSAR) at the northern end of the Oslo Region (*after* Kanestrøm 1973).

various shapes of the inferred body have been calculated for all the profiles (Figs. 68–73). At this depth the anomalous body is on the average about 60 km wide, that is about 20 km wider than the surface expression of the graben.

With the lower level of the body being identical to the Moho I discontinuity, the upper level occurs on the average at about 26 km, 23.5 km and 20 km below surface with the density contrasts of 0.2, 0.15, and 0.1 g/cm^3, respectively. Considering the maximum depth calculations and the seismic interpretations (Sellevoll 1972), the contrast of 0.1 g/cm^3 would seem the most plausible. In that case the dense body would occupy more than ⅓ of the already thinned crust directly below the graben axis. This interpretation (alternative A) implies a density of the anomalous body of about 3.10 g/cm^3.

Recalling the fact that many uncertainties occur, such as the choice of crust/mantle density contrast, choice of regional, seismic depth estimates, density contrast of anomalous body versus crustal rocks, etc., a whole family of depth curves might be drawn dependent on the various assumptions made.

In spite of this the conclusion holds that some kind of *dense, intrusive zone present beneath the graben at deep to intermediate levels is consistent with gravity and other geophysical observations.*

The magnetic signature across the graben (Haines et al. 1970; see also profiles AA' and BB' in Ramberg 1972b) could perhaps imply instead the existence of an axial dike (or dikes) higher in the crust, such as in the Red Sea (Girdler 1958) or parts of the East African rifts (Griffiths et al. 1971, Searle & Gouin 1972). More detailed measurements by NGU show that the large aeromagnetic anomalies seem confined to the large composite batholiths of chiefly syenitic and monzonitic rocks, and that no conclusions bearing upon the postulated dense rocks at depth can be drawn on the basis of the aeromagnetic anomalies (K. Åm, pers. comm. 1972).

The broad Oslo gravity high would, however, evidently be satisfied by more shallow models becoming wider as they near the surface. Hence, alternative solutions are depicted in Fig. 74b. One alternative (B) assumes that the density contrast occurs at intermediate crustal levels. This solution could represent an intrusion at intermediate crustal levels, but the principal difference with respect to alternative A may only be that the material in alternative B has a somewhat lower density and (if continuous down to the Moho) consequently shows no density contrast to the deeper part of the crust (having a density of about 3.05 g/cm^3) but to the intermediate layer (about 2.92 g/cm^3). Solution (B) is equally satisfactory as the deep crustal solution (A) with regard to the gravity and geological data, but has not received seismic support (Kanestrøm & Haugland 1971, Sellevoll 1972).

As a third alternative (C), a shallow model is presented (Fig. 74b). Since the broad anomaly exists also over the nordmarkites and other felsic occurrences, in places possibly reaching down to 8–12 km or more (Chapter 6), this alternative is represented by a mass excess located as a plate-like intrusion just below the lower level of the various felsic rocks. In the profile chosen, a slab 100 km wide and 7–8 km thick (using a density contrast of 0.15 g/cm^3) will fit the gravity data.

If the postulated shallow intrusive (alt. C) of dense rocks was emplaced contemporaneously with the extrusion of the major, regional, basaltic flows at the surface, the intrusion pre-dates the period of major normal faulting. The larger normal faults, having estimated dip slips of 1–3 km, would probably also have affected the shallow dense slab and thereby created consistent marginal gravity highs with a central, relative low not depending upon surface lithological features. Since this is not the case, either normal faults did not reach sufficiently far down, or the dense rocks were intruded at a later stage and *after* the intrusion of monzonite, correlated with the time of RP extrusions. A shallow intrusion of regional extent, underlying the Oslo Graben over at least its full length and having a width of more than twice that of the graben, would seem to leave the foundering and formation of the graben unexplained. The formation of the dense shallow body would have to be a major tectonomagmatic event, the traces of which, however, have not been recognized on the surface. The one possible exception is the frequent occurrence of diabase dikes that formed at the present surface throughout the whole period of igneous activity in the Oslo Province. From the aeromagnetic anomalies as shown on the Canadian (Haines et al. 1970) and NGU maps there are no indications of such a broad, dense and shallow body. Moreover, the simple model of alt. C cannot be fitted perfectly ($\pm$ 3 mgal) to the profile in Fig. 74b, and the shallow model thus seems the least likely solution.

A new development has arisen from current research at NORSAR, indicating that P-wave travel time anomalies are not consistent either with a large Moho bulge or with the presence of highly concentrated mass anomalies in the crust or upper mantle (Berteussen 1975, E. Husebye, pers. comm. 1974). With the mass correctly recorded by gravity data, this would lead to small density contrasts and dispersed masses over large volumes, much in harmony with alt. B but with possible mass excesses located also higher up in the crust and even in the upper mantle.

In conclusion, of the many possible gravimetrical models, that requiring deep to intermediate (dispersed) dense masses is regarded as the most reasonable interpretation consistent with the overall geological and geophysical evidence.

7.5 MASS CALCULATIONS OF DEEP-SEATED BODY

On the assumptions that the regional I (Plate 3) correctly reflects the variation in crustal thickness, the gravity effect of the postulated dense mass is constructed by taking the difference between the regional II (Plate 4) and the regional I. This is the same as taking the difference between the two regional curves in all the profiles (Figs. 67–73) and combining them into a residual gravity map (II), Fig. 76. If the magnitude of the regional I below the Oslo Region is erroneously *small,* it means that the residual II includes too much mass that correctly belongs to the upwarped upper mantle. On the other hand, considering the preliminary results of Husebye and coworkers (E. Husebye, pers. comm. 1974), if the Moho upwarp has been overestimated, the mass estimates based on Fig. 76 will represent minimum figures. If we consider only the anomaly below the outcropping part of the graben (profiles II to XI), then the net mass excess as revealed by Fig. 76 has been calculated to $M = 1.21 \times 10^{19}$g.

With the deep crustal body (alternative A) and an assumed density contrast of 0.1 g/cm^3, the *total* mass is accordingly $M \approx 3.75 \cdot 10^{20}$g and the volume $V \approx 121{,}000$ km^3. If the intermediate crustal body is considered (alternative B), assuming a density contrast of about 0.12 g/cm^3, the total mass will be $3.1 \cdot 10^{20}$g (*plus* possible mass of Permian origin in the deep crustal layer of density 3.05 g/cm^3). Similarly, the volume estimate in this case will be about 100.000 km^3 (*plus* possible material intruded into the deeper crust). For the shallow model (alternative C) almost the same values and arguments will apply as for alternative B.

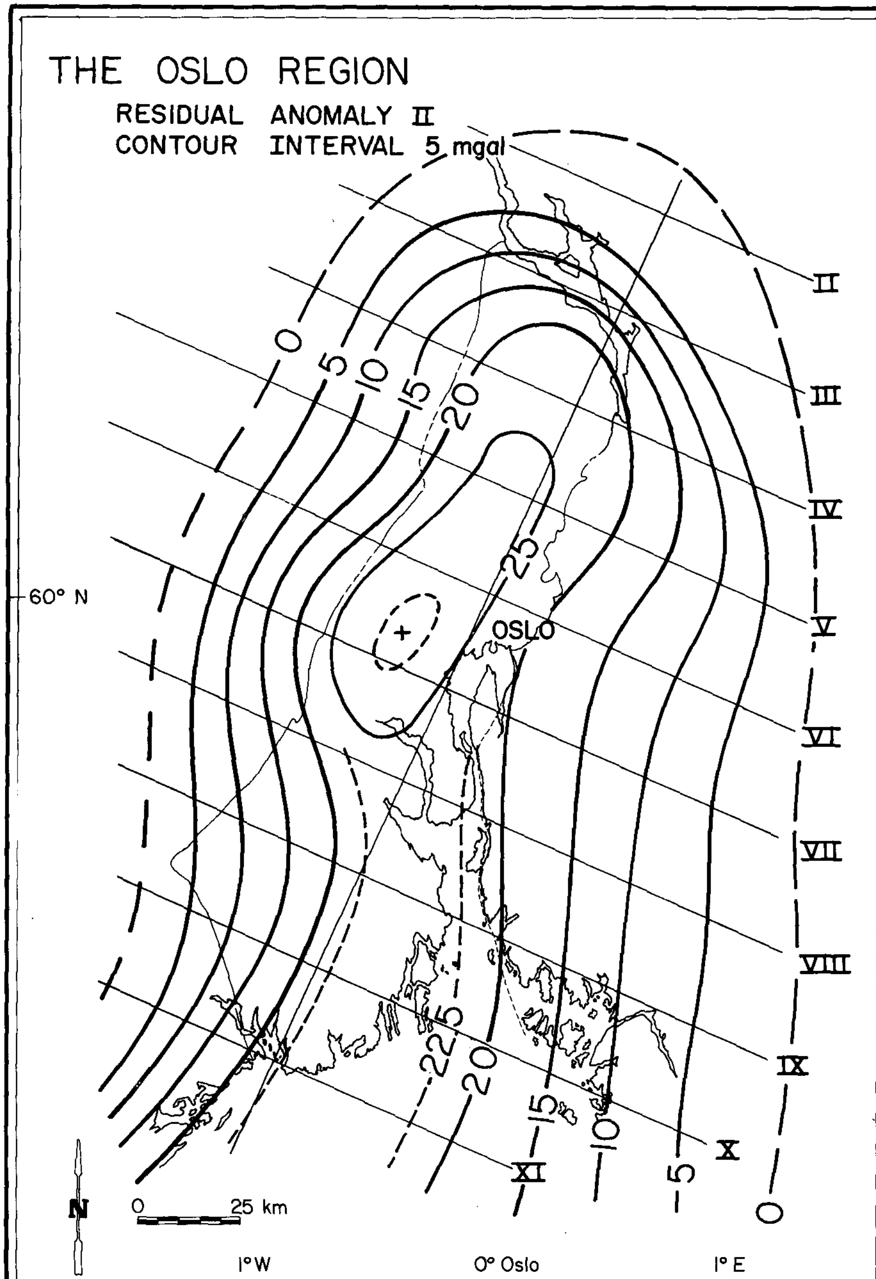


Fig. 76. Residual anomaly (II) of the Oslo Region obtained by taking the difference between regional II (Plate 4) and regional I (Plate 3).

The possible occurrence of low density material in the crust adjacent to the extensive dense body, and not dealt with in Chapter 6, will in effect reduce the calculated net mass and hence also the total mass and volume estimates. This, combined with a tendency to underestimate the residual anomalies at their tails (Parasnis 1973), will usually mean that the mass estimates represent *minimum* values. Considered together with the possibility of a smaller Moho upwarp than earlier assumed, this means that the total mass and volume are probably not less than $M = 3.75 \cdot 10^{20}$g and $V = 1.21 \cdot 10^5$ km^3, respectively, and might be increased by a factor of two. An additional factor is the contribution from the likely lateral continuation of the dense mass below the Skagerrak.

7.6 SKAGERRAK PROFILES

The crustal thinning and suggested deep-seated dense rocks do not seem to extend towards NNE much farther than the visible limits of the Oslo Graben. Towards SSW, however, the gravity field (Plate 1) implies that the rift structures may be continuous in the axial direction also below the Skagerrak.

Skagerrak is part of the epicontinental and generally rather shallow North Sea which in this area is marked by the Norwegian Channel, a 60–80 km wide topographic depression which runs for 800–900 km just outside the curved coastline of southern Norway (Fig. 77). The Norwegian Channel has been interpreted as originally formed by graben subsidence (Holtedahl 1950, 1960, 1963, Pratje 1952), while others favor the hypothesis that glacial erosion was of prime importance (Shepard 1931, Kuenen 1950, Sellevoll & Aalstad 1971).

Seismic refraction and reflection studies (Aric 1968, Weigel et al. 1970) in the area between Grimstad and Jammerbugt (Fig. 77) have provided information about the crustal structure below the Skagerrak. Sedimentary layers of about 5 km thickness extend from about 20–30 km off-shore Norway to Jutland, and increase to 7 km or more in central parts of the Danish Embayment (Casten & Hirschleber 1971). The depth to the Conrad discontinuity ($V_P = 6.6$ km/s) was found at 13–14 km below Skagerrak, while intermediate velocities of about 7.2 km/s were reported from depths of about 20 km. Depth estimates to Moho are all in the range 29–32 km below the Skagerrak, and in close agreement with those below Jutland as well as with those presented for the Oslo Graben (Kanestrøm & Haugland 1971, Sellevoll 1972). An E–W seismic model from Jutland (Casten & Hirschleber 1971) shows an increased depth to the Moho towards the east, in accordance with the inferred Moho depths of 36–38 km in southern Sweden (Gregersen 1971).

From aeromagnetic measurements, Sellevoll & Aalstad (1971) presented a contour map of depth to the magnetic basement in the Skagerrak. This indicates that the deep sedimentary basin thins out rather abruptly towards the north and east, that is towards the south coast of Norway and the Fennoscandian border zone.

Gravity surveys in Skagerrak have been carried out by Collette (1960) and Bedsted Andersen (1966); their data are included in the Bouguer anomaly

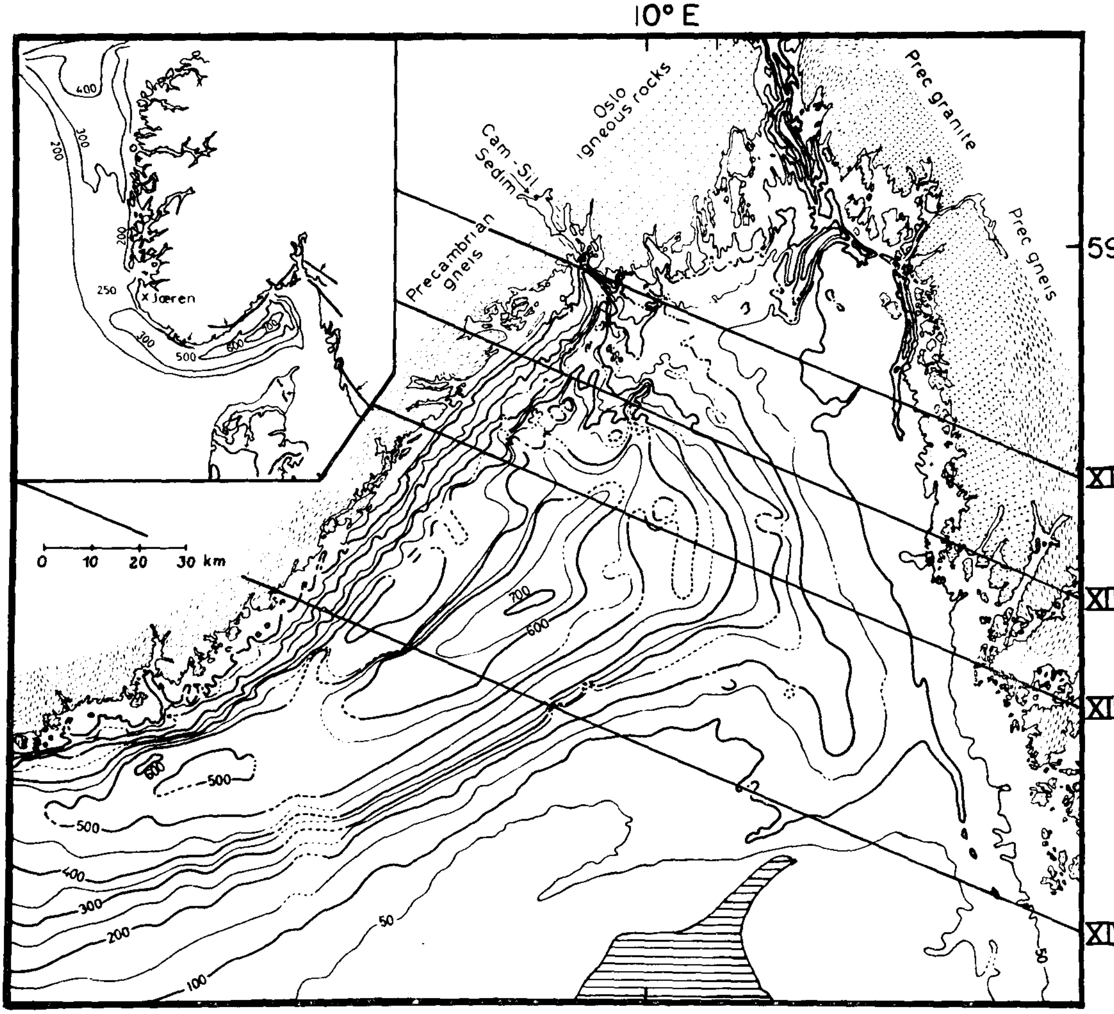


Fig. 77. Morphology (nos. in metres below sea-level) of the Norwegian Channel (*after* Holtedahl 1963) and location of gravity profiles XI to XIV.

map of Plate 1. From this map, and from the three profiles XII to XIV (Figs. 72–73) it is apparent that the characteristic 'Oslo high' is still present some 100 km SSW of the graben despite the considerable mass deficiency represented by the thick sedimentary sequence. Maximum Bouguer values (about + 60 mgal) do not coincide with maximum depths of the Norwegian Channel, but occur close to profile XII at shallow water depth (about 200 m), testifying that at least part of the gravity high is caused by density contrasts within the crystalline basement.

The main purpose in analyzing the profiles XII to XIV has been to test whether or not the principal structural interpretation suggested for the Oslo Graben to the northeast will also fit the gravity observations across the Skagerrak. In order to do so, the gravity effects of surface geological features close to the area of interest were stripped off, where possible. A gravity minimum on the east side of the broad 'Oslo high' extends from Sweden to the northernmost tip of Jutland; this is due to the Precambrian Bohus granite (Lind 1967a) and its likely submarine extension below the Kattegat (Sorgenfrei 1969). Likewise, to the west of the 'Oslo high' the gravity effect of some minor granite bodies and the dense Bamble rocks has been inferred (Ramberg & Smithson 1975). The seaward continuation of the Bamble rocks is substantiated by the aeromagnetic anomaly pattern observed outside the coast (Sellevoll & Aalstad 1971). Reduced sharpness in the gravity gradient at the Telemark/Bamble rock contact zone in profile XIV as compared with the profiles to the north, indicates that no notable mass of Bamble rocks is present that far south (outside Grimstad).

The depths to Moho were fixed by means of seismic data (Aric 1968, Weigel et al. 1970, Kanestrøm & Haugland 1971), and for the rest of the profiles calculated by means of the regional gravity anomalies assuming a crust/upper mantle density contrast of 0.45 g/cm^3 and applying the same depth correction as for the graben profiles (nos. III–IX). A simple, deep, dense body ($\Delta\varrho = 0.1$ g/cm^3) was assumed, the top of the body being located at ca. 20 km depth, at the intermediate discontinuity recorded by Weigel et al. (1970). The geometry of the sedimentary basin was based on the various seismic and magnetic interpretations cited above; the overall density contrast for the sedimentary layers (having reported P wave velocities in the range of 2.8–5.0 km/s) was set to 0.3 g/cm^3 relative to the Precambrian basement.

As can be seen from Figs. 72–73 the structural solution suggested fits the gravity observations in all the three profiles but for two minor exceptions. In profile XII (Fig. 72) a small gravity high on top of the broad one is not accounted for, and some rather shallow mass surplus has to be assumed below the sea floor. In profile XIV (Fig. 73) the deep block will not compensate the sedimentary sequence, and some rather *arbitrary* dense mass has been introduced to fit the data. These masses, which all appear to have subcircular outline, could either represent basic intrusions at intermediate to shallow depths, consistent with the presence of large basaltic volcanoes in the Skagerrak already indicated on the basis of aeromagnetic and gravity data (Åm 1973), or represent more dispersed masses in the whole 'sub-Skagerrak' column. It is also clear that the degree of accordance between calculated and observed gravity is critically dependent on the mass of the sedimentary sequence. Any change in structure or density contrast will have to enforce proportional changes on the remaining bodies.

One may conclude that the structural model favored for the Oslo Region with some modifications also represents a solution for the eastern part of the Skagerrak. Although less clear, the gravimetric signature of the Oslo Graben seems to continue still farther towards SSW where it intersects the axis of the Danish Embayment at the Jutland junction. The remarkable gravity high south of Kristiansand (at about 58° N/8° E) has been interpreted as being due to multiple intrusions of Tertiary and Permian age (Åm 1973). The magnetic anomalies northwest of Jutland are similar to those confined to the syenite and monzonite plutons within the graben area to the NNE, and various geological evidence from the south coast of Norway (Chapter 10) also indicates that the Permian igneous and tectonic activity was not confined to the visible graben.

8. Petrogenetic discussion

In a graben area, with its implied deep-seated foundations, tectonic and petrogenetic processes are so closely involved that separate discussions are hardly justified. However, for the sake of clarity, this chapter is chiefly concerned with the petrogenetic aspects, while structural and tectonic implications are dealt with in the last chapter.

Some major petrogenetic problems of the Oslo Region concern the origin of 1) the large quantity of felsic and intermediate rocks, 2) the dense mass at depth, its nature and possible relation to the exposed plutonic series, 3) the 'Oslo-essexites' and their relationship to the main plutonic series, and 4) the voluminous extrusives, their apparent bimodality, and the dominating role of rhomb-porphyry flows. New insight into these and related problems has been gained through the geophysical analysis and will be discussed briefly below.

8.1 PETROGENETIC HYPOTHESES AND THE FREQUENCY DISTRIBUTION

Bowen (1948, p. 87) maintained the hypothesis that "most granites have been produced throughout geologic time by differentiation of basic (basaltic) magma" and that "the granite is the normal uppermost differentiate produced by repeated gravitative and tectonic (filter-pressing) separation of a series of crystals from an everchanging mother liquor". In contrast to this view, students of the Oslo Region igneous province have for some time generally agreed that "the evolution of the large quantities of monzo-syenitic Oslo magma hardly could have proceeded from a basaltic 'mother magma' through fractional crystallization" (Barth 1954, p. 10), mainly because of the small amounts of mafic rocks exposed and the large amounts of intermediate and felsic rocks. While Barth (1945, 1954) and Oftedahl (1952, 1959, 1960, 1967) both considered the 'Oslo-essexite' series (Fig. 6a) to be the only mantle-derived intrusives, Barth (1954) envisaged the major portion of the igneous province (the Kjelsåsite-Granite plutonic series) to have formed by refusion and remobilization of the pre-existing Precambrian crust. By means of percolating emanations and heat brought about by a burst of degassing of the earth's interior, the Precambrian rocks were totally melted, the resulting monzonitic magma giving rise to the main differentiated series.

More recently Oftedahl (1967) looked into the various possibilities of magma formation within the Oslo Region: 1) anatectic magma formation in the deeper crust, moving upwards and changing gradually to more acidic products; 2) anatectic magma formation in the deeper crust, accompanied by gravitative differentiation *in situ;* 3) magma formation in the deeper crust, with comprehensive differentiation near the surface; 4) basaltic magmas from the upper mantle, undergoing a comprehensive crustal differentiation; 5) possible differentiation in the upper mantle. Oftedahl concluded that the alternatives (1) and (2) seemed the most likely processes for the formation of the plutonic series.

An opposing view arose from the Rb–Sr isotopic analyses of 28 rock samples from the main plutonic series (Heier & Compston 1969). The initial Sr^{87}/Sr^{86} ratio of the rock series of 0.7041 $\pm$ 0.0002 is within the range previously given by Czamanske (1965) on two granite samples from the Finnemarka complex. Heier & Compston concluded (p. 143) that the ratio "restricts the magma source to the upper mantle, but a deeper crustal origin cannot be excluded because of the lack of knowledge of the Rb/Sr ratio in this material", and further that "contamination from the Precambrian gneisses and granites and the Paleozoic sediments into which the plutonic rocks were intruded, may safely be ruled out." In spite of this, "the stupendous volume of syenitic rocks in the Oslo province" seems so large that "the classical concept of fractionation of basic magmas appears inadequate" (Bose 1969, p. 24). The frequency distribution of the Oslo plutonic rocks as given by Barth (1954) and reproduced here as Fig. 78, has apparently represented a major obstacle to the acknowledgement of rivalling ideas on the hypothesis of anatexis. Although Barth (1954) stressed that the frequency distribution pictured only the present surface conditions and suggested the possible presence of an approximately ten kilometer thick layer of 'Oslo-essexitic' composition in the deep crust, the frequency distribution curve (Fig. 78) led him to the conclusion that the gabbroic rocks of the Oslo Region can hardly bear any parental relations to the granitic-syenitic rocks, but that the main plutonic series formed by crystallization-differentiation of a monzonitic-syenitic magma.

The gravity analysis presented here shows quite clearly that *the Oslo Region felsic rocks are associated with vast amounts of dense, mafic to ultramafic rocks at depth.* The mass and volume estimates given in the previous chapters provide a more quantitative basis for the construction of a revised frequency distribution diagram (Fig. 79). The diagram, which reveals the *volume* estimates of the main rocks groups, also includes the equivalent lava volumes assumed to total about 10,000 km^3 (Chapter 2). The mafic to ultramafic group comprises the dense, intermediate to deep, crustal body, the hypothetic shallow intrusions as marked by the Narrefjell, Asker–Lier and Horten–Tønsberg gravity highs (Table 12), and the 'Oslo-essexites' (Table 11). Changes in the postulated depth to Moho could increase the total mass of the mafic rocks by a factor of two. The volumes of the felsic rocks are based on data given in Table 9, while no definite estimate can be offered for the plutonic rocks of intermediate composition. However, various lines of evidence (Sections 6.2, 6.4, and 6.5) suggest that large portions of the

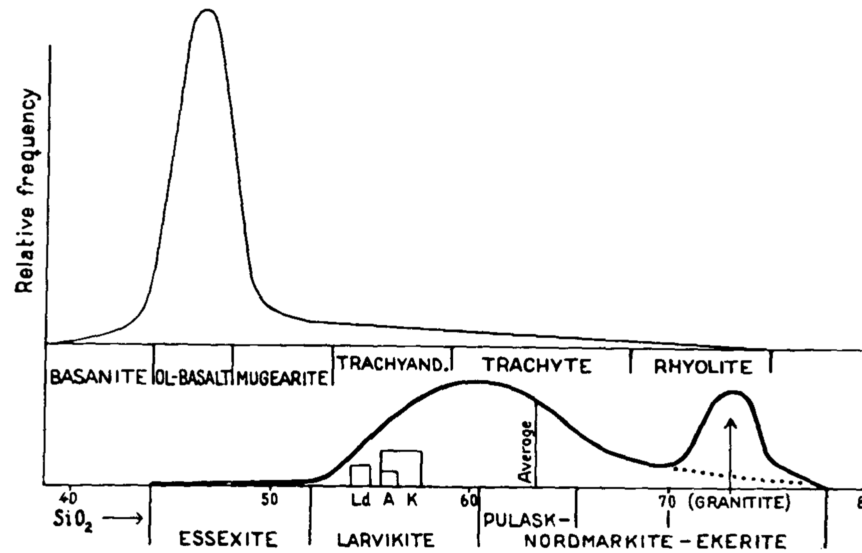


Fig. 78. Areal frequency distribution of the Oslo plutonic rocks (*after* Barth 1945) compared with the lava series of Midland Valley, Scotland (*after* Tomkeieff 1937). Ld = lardalite, A = akerite, K = kjelsåsite; see Table 3.

upper and middle crust consist of syenodioritic rocks and that they probably constitute an equivalent or larger volume than that of the felsic rocks (Section 6.4).

The great preponderance of dense (mafic/ultramafic) rocks in Fig. 79 is striking, the felsic rocks composing only about 12.5 vol. % with respect to the mafics/ultramafics. A similar calculation for the estimated *total mass* involved shows that the felsic rocks constitute about 10%. The less ambiguous *anomalous mass* estimates show the same trend as in Fig. 79 and give about 14.5% felsic rocks relative to the anomalous mass of the mafic/ultramafic rocks. If we assume that the mass and volume of the intermediate rocks is approximately twice that of the felsic rocks, the latter represents about 10%, 8% or 9% of the anomalous mass, total mass or volume, respectively, if the total amount of all the igneous rock suite in the region (north of profile XI) is considered. Since a number of uncertainties are involved in mass and volume estimates, it is the general trend of the frequency curve rather than the indivdual numbers that should be emphasized.

Because granites formed by crystal fractionation from a gabbroic source will constitute approximately 10% of the original reservoir (Grout 1948), *there is no volumetric objection to the idea that the felsic rocks of the Oslo Region originated from a basaltic mother magma.* It is suggested that while the magma formation *in general* might be due to 'classical' differentiation in the sense of Bowen (1928, 1948) and that the basaltic magma represents the ultimate source, the evolutionary process may in detail have been very complex (e.g., Raade 1973, Czamanske & Mihálik 1972) involving processes such as gas transfer, refusion and liquid immiscibility.

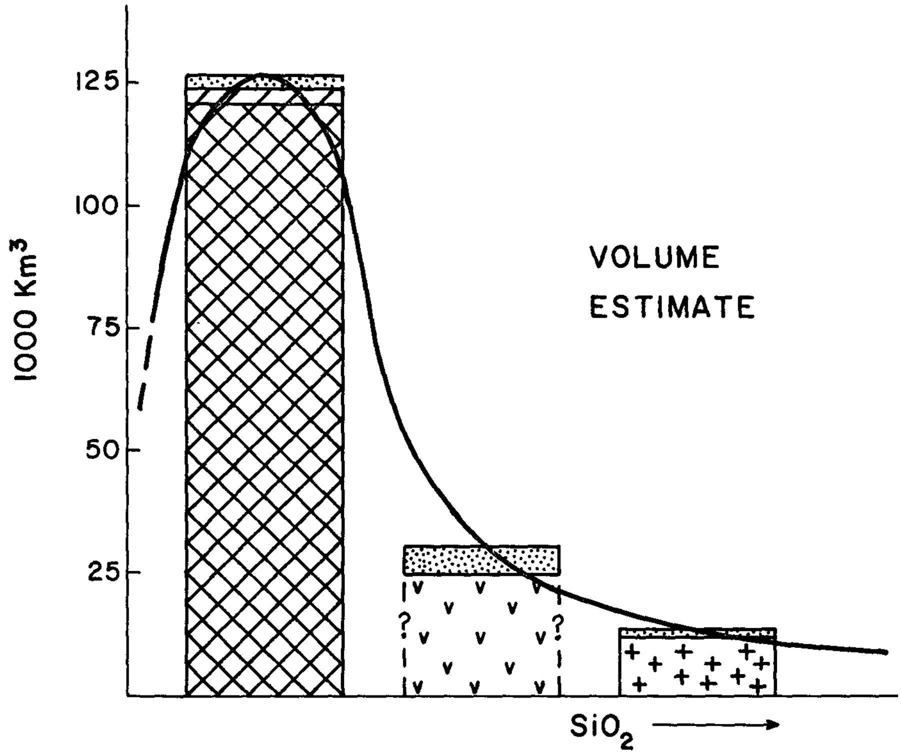


Fig. 79. Volume frequency distribution of the Oslo igneous rock series, based on mass calculations (deep-seated rocks) and geological reasoning (lavas [dotted], see Chapter 2.5). Deep-seated mafic/ultramafic rocks (cross-hatched), more shallow mafic rocks (obliquely lined), intermediate rocks (V's), felsic intrusives (crosses).

The frequency curve of Fig. 79, with reservations especially for the unknown amount of rocks of intermediate composition, is strikingly similar to the curve given for the lava series of the almost contemporaneous Midland Valley graben, Scotland (Kennedy 1958, Dunning 1966) (Fig. 78). The predominance of olivine basalt as well as the shape of the frequency curve has been thought to be indicative of a normal comagmatic series with olivine basalt as the parental magma (Tomkeieff 1937).

In Iceland, young acidic intrusions occur in a zone parallel to the central rift (Gale et al. 1966), while about 10% of all extrusives are of acidic and intermediate compositions (Thorarinson 1967). While Walker (1966), Thorarinson (1967) and others have suggested that Iceland may have a sialic basement to account for the felsic igneous products, Ward (1971), in a review paper, concluded that the island is part of the mid-oceanic ridge-rift system and suggested that felsic rocks may be just as common along other oceanic ridge crests. Hence, the frequency distribution of the volcanics from the Permo-Carboniferous continental rift in Scotland (and from the plutonic Oslo series) may be more similar than commonly conceived to the distribution in slow-spreading oceanic rifts where the extrusives are thought to be derived from the upper mantle primarily by the process of partial melting (Wyllie 1971). It is also of interest that alkali basalts may be more common along oceanic rifts than previously assumed (Jacobsson 1972).

Voluminous alkaline volcanism also accompanied the development of the eastern rift system of Africa (Williams 1970, 1972, King & Chapman 1972, Baker et al. 1972). As in the Midland Valley, the volume estimate of the volcanics in Kenya and Ethiopia indicates that the volume of basaltic flows heavily outweighs the occurrences of trachyte, rhyolite, etc., and Saggerson (1970) suggested that all the volcanic rock-types in central Kenya were differentiation products of parental alkali olivine basalt. However, Wright (1965) previously argued that in the case of differentiation, enormous volumes of unexposed basaltic rocks had to be postulated. The existence of such rocks now seems almost proved. While volumetric ratios of extrusives may be somewhat accidental due to the differential ability of various magmas to extrude, and to subsequent erosion, etc., the frequency distribution of the *plutonic* rocks may be more significant. From seismic and gravity studies it is clear that vast amounts of dense rocks are present at depth beneath the East African rifts (Searle 1970, Kahn & Mansfield 1971, Searle & Gouin 1972, Griffiths 1972). Likewise, from their detailed study of the Silali caldera volcano, Kenya, McCall & Hornung (1972) concluded that the volcanoes are inevitably differentiation products of mantle-derived material, and that the Gregory Rift is possibly underlain by a chain of cupolas of felsic composition. Similar conclusions were reached by Weaver et al. (1972) who suggested that the differentiated lava series from Kenyan and Ethiopian volcanoes derive from cupolas of salic magma on top of large basaltic reservoirs. These subsurface intrusives may be equivalents of the many felsic batholiths exposed in the much deeper eroded Oslo Region.

The similarities between the Oslo Graben and other rifted regions would suggest a common pattern of magmatic and structural development. Thus, the detailed geological and gravity data from the Oslo Graben may not only provide evidence for a revised petrological model for this region, but, by analogy, may also have some influence on geological reasoning in other, less exposed and more recent continental rift regions.

Magma generation associated with cratogenic rifting is commonly associated with alkaline rock suites evidently related to deep-seated tectonomagmatic processes (Wyllie 1967). However, granite formation in general cannot be viewed in the light of the suggested genesis of the Oslo Region felsic rocks. Orogenic granites *(sensu lato)*, like the huge batholiths of the Andes mountain belt, and the Sierra Nevada, Idaho and British Columbia batholiths, etc., all seem characteristically associated with calc-alkaline andesitic series and are thought to have formed by melting at high temperatures near the base of a highly desiccated crust (Brown 1973). These magmas probably generated by partial fusion of oceanic lithosphere along subduction zones (Oxburgh & Turcotte 1970, Gilluly 1971, and others). Although rather low, the Sr^{87}/Sr^{86} isotopic ratios of many orogenic granites (Hurley et al. 1965) are still significantly higher than those in the Oslo plutonic series, and may reflect a complex origin with contributions both from oceanic basalts and crustal material.

8.2 NATURE OF DEEP-SEATED, DENSE MASS

The subsurface dense masses, as revealed for example by Fig. 76, are located in the same area as the felsic rocks of the graben and closely follow the axis of the presumed crustal thinning. This is certainly not accidental, and the phenomena are regarded as most likely representing various responses to the same causal process which may be rifting.

Continental rifting in general is associated with the extrusion of large volumes of basaltic rocks commonly of alkaline character (Turner & Verhoogen 1960, Freund 1966, Beloussov 1969, Baker et al. 1972). In the 1960's it became clear that various depths of magma formation and fractionation would lead to the production of different types of basaltic magma (Kuno 1959, Yoder & Tilley 1962, Green & Ringwood 1967a, O'Hara 1968). The commonly occurring pyrope-bearing nodules of alkali olivine basalts indicate that the source area of the alkaline magmas is to be found in the pressure range equivalent to 70–150 km depth (Green & Ringwood 1970, Turcotte & Oxburgh 1969), corresponding to the upper part of the low velocity layer. With fractional melting of perhaps 10–20% of mantle pyrolite, alkali basaltic magma will form and rise towards the base of the crust presumably due to the density contrast (the calculated magmatic liquid density is about 2.63 g/cm^3 for olivine basalt at 1100° C (Bottinga & Weill 1970)).

In the Oslo Region the alkali basaltic material rose towards the surface in response to sub-crustal movements, whether by simple buoyancy, by convection, or by other, as yet unknown movements connected with rifting. Although large volumes were extruded on the surface as basaltic lavas, it seems *a major portion of the rising mass was trapped in the crust and differentiated there,* as indicated by the gravity anomalies.

Seismic refraction data from the Oslo Region have revealed P_i wave velocities of 7.5–7.6 km/sec. from an 'intermediate layer' at depths between 22–23 km and that of the Moho discontinuity at about 32 km (Sellevoll 1972). This implies rock densities in the range 3.10–3.15 g/cm^3 (see Table 8). Unless the high P_i wave velocities are modified by later studies, the postulated, deep, graben 'block' must consist of rather unusual rocks. While the major part may be gabbro, the block may include olivine- and pyroxene-rich gabbro, peridotite or other ultramafics, in order to justify the inferred high rock density. This may be explained by the hypothesis advanced in the previous section, that the felsic and intermediate rocks of the igneous province originated principally by crystal fractionation from a gabbroic source rock. If the deep crustal body acted as such a source, it will inevitably represent a residuum including significant amounts of cumulates, bringing the overall density above the assumed density of about 3.00–3.05 for common gabbroic rocks.

Based on experimental studies of the gabbro-eclogite transition, Green & Ringwood (1967b) suggested that eclogite, as opposed to earlier views, is thermodynamically stable through most of the continental crust, Fig. 80. 'Basaltic' rocks at the base of the crust would consequently be transformed to eclogite, or, in tectonically active regions characterized by a high heat flow, to the transitional garnet-granulite phase (Ringwood & Green 1966). Green & Ringwood's suggestion is based on an extrapolation of the boundaries of the garnet granulite field to lower temperatures outside the area of experimental determination. From Fig. 80 it is clear that conflicting interpretations of the stability field at low temperatures have been presented. This uncertainty, considered with the fact that eclogite has a P wave velocity of about 8.4 km/sec (and corresponding density of about 3.5 g/cm^3) compared with the observed values in the Oslo Region of 7.5–7.6 km/sec (and 3.10 –3.15 g/cm^3), makes it a reasonable working hypothesis that *the deep-seated mass below the Oslo Graben presently consists either of gabbroic rocks with minor amounts of ultramafics or of the isochemical garnet granulite phase.* With a smaller density contrast (in accordance with the interpretation of the P wave travel time anomalies (Berteussen 1975, Husebye, pers. comm. 1974)) less ultramafic rocks are required or, more likely, *the body represents an inhomogeneous mixture of pre-existing crustal rocks and (meta-)intrusives of mantle provenance.*

It is well known that many active tectonic zones of the world are characterized by high heat flow values and by anomalous low upper mantle velocities in the range 7.2–7.8 km/sec. Rift 'pillows', 'cushions', or asthenoliths with similar seismic velocities and modes of occurrence as in the Oslo Region have in the last few years been inferred from a number of rifted regions. From the Cenozoic Rhine Graben, Meissner et al. (1970) reported a pillow-shaped 'transition zone' with compressional velocities of 7.2 to 7.4 km/sec at about 25 to 35 km depth between crust and mantle. In another profile further to the south, Ansorge et al. (1970) recognized the same structure, the velocity here being estimated to 7.6–7.7 km/sec. Gravity profiles (Closs & Plaumann 1967, Meissner et al. 1970, Mueller 1970, Lecolazet 1970) show high flank values and a marked low above the subsided sedimentary fill. By compensating for the more than 4 km-thick sedimentary sequence (Boigk et al. 1967) and subsurface felsic intrusives (Mueller 1970) the gravity anomaly suggests a considerable mass surplus along the graben. Illies (1975) shows how a pronounced Moho upwarp reaches a maximum below the Kaiserstuhl volcanic center. Similarly, the Permo-Carboniferous Midland Valley in Scotland is associated with a broad gravity high of about 25–30 mgal (McLean & Qureshi 1966), caused by a relative thinning o fthe crust below the Midland Valley, by mass anomalies wihtin the crust, or by a combination of the two features.

The same conclusion may apply to the Cenozoic Baikal Rift where, again, a wide gravity high becomes apparent when correcting for a sediment thickness up to 6 km (Zorin 1966). Artemjev & Artyushkov (1971, p. 1201)

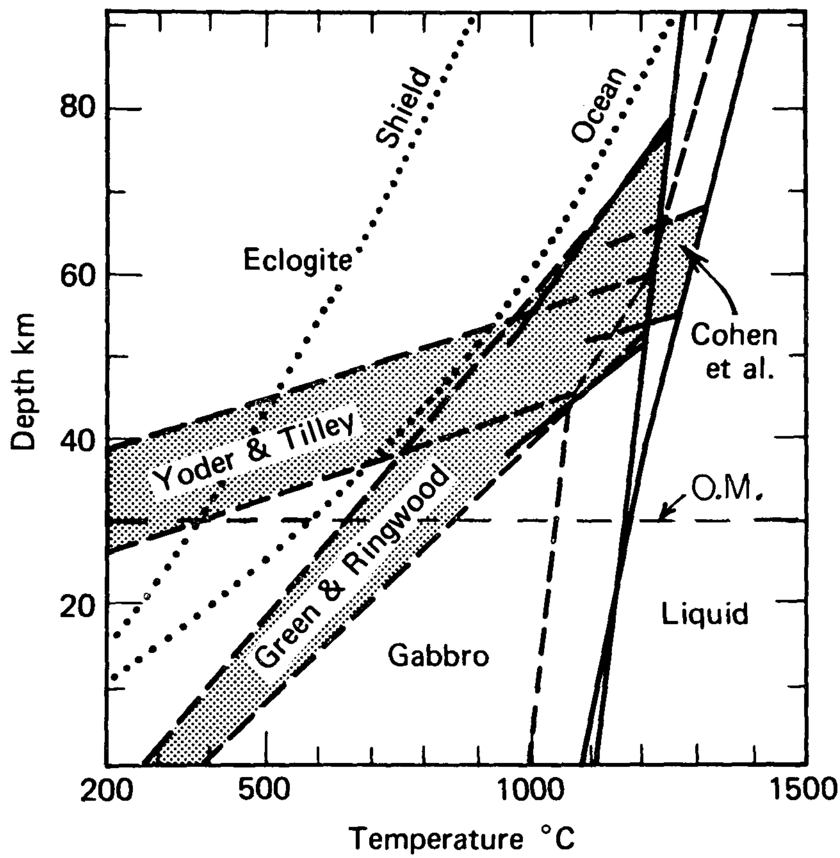


Fig. 80. Gabbro-eclogite stability fields according to various sources. Shaded areas represent transitional zones. Stippled lines show assumed normal thermal gradients in shield and oceanic areas. Horizontal dashed line (O.M.) indicates crustal thickness in the Oslo Region above which the dense intrusives may be located.

stated that "... as usual, one assumes that the obtained anomaly is connected mainly with the Mohorovicic discontinuity," and a considerable crustal thinning is suggested. The subcrustal seismic velocities immediately beneath the Baikal rift are in the range 7.1–7.5 km/sec while the upwarp of a layer of high electric conductivity is supposedly connected with anomalous heating of the upper mantle material by a few hundred degrees at depths of about 100 km (Artemjev & Artyushkov 1971).

In the East African rifts, Fairhead & Girdler (1971, 1972) pointed out that the regions of travel time delay, S_n attentuation, and the long-wavelength negative Bouguer anomalies under the rifts are all related to the presence of a low-velocity layer considered to be continuous with the major low velocity zone (asthenosphere) below the lithosphere plate. In addition, seismic refraction data in the Kenya rift (Griffiths et al. 1971, Griffiths 1972) indicate the existence of dense material with seismic velocities of about 7.5 km/sec at deep and intermediate crustal levels, consistent with gravity data from Kenya and Ethiopia (Searle 1970, Khan & Mansfield 1971, Searle

& Gouin 1972, Darracott et al. 1972), showing a marked axial high and implying that the asthenosphere has penetrated the crust to within 20 km of the surface.

Low-velocity, low-density, upper mantle material apparently underlies the anomalously thin crust below the Basin and Range System of the Western United States (Pakiser & Steinhart 1964, Cook 1966) as well as the Afar triangle (Searle & Gouin 1971, Makris et al. 1972); the latter area has a transitional position between the African rifts and the Red Sea and Gulf of Aden. In the Red Sea, a positive gravity anomaly is associated with the central trough (Girdler 1958, Drake & Girdler 1964), and it has been interpreted to be due to a mass of basic material intruded through down-faulted basement blocks. The basic material has varying seismic velocities in the range 6.8–7.3 km/sec.

Since the early work of Bullard (1936) it has been customary to assume that continental rifts usually are characterized by negative gravity signatures. It now seems justified to state that *continental rifts are frequently associated with major gravity highs, implying related subsurface structures and a likely deep-seated common origin for many of these features.* However, negative gravity anomalies may occur especially when volcanic and sedimentary deposits as well as shallow felsic intrusives have developed in response to the graben formation. In particular, some of the grabens of the western branch of the African rifts that are virtually devoid of volcanic products exhibit negative anomalies (Wohlenberg 1970) that cannot easily be ascribed to the commonly rather thin sedimentary cover.

In active rift systems, no clear Moho reflections are usually observed. The low-velocity material is regarded as a part of the uppermost mantle, possibly representing the top portion of a diapiric uprise from the low velocity zone or asthenosphere, while in older rifts the 'rift-cushion' has been injected up into the crust. In the Oslo Graben the 'intermediate layer' overlies a Moho discontinuity which seems to be clearly restored (Kanestrøm & Haugland 1971, E. Husebye, pers. comm. 1974) while the upper mantle apparently has normal seismic velocities (Kanestrøm 1972). The Paleozoic Oslo Graben may be typical for regions where the rift-forming processes culminated long ago. The anomalous thermal conditions that brought about phase transformation and partial fusion of the upper mantle material have ceased, but the dense masses, and associated derivatives, thought to have been intruded from below as products of these differentiation processes, remain.

8.3 CONTINENTAL VERSUS OCEANIC RIFTS. A COMMENT

If the rift-forming processes are of a global nature, they will probably occur beneath both continental and oceanic regions. Continental rifts, however, may be regarded as intra-plate features in contrast to the typical accreting plate boundaries defined by the oceanic rifts. The forces involved will thus have to deal with contrasted types of crust or 'overburden', leading to different

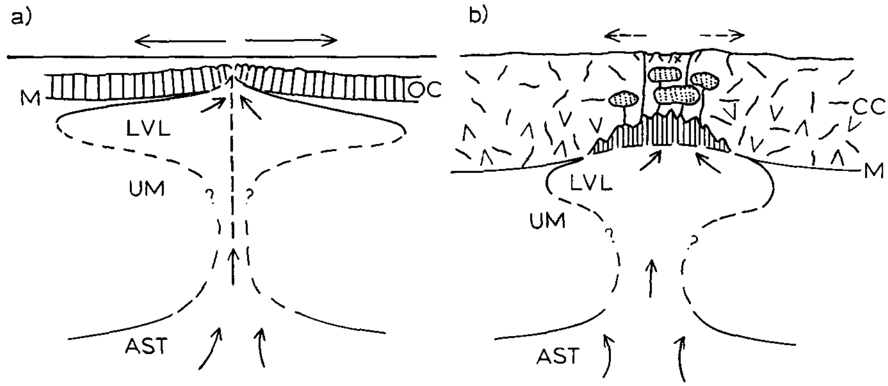


Fig. 81. Schematic models for the formation of (a) oceanic rifts (see e.g. Miyashiro et al. 1970) and (b) continental rifts. Postulated low-velocity layer (LVL) or 'rift cushion' exhibits negative density contrast with respect to the continental crust (CC) and possibly also to the much thinner oceanic crust (OC) when solidified, causing broad negative anomalies in active oceanic rifts (a) and negative anomalies superimposed by marked gravity highs in continental rifts (b). M = Moho, AST = Asthenosphere or low velocity zone. Horizontal arrows proportional to spreading rate.

trends of structural and magmatic evolution and to partly different geophysical characteristics.

Axial gravity highs, for instance, characterize many recent and paleocontinental rifts, while mid-oceanic ridge-rift systems seem to be associated with long wavelength negative Bouguer anomalies correlated with the presence of asthenoliths (Båth 1960, Le Pichon et al. 1965, Talwani et al. 1965). In continental rifts, the intrusion of basaltic materials into the thick and relatively less dense crust creates the gravity high that might outweigh possible long wave-length anomalies caused by the replacement of normal mantle by the upward rising, light asthenolithic material (Sowerbutts 1969, Girdler et al. 1969). Thus, the difference in geophysical characteristics may merely reflect different stages of the same general process (Fig. 81), the continental rift representing the potential or incipient plate boundary.

While the thick and heterogeneous continental crust is believed to resist tensional forces to a greater extent than the oceanic regions where the plates supposedly will move more freely, the difference in spreading rates is believed to have some effect on the magmatic development. Where spreading is rapid, as in many mid-oceanic rifts, a relatively regular and voluminous rise of basaltic material takes place. In areas with slow (0.1–0.5 cm/year) or even aborted spreading, long-term crustal accretion is feasible (Baker et al. 1972). The thick continental crust is also cooler at corresponding depths compared to oceanic crust. Proportionally less material is expected to reach the surface as volcanics while a major portion undergoes differentiation and possibly contamination within the crust, leading to relatively more felsic intrusives and to a wider range of igneous rocks than normally found in oceanic regions.

It has been argued (Gilluly 1971) that the more alkaline character of the continental rift rock series as compared with the oceanic rifts implies different origins and that they "represent two quite distinct types of world tectonics" (Le Bas 1971, p. 85). This difference might also relate to depth of magma formation. The incipient melting below an embryonic (and continental) rift presumably takes place at great depth within the pressure range of olivine alkali basalt formation (Green & Ringwood 1967). In oceanic regions where sea-floor spreading and copious basalt formation have been going on for some geological time, the isotherms have been moved upwards (Oxburgh & Turcotte 1968) into the pressure regime of tholeiite formation. This model implies that the change from alkaline to tholeiitic basalt formation may be gradational, and that the most typical alkaline provinces, apart from continental rifts, are to be found close to the original plate margins away from the present axial zones of the mid-oceanic ridges. The occurrence of alkaline provinces along the contact of Africa and South America in a pre-drift construction as demonstrated by Le Bas (1971, Fig. 1) is in agreement with this concept, and McBirney & Gass (1967) have shown that volcanic islands generally tend to be more alkalic the farther they are from the ridges. In the Kenya rift there is a marked transition from the strongly alkaline (pre-drift) Miocene basalt flows to the more mildly alkaline post-Miocene basalts of the rift floor (Williams 1972). In the Afar depression, Mohr (1971) and Barberi et al. (1972) subdivided the volcanism into continental and oceanic types. The former is more alkaline and believed to be an early manifestation of the continental rifting, while the latter, being confined to the central rift axis, is related to crustal separation of oceanic type. Lipman (1969) came to similar conclusions for the distribution of tholeiitic and alkali basalts in the Rio Grande rift in southwest USA.

The contrasted petrographical characteristics of oceanic and continental rifts are, therefore, in part inherited from various crustal thicknesses and depths of magma formation, and the great masses of largely alkaline extrusive and intrusive rocks associated with the Oslo Graben may be typical for continental rifts that failed to spread. In parts of the Oslo Graben, tholeiitic basalts apparently occur in the stratigraphically lowermost lava units only (Weigand 1975). These features need to be studied further, but variations in the alkali/lime ratio evidently also depend on factors other than the overall tectonic setting; such factors include, for instance, variations in the degree of partial melting, and melting of parts of the mantle with different water content (Green 1970, 1971).

8.4 ANATEXIS OF CRUSTAL ROCKS

It has been shown (Section 5.7) that gravity and magnetic anomaly trends in the NNW and also the NNE directions can be traced from the surrounding Precambrian terrain into the Oslo Graben, suggesting that some of the Oslo rocks and their spatial distribution result from partial melting and remobilization of the pre-existing Precambrian crustal rocks.

Based on experimental studies in the system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2–H_2O Tuttle & Bowen (1958) showed that large batholitic masses of granitic rocks may form by partial melting of crustal rocks in the presence of small amounts of water. Recent experimental studies (Brown & Fyfe 1970, 1972, Wyllie 1971) show that incipient melting of granite takes place at about 700° C and 25–30 km depth provided there is a release of water through dehydration reactions. Heat flow values in active rift regions (> 2HFU) (Le Pichon & Langseth 1969, Decker & Smithson 1975) and thermal conductivity of crustal rocks suggest that the deeper part of the crust in such regions has temperatures in excess of 700° C. The input of excess heat in the form of large, hot, mantle-derived intrusives increases the possibility of having granitic, syenitic, or even monzonitic magma formation according to the concept of anatexis. As an example, Hodge (1974) calculated that in an area of a normal thermal gradient of 20° C/km, a 4 km-thick slab of magma (T_M = 1200° C) at 30 km depth would provide maximum temperatures such that about one kilometer of country rock above the upper contact of the magma and about four kilometers below the magma would be subjected to partial melting. Principally being a function of the thickness of the intrusive and of the initial temperature and composition of the country rock, the size of the aureole of partial melting and the proportion of the low-melt fraction in relation to the emplaced magma will vary; but the important result is that significant volumes of partial melt are likely to occur under the conditions described.

These results have a bearing upon the petrogenesis of the Oslo rocks in that they seem to substantiate the idea of Barth (1954) that the main plutonic series originated by complete fusion of the Precambrian block, aided by the input of heat and alkaline solutions from the mantle. This conclusion appears to be contradicted by the Sr^{87}/Sr^{86} isotopic data presented by Czamanske (1965) and Heier & Compston (1969), which suggest that the rock province as a whole most likely is of deep crustal or preferably of upper mantle origin. If this is so, then the observed gravity and magnetic trends may be attributed to tectonic control whereas the postulated dense masses of intrusive rocks evidently must have had a substantial effect upon the temperature distribution in the lower crust. One might envisage that, (a) the temperature rose above the minimum level required for incipient melting of a granitic fraction, but that the crust represented a differentiated, mature type that was already efficiently depleted for potential melt; hence, virtually no fractional melt was formed; or (b) that some partial melting actually took place. Once the light, anatectic granitic melts were formed at the lowest temperatures of crustal fusion, they rose into lower pressure regimes where trapped by their wet melting curve (Cann 1970), they were forced to crystallize.

A possible anatectic origin for *some* of the Oslo Region felsic intrusives has recently been indicated (Raade 1973). Larger volumes may be present at *depth,* since anatectic magmas generated in the lower crust are likely to crystallize during their ascent to higher crustal levels unless they are exceptionally dry. However, while granitic magmas have a rather restricted mobility, magmas of intermediate composition generated at high temperatures in the deep crust, either by crystal fractionation or by partial fusion, may rise to much shallower levels before they crystallize (Brown 1973).

In conclusion, while gravimetric evidence indicates that *the major part of the igneous rocks of the Oslo province might have originated by differentiation from a gabbroic source, the gabbroic source possibly caused some partial fusion in the lower crust.* How much of the anatectic melt that eventually rose to the surface, if any, or mixed with other magmas formed, is a problem that invites further studies.

8.5 SUMMARY OF EVENTS

A summary of the petrological events in the Oslo Region will at the present stage serve as a preliminary model. Numerous problems remain for future research; however, the scheme presented may offer a framework for such studies.

The Permian igneous activity, as manifest from surface geological features, commenced with the outpouring of mainly alkaline basaltic flows generally believed to be derived by partial melting at relatively great depths (70–150 km), possibly within the upper part of the low velocity zone (Green & Ringwood 1967a, 1970, Ito & Kennedy 1968, Turcotte & Oxburgh 1969). Lack of ultramafic nodules within the Oslo basalts implies either that some differentiation took place in the lower crust or, more likely, that the composition of the rising magma has been slightly modified by gravitational differentiation due to a slow ascent.

Above the locally very thick B1 basaltic flows, the alternate occurrence and interfingering of basaltic and rhomb-porphyry flows demonstrate an apparent bimodality and coexistence of chemically contrasted magmas at this time. In the present case, neither the apparent bimodal distribution nor the preponderance of volcanics of intermediate composition are likely criteria by which one could rule out an origin by differentiation from a gabbroic source. The gravity study has revealed that surface geology is not quantitatively representative for the whole igneous province; the apparent bimodality could be due to disproportionate surface distribution. Further, there is little chemical difference between some of the stratigraphically higher basaltic flows and the most basic rhomb-porphyries, and a gradational trend also occurs from the stratigraphically lower rhomb-porphyries (and other trachyte flows) to the uppermost rhyolitic beds (Oftedahl 1967). The possible development of a monzonitic (monzodioritic) magma by differentiation from a deep crustal gabbroic source is suggested by the gravity interpretation and conforms with the reported low initial Sr^{87}/Sr^{86} isotopic ratio (Heier & Compston 1969). Furthermore, the RE fractionation pattern (Finstad 1972) is the same for both the rhomb-porphyry and the basaltic flows as well as for the 'Oslo-essexites', indicating that the rhomb-porphyries and basic rocks ultimately derived from the same source magma. Thus, continued fracturing of the lithosphere could lead to simultaneous tapping of the two major magmas and to an apparent volcanic bimodality similar to that reported from other rift provinces (McCall & Hornung 1972).

Considering the plutonic rocks, Raade (1973) established two distinct groups on the basis of U-Th distribution. One group, including kjelsåsite, larvikite, lardalite, pulaskite and various syenites, shows low and uniform Th/U ratios and was assumed to have originated directly from a mantle-derived source rock by differentiation. The other group, including nordmarkite, ekerite and biotite-granite, has variable Th/U ratios and was thought to have formed through differentiation at higher levels in the crust. The high and variable Th/U ratios for some of the felsic rocks imply an anatectic origin but could also result from selective loss of U by oxidation due to a relatively higher oxygen fugacity in the upper part of the crust. This is supported by the magmatic evolution in the Finnemarka complex that seems to follow a strongly oxidizing trend (Czamanske & Mihálik 1972), and so the question of whether anatexis operated is still open.

Based on various geological and geochemical data, as well as on the present gravity interpretation, Fig. 82 summarizes a simplified evolutionary model for the Oslo igneous rock series. The hypothetical monzodioritic magma is shown as a separate body at intermediate crustal depths, but could alternatively represent the top portion of a huge, stratified magma chamber in which the basaltic magma constitutes the major, lower portion. It is also conceivable, as suggested for, e.g., the Kenya Rift (McCall & Hornung 1972), that the felsic volcanic and subvolcanic series were derived directly from the subjacent basalts by extreme differentiation aided by volatile concentration in the near-surface chambers, or by liquid immiscibility (Philpotts 1971). However, the copious rhomb-porphyry flows and the monzonitic intrusives known and indicated (Section 6.4) strongly imply the existence of a monzodioritic reservoir.

An elongate, gabbroic intrusive underlies the whole province at depth, as inferred from gravity data. While it was still in a fluid state, crystallization supposedly occurred most rapidly near the relatively cold upper part of the body. The silica-enriched residual liquid was consequently displaced upwards due to buoyancy, collecting in large monzodioritic magma chambers and partly also in more near-surface chambers or cupolas. Continued differentiation processes would operate in the discrete magma chambers quite separated from the basaltic reservoirs, forming a diversity of rock-types. In the basaltic reservoir, the formed crystals would sink and be replaced by relatively undifferentiated material leading to a constant production of monzodioritic magma and to cumulate facies at depth. Occasionally, largely undifferentiated basaltic and monzodioritic material would ascend along the tensional cracks and extrude at the surface. The 'Oslo-essexites' probably derived from the same deep-seated source but underwent rapid ascent and continued differentiation in subvolcanic chambers. Further crystal fractionation and gas

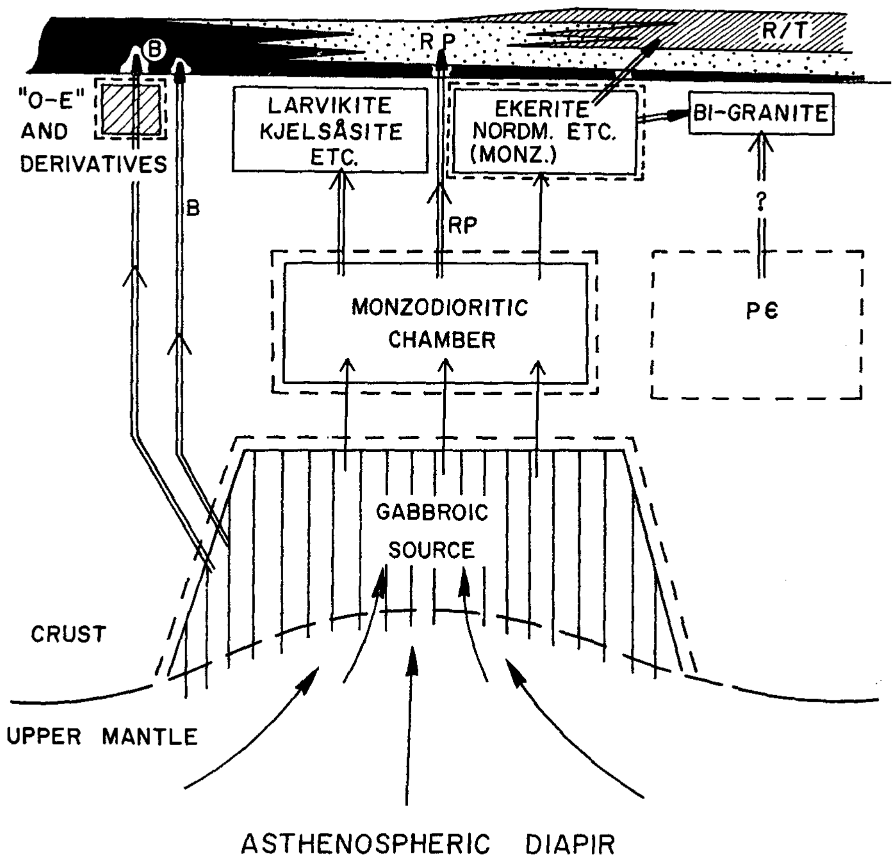


Fig. 82. Schematic evolutionary model for the Oslo Region igneous series. R/T = Rhyolites and trachytes, RP = Rhomb-porphyry, B = Basalt, 'O-E' = 'Oslo-essexites'. Double arrows indicate rapid ascent without significant differentiation, single arrows slower ascent accompanied by differentiation processes. Double-walled boxes signify bodies differentiating *in situ,* the other boxes represent rock bodies principally undifferentiated after emplacement.

transfer in the shallow chambers or cupolas would produce acidic fractions and gas pressure locally exceeding the lithospheric pressure, leading to explosive eruptions of felsic volcanics and to cauldron subsidence. The larger syenitic and granitic batholiths were probably derived from the inferred monzo-dioritic reservoir, rising towards the surface by means of piecemeal, magmatic stoping.

A number of fundamental questions may be raised and studied in the future, such as: how and why did the rise of primordial alkaline magmas start and stop; why did the main episode apparently take place within a short time span in the Permo-Carboniferous; and why was the igneous activity restricted to a geographically small area such as the Oslo Region? Was perhaps the Oslo Region, as indicated by its voluminous igneous products and radiating dikes, underlain by a 'hot spot' or 'mantle plume' (Wilson 1963, 1965, Morgan 1971, 1972, Gass 1972, Burke et al. 1973) which subsequently moved in a relative sense generally to the west? From the postulated Afar and Iceland mantle plumes, Schilling (1973a, b) has demonstrated patterns of geochemical zoning. This offers a quantitative approach that combined with radiometric dating and other geochemical, geophysical and structural evidence might test the validity of such hypotheses in the case of the Oslo Region.

The Oslo Region clearly was the site of igneous activity long before the Permian. The Precambrian mafic intrusions along the west border of the region have already been emphasized. Another example is the Fen complex of alkaline silicate rocks and carbonatites, apparently occurring on the same zone of weakness as the Precambrian intrusives and the western line of 'Oslo-essexites' (Fig. 2). Griffin (1973) showed that lherzolite inclusions within the damkjernite (i.e. kimberlitic, alkaline ultrabasic rock) in the Fen area probably formed at about 10 kb, corresponding approximately to the present crustal base under the 'thinned' Oslo Region. The inclusions may represent fragments of a rising mantle diapir from which the lowest-melting fraction had been extracted as damkjernite, or may be cumulates from a proto-damkjernite magma which separated from its parent material and began to crystallize at depths of about 30–35 km. This might indicate that the Oslo Region was an area of anomalously thinned crust already in Eocambrian/early Cambrian time, thereby representing a favorable site for crustal rifting and volcanism in subsequent geological periods.

Carbonatite and alkaline-ultrabasic complexes in Fennoscandia seem to be related to a well-defined fracture system that also includes the Oslo Graben (Ramberg 1973, fig. 1). Similar features have been observed in a number of other regions, and there seems to be a world-wide association of carbonatites, alkaline rocks and kimberlites with continental rift valleys, swells, step faults, and tensional or fracture zones (Holmes 1965, Verwoerd 1966, Tuttle & Gittins 1966, Kumarapeli & Saull 1966, Gold 1967, Kukharenko 1967, Wyllie 1967, Egorov 1970, Doig 1970). Even though they are generally confined to cratogenic regions, the variety of tectonic environment is evident. It is suggested that the close spatial association of the Fen complex and the Oslo Region is merely a result of tectonic control imposed by a pre-existing, major fracture system.

In conclusion, it is re-emphasized that the major event in the rift formation was the emplacement of large amounts of buried mafic rocks that might have given rise to the Oslo igneous series by crystal fractionation. The mass relations have put constraints on current petrological models for the Oslo Region, stressing the ultimate upper mantle provenance for most of the intrusives. Protrusions of mafic rocks from below seem to represent a key factor in the development of rift systems in general; it has been shown that this also applies to the Paleozoic Oslo Graben.

9. Structural implications

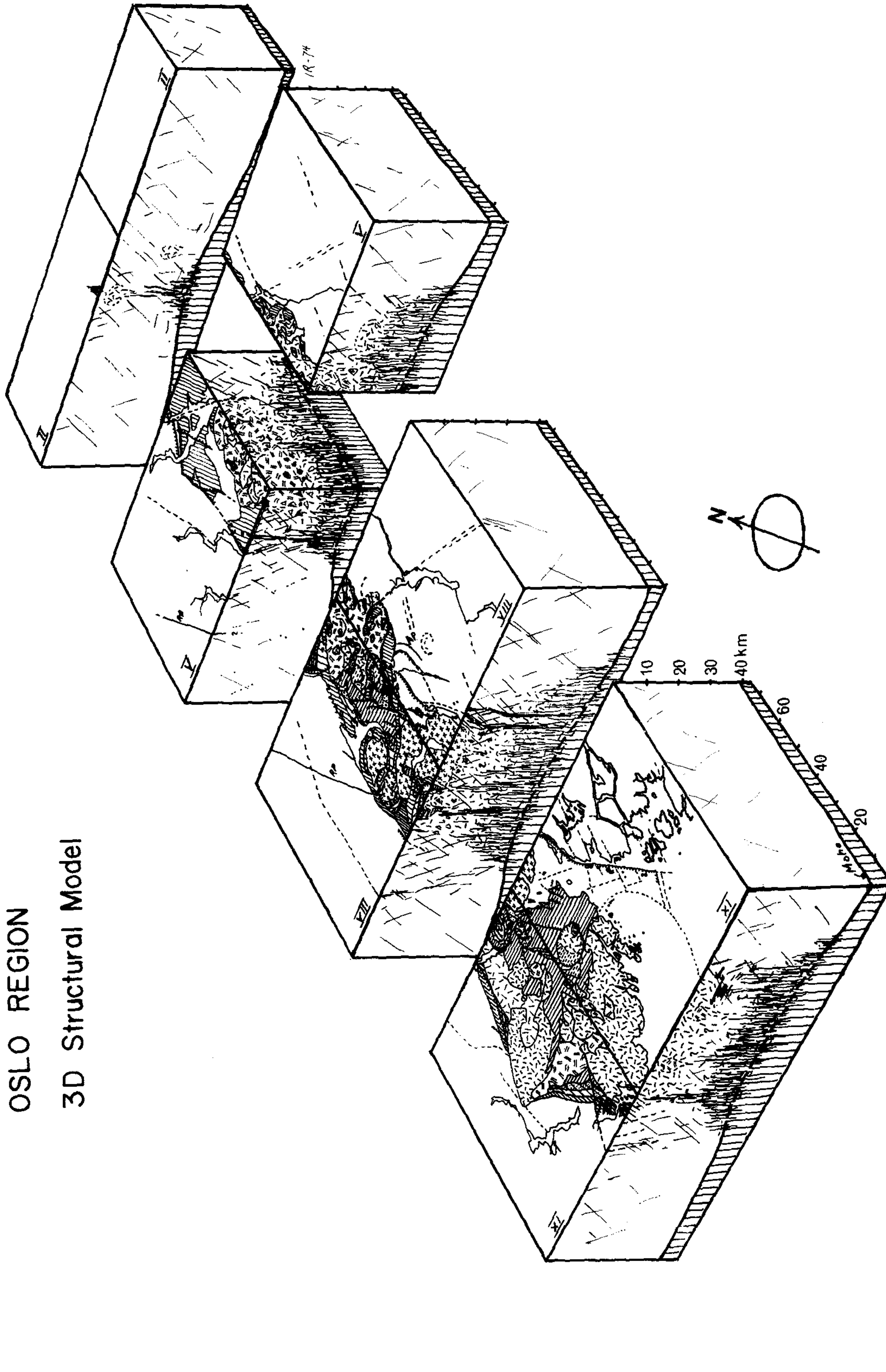
9.1 THREE-DIMENSIONAL MODEL. THE ROOM PROBLEM

A block diagram illustrating the structural features of the Oslo Region as discussed in Chapters 6 & 7 is shown in Fig. 83. The major features are the inferred crustal thinning and the large masses of mafic to ultramafic rocks below the relatively more shallow intermediate to felsic intrusives along the total length of the graben. The deep-seated intrusive body has a rather constant width; it is wider than the graben as defined by surfacial faults and contacts, and much wider than the outcropping igneous suite except in the southernmost (Vestfold) area. It has an overall density only slightly higher than that of the surrounding crustal rocks, and it is envisaged that the inhomogeneous, bulky body grades upward into rocks of more intermediate composition. At the surface, the area north of Oslo differs significantly from the southern part by having large amounts of syenitic rocks. These rocks occupy most of the upper third of the crustal section, the middle part probably consisting largely of monzodioritic intrusives and pre-existing crustal rocks. To the south, syenitic rocks are much more scarce, being confined to cauldron complexes and to rather small occurrences in between the monzonitic masses. Shallow mafic intrusions also occur while the volume and vertical distribution of the large monzonitic batholiths are hypothetical.

Following the interpretation and discussion in the previous chapters, Fig. 83 presents us with a room problem: we have graben subsidence, intrusion by stoping, and rise of mafic material from below, so that the room problem becomes acute. Everything would seem to have been 'squeezed' into the middle crust.

If we disregard magma formation by anatexis of the crustal rocks, then principally four possibilities exist: 1) the crustal rocks were domed or forcefully lifted and removed by subsequent erosion; 2) the pre-existing crustal rocks traded place with the intrusives; 3) the intrusives occupied space made available by extensive rifting and separation of the crustal blocks; and 4) crustal rocks were displaced laterally by means of plastic flow and necking.

Except for minor signs in the northern part (Chapter 2), forceful intrusions are not observed. This is consistent with the notion that at shallow levels the large viscosity contrast between country rock and rising melt makes the latter too weak to push the country rock aside unless by some sort of overhead stoping (H. Ramberg 1970). However, at greater depths with less viscosity contrast the intrusives may act differently, giving rise to some pushing and shouldering of country rocks (Buddington 1959). On a broader scale, there is at present no evidence of crustal arching or doming in the Oslo Region. This may simply be due to the Permian age of the rift, since the postulated low-density mantle diapir and heat input probably caused isostatic lifting and thermal expansion of the crustal rocks (Osmaston 1971). With a possible domal uplift of 2–3 km, as found for instance in the Rhine Graben (Illies 1970) and parts of the East African rifts (Baker et al. 1972), this


OSLO REGION
3D Structural Model
10
20
30
40 km
20
40
60
Moho
N

would still lead to negligible amounts of extension (Freund 1966, Pease 1969) and would hardly provide an explanation for more than a small fraction of the voluminous Oslo intrusives. Hence, the first alternative, forceful intrusion and crustal arching, is not considered an important factor in the emplacement of the igneous suite.

Geological and gravity evidence suggests the existence of a large number of subsided xenoliths within the felsic rocks, and magmatic stoping is generally accepted as the mechanism of emplacement for the majority of the exposed Oslo intrusives (Holtedahl 1943, Oftedahl 1960). It is evident that if *all* the Permian intrusives simply traded place with the pre-existing Precambrian gneisses, etc., then the thickness of the crust below the graben would necessarily have to be considerably increased (accepting that the intrusives are of mantle origin), while the opposite seems to be the case. Also, the xenoliths would not easily descend into the mantle since they supposedly exhibit negative density contrasts with respect to the upper mantle material unless this was totally melted. Thus, while the upper crustal intrusives to a large extent were emplaced by stoping and physical replacement (in the sense of Daly, 1933), this mechanism alone cannot adequately explain the *mis en place* of the entire volume of intrusives, and other factors must have played an important role.

Crustal separation is generally believed to have taken place in continental rifts although rather extreme views and considerable disagreement have been expressed with regard to the actual amount of separation, for instance in the East African rifts (McKenzie et al. 1970, Mohr 1970, Baker & Wohlenberg 1971, Logatchev et al. 1972). Applying revised pole positions, Baker et al. (1972) calculated that the previously used three-plate three-pole models would yield more moderate values of extension. They arrived at a maximum amount of crustal extension across the eastern rift of 12 km in central Kenya and 25 km in central Ethiopia, more in conformity with geophysical and geological data. From the approximately 36 km-wide Rhine Graben, Illies (1967, 1970) has estimated the crustal separation to about 4.8 km, the calculations being based on rather detailed surface maps and knowledge of fault attitude with depth from numerous drill holes.

In the Oslo Region where the major part of the present surface is occupied by intrusive rocks, the amount of lateral spreading or extension can only be estimated roughly. Extension is indicated by features such as normal and oblique slip faults, fractures and dilatational dikes, and possibly also by the presence of batholithic structures.

Fig. 83. Three-dimensional structural model of the Oslo Region, pulled apart at the profiles II, V, VIII, and XI, and also along the central axis between the profiles II and V. Rock symbols as in Fig. 2, with the addition of upper mantle material (regular vertical ruling), deepseated mafic body (irregular vertical lines), Precambrian (and Cambro-Silurian rocks north of the graben) (white with indications of fracture zones and faults). Block diagram drawn from photographs of actual 'blocks' put together on the basis of gravity interpretation.

Dike frequency in the Cambro-Silurian terrain in the vicinity of the city of Oslo (Section 2.5e) makes it reasonable to assume that the dikes may compose about 5 % of the total volume. A majority of the dikes have strikes from NNW to NNE. On the assumption that the estimated dike frequency is representative for the whole graben, the volume increase imposed by the dikes would correspond to an E–W extension of about 1.5–2.5 km.

Oftedahl (1952) has calculated that the systematic tilting of the step faults in Vestfold, having an average dip of about 75°, corresponds to a horizontal extension in the lavas of about 0.45 km from Horten and westward. The major fault along the east side of the Oslofjord has an estimated minimum vertical displacement of about 1 km at Jeløya east of Horten. With an inward dip of about 60° this is equivalent to a horizontal extension of at least 0.6 km. At the west side of the profile, no major faults occur (unless concealed by the intrusives) but minor antithetic and normal faults are found (Oftedahl 1960, Rohr-Torp 1974). If we consider the various contributing effects mentioned, a crustal separation on the average of about 4 km is reasonable, disregarding possible contributions from within the intrusive masses. The indicated, inclusion-free, wedge-shaped, vertical extensions of the intrusion complexes (Chapter 6) may mark zones of pronounced separation. Also, the larger vertical displacement (about 3000 m (?) or more) along the southernmost extension of the Oslofjord fault indicates that the southern part of the graben was subjected to a greater amount of crustal extension than to the north where extension gradually diminished.

The fourth factor is crustal thinning by the process of necking (Artemjev & Artyushkov 1971). In response to extensional forces and the rise of hot mantle material from below, the lithosphere and lower parts of the crust will yield and flow laterally. The rift-forming processes will, in effect, erode the lithosphere and cause the thinning as observed in the Oslo Region and other grabens.

In conclusion, all the four factors discussed may have contributed to the emplacement of the intrusive series, but it is believed that stoping, crustal spreading and necking were the more important ones; they also operated to different degrees at different crustal and lithospheric depths. Diapiric uprise of (partly melted) mantle material and injection of dike swarms into the lower crust, combined with stretching, lead to softening and attenuation of the crust. Plastic deformation and shouldering might have played important roles in the deeper parts while magmatic stoping marshalled the emplacement at shallow levels. Crustal rupture and separation due to a moderate amount of rotation of the blocks on both sides of the graben added to the space available; but it seems that the suggested mechanisms and specifically the minimum value of extension at the surface (ca. 4 km) are not enough to account for the major room problem. A further test for the hypothesis and for the degree of crustal separation that actually occurred might come from a close study of the narrow belts of Cambro-Silurian rocks that extend continuously across the graben floor. Also, more detailed information about Sr isotopic ratios of especially the deep crustal rocks might help decide whether to revive the idea of a significant degree of crustal anatexis at depth, and possibly ease the room problem.

9.2 EVOLUTIONARY SEQUENCE

There is clearly a relationship between intracontinental alkaline magmatism and rifting; but it is not generally clear whether rifting may be an effect of magmatism or vice versa. Similarly, crustal arching is usually but not always associated with rifting, and so the precise relationship is again uncertain. Contrasted evolutionary sequences have been suggested from different rift areas (Illies 1970, Baker et al. 1972, Logatchev et al. 1972, Barker 1975) implying that the characteristic magmatism, arching and graben formation result instead from the same tectonic event. This event could be crustal extension combined with the fairly sudden development and rise of a hot asthenospheric diapir and discharge of basaltic material into the base of the gradually thinned crust.

In the southern part of the Oslo Region, the thick alkali basalt and first rhomb-porphyry flows appear to have preceded major extensional faulting. To the north volcanism began later; here faulting seems to have taken place prior to the extrusion of the thin basal flows. Certainly the major fault along the eastern side of the Oslofjord was cutting through strata of previously deposited and solidified rhomb-porphyry flows, so apparently the initial graben trough formed well after the extrusion of at least the stratigraphically lower basalt and rhomb-porphyry units. Since the various extrusives may have been derived from magma chambers at different crustal levels (but ultimately of a common, upper mantle provenance), the formation of the superficial graben features must largely post-date the mantle diapirism and emplacement of the mantle-derived mafic source in the lower crust.

In summary, evolutionary stages of the Oslo Graben are tentatively sketched in Fig. 84. The graben formation (taphrogenesis) was a unique event and a direct response to the subcrustal swelling and stretching. The diapir rising to the base of the crust gave doming, tensional fractures and a collapse of the graben along the crest. While the overall trend of the graben reflects the position and geometry of the deep-seated diapirism, the emplacement of the more shallow magmas and spatial location of faults in the upper, brittle crust were partly controlled by the pre-existing fracture pattern. The subvolcanic 'Oslo-essexites' and the highly differentiated, larger, felsic plutons appear as first and final phases in the plutonic development, as seen in a warping-rifting-plate tectonic situation. The present Oslo Graben is dissimilar to recent rifts in having a normal heat flow ($\sim$ 1 HFU, Swanberg et al. 1974), no marginal uplift, sharp Moho reflections, and normal upper mantle velocities. These differences may all be attributed to the difference in age; the Oslo Graben may be typical of paleo-rifts where the deep-seated thermal processes culminated long ago. While the Oslo Graben is at the final stage

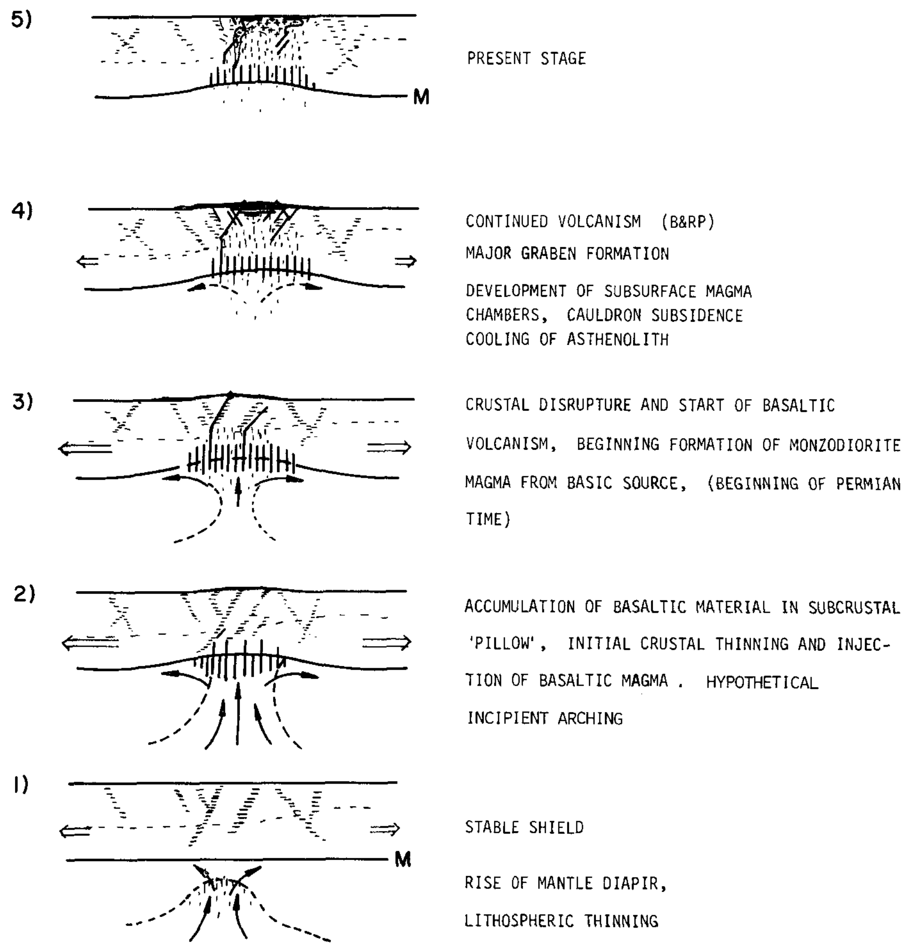


Fig. 84. Evolutionary stages of the Oslo Graben illustrated by a series of cross-sections. Graben formation and volcanism are viewed as secondary effects of the mantle diapirism, crustal thinning and injection of mantle-derived basic material.

of the development (stage 5, Fig. 84), recent continental rifts are in the revolutionary stages (3 to 4) which in the case of the Oslo Graben occurred during early Permian (or Permo-Carboniferous) time. At this time the thermal input evidently was 'switched off' and the graben failed to develop further. However, later tectonic and magmatic events have occurred along the rift zone and may correlate with major phases of tensional deformation of the Laurasian continent.

9.3 REGIONAL EXTENT OF THE RIFT SYSTEM

Since Hans Stille's suggestion in 1925, the Oslo Graben has generally been regarded as the northern extremity of the postulated trans-European 'Mjøsen-Mittelmeer' fracture zone passing through the Rhône Valley, Rhine Graben, Hessen Depression and the area of NNE-trending salt walls within the North German salt dome basin.

No such direct link has been found, on the contrary, the gravity data strongly suggest a southwesterly axial extension of the Oslo Graben below the Skagerrak and the Norwegian Channel between southern Norway and Denmark (Fig. 85). The continuation of the regional gravity high into these areas is compatible with the presence of a rather deepseated mafic body beneath the Skagerrak (Section 7.5). The conclusion is supported by a number of other geophysical and geological observations (Ramberg 1971, Åm 1973, Vokes 1973); important features include:

1) *Permian dikes.* Rhomb-porphyry and diabase dikes commonly occur in the coastal areas from southwestern Norway (Skagerrak) along the Swedish west coast to Scania (Suleng 1919, Ljungner 1931, Hjelmquist 1939, Oftedahl 1952, 1960, Samuelsson 1971) and a Permian age has been suggested for these. This has been supported by paleomagnetic and partly also by K–Ar radiometric dating for a number of dikes in the Kragerø, Arendal and Kristiansand districts along the Skagerrak coast (Storetvedt 1966, 1968, Halvorsen 1970, 1972), while some of the Kragerø dikes appeared to be of Tertiary age (Storetvedt 1968).

2) *Other intrusives.* Ultramafic bodies (Barth 1970, Touret 1970) occur along the 'Great friction breccia' of southern Norway. Preliminary K–Ar dating on micas indicates a late Carboniferous age for one of the occurrences (Touret 1970). Explosion breccias in Bamble occur along a line approximately parallel to the friction breccia and have been related to the Permian volcanic activity in the Oslo Region (Morton et al. 1970).

Gravity and magnetic data indicate the existence of basaltic volcanic centers in the Skagerrak outside Kristiansand. The interpreted directions of magnetization of the magnetic bodies associated with the Kristiansand anomalies indicate that "the plug associated with the negative magnetic anomaly solidified in Permian times, while the positive anomaly is probably associated with a feeder channel for Tertiary volcanism" (Åm 1973, p. 22). An anomaly outside Kragerø is likewise interpreted as a dike-like feeder for Tertiary volcanism.

Marked magnetic anomalies over the Oslo Region are associated with outcropping, or shallow, monzonitic or syenitic batholiths (Åm 1973). They follow the same NNE trend as outlined by the regional gravity high, and a third magnetic anomaly is apparent in the Skagerrak at about 57°40′ N and 8°40′ E. This anomaly may possibly be caused by a similar 'Oslo-batholith'.

3) *Metallogenic province.* The Oslo Region is associated with a characteristic metallogenic province (Vogt 1884 a, b, Goldschmidt 1911). The ore deposits occur mostly within an aureole approximately 25 km wide and especially to the west of the graben (Foslie 1925). Typical epigenetic vein deposits are found along the Skagerrak coast as far south as the Kristiansand area (Fig. 85), implying the presence of plutonic Oslo-type rocks in the Skagerrak. The

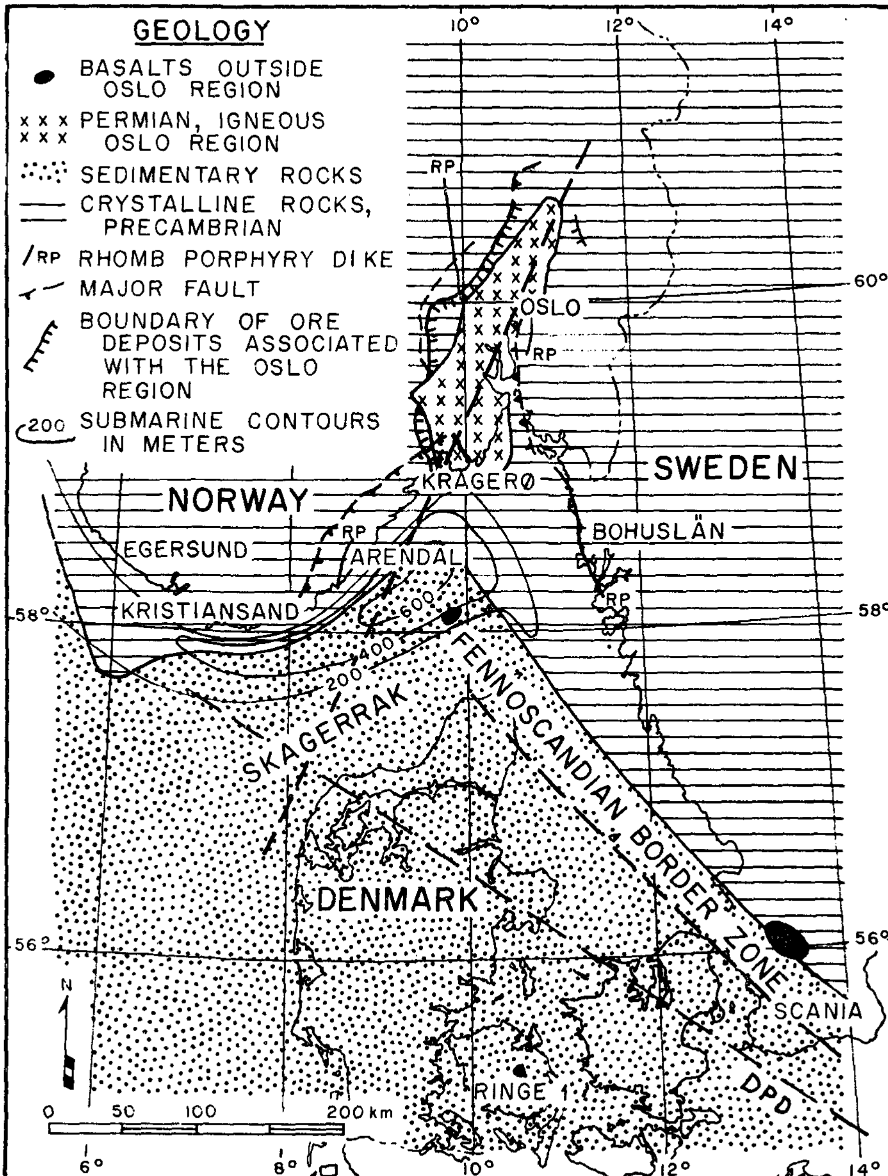


Fig 85. Sketch map of the Oslo Region, Skagerrak and adjacent land areas.

metallogenic province has been related to the major crustal fracturing and magmatism, associated with the initial breaking-up of the Laurasian continent (Vokes 1973).

Similarly, the occurrence of pitchblende and uraniferous hydrocarbon seems to be confined to the surroundings of the Oslo Region, and a hydrothermal origin has been invoked (Dons 1956b). Several occurrences of these minerals near Arendal and Kragerø extend this 'province' to include the Skagerrak coast.

4) *Structural evidence.* The 'Great friction breccia' (Bugge 1928, 1936, 1965), or the Porsgrunn-Kristiansand fault (Morton et al. 1970), and associated zones of breccias and mylonitization continue from the Oslo Region and southwestward, subparallel to the coastline and the extension of the Oslo axial gravity high. The breccia marks a major tectonic zone along which movements have repeatedly taken place from Precambrian to Permian times, and possibly also later. It has been suggested that the friction breccia itself represents the main western fault of the extended Oslo Graben (Bugge 1928, Morton et al. 1970).

In conclusion, the various geophysical and geological evidence strongly indicates that the Oslo Region extends in its axial direction beneath the Skagerrak and subparallel to the southern coastline of Norway, at least as far as south of Kristiansand.

The spatial overlap between the suggested submarine extension of the Oslo Graben and the eastern part of the Norwegian Channel has some interesting aspects. Although a largely tectonic origin for the Norwegian Channel has been invoked by several authors (Holtedahl 1950, 1960, 1963, Pratje 1952, etc.), recent seismic studies have failed to confirm the inferred Tertiary faults and favored the hypothesis that the Channel formed predominantly by the action of glacial erosion (Sellevoll & Aalstad 1971, Talwani & Eldholm 1972, Sellevoll & Sundvor 1974). However, the rather abrupt transition along the Skagerrak coast from Precambrian basement rock to the deep sedimentary basin (Holtedahl 1963, Aric 1968, Sellevoll & Aalstad 1972) has the character of a tectonic contact. Further north, the morphologically sharp fault escarpment along the Oslofjord suggests post-Permian movements. It therefore seems feasible, as argued along somewhat different lines of evidence by Åm (1973), Roberts (1974) and Ramberg & Smithson (1975), that at least the eastern part of the Channel represents the present-day expression of an essentially older structure.

Global implications. The inferred continuation of the Oslo Region beneath the Skagerrak immediately raises the question of its further extension and possible connection with other major fractures or rift zones. A major NW–SE tectonic lineament, in this area commonly referred to as the Danish–Polish Depression, intersects the southern extension of the Oslo Rift offshore in the Skagerrak and can be followed southeastward to the Carpathians (Tectonic map of Eurasia, Yanshin 1966). In Jutland, it is identical to the graben-like Danish Embayment which possibly represented a depression already in Lower Paleozoic time (Bartenstein 1968) (Fig. 4b) and which exhibits maximum accumulation of Zechstein salt (Sanneman 1968) and later sedimentary deposits (Sorgenfrei 1969). The embayment is flanked by structural highs on both sides, the Ringkøbing–Fyn high and the Fennoscandian Border Zone (Fig. 85). In Scania, the Fennoscandian Border Zone is marked by NW–SE trending block faults, Precambrian basement with overlying Paleozoic and

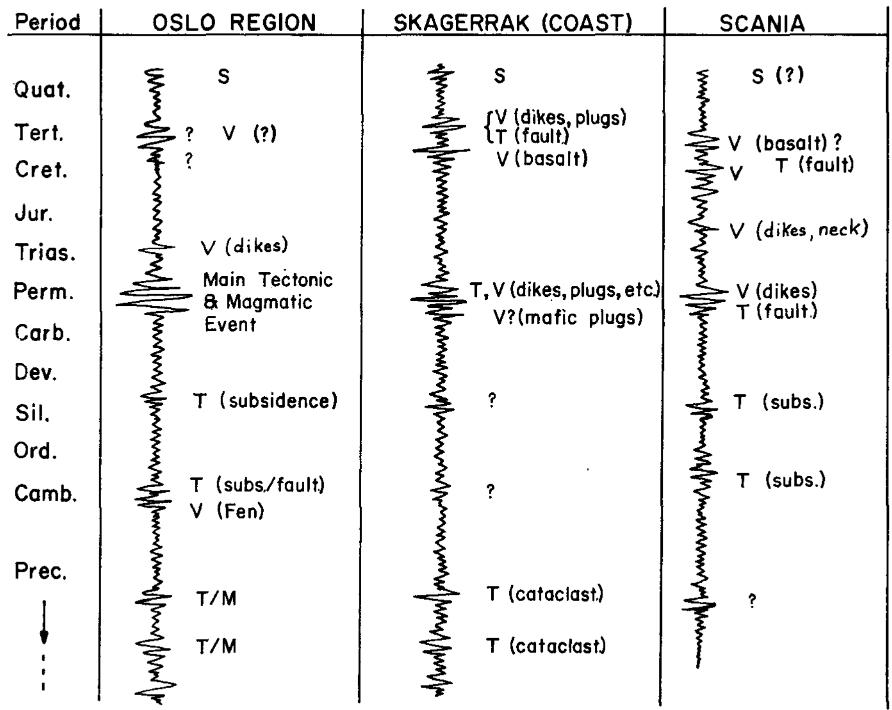


Fig. 86. Tentative sketch illustrating and comparing periods of tectonic and magmatic activity in the Oslo Region, Skagerrak and Scania. Number of peaks has no quantitative meaning; the figure is purely qualitative. S = seismicity, V = volcanism, M = magmatism in general, T = tectonism.

Mesozoic rocks being found at subsequently deeper levels towards the axis of the Danish–Polish Depression. According to Regnéll & Hede (1960), this faulting took place to a large extent in the Permian and was accompanied by volcanism. Troedsson (1932) showed that apart from the Permian, important movements took place also in the Silurian and again repeatedly in the Cretaceous at least up to Lower Tertiary.

Recent radiometric (K/Ar) datings indicate that several of the basalt necks are of Lower and Upper Cretaceous age (Printzlau & Larsen 1972) while other studies from one of the same localities imply a Jurassic age (Klingspor 1973). Priem et al. (1968b) found Permian paleomagnetic pole directions from NW–SE trending diabase dikes in Scania as well as from dikes further north in the Västergötland district, the age being in agreement with K/Ar datings (about 282 m.y.) from the latter region. Thus, faulting and related basaltic activity in the Scania region took place at least in both Permian and Cretaceous times.

The rather synchronous and similar development in the Oslo Region, Skagerrak and Scania (Danish–Polish Depression) is sketchily summarized in Fig. 86. Details remain to be filled in, but the evidence seems sufficient to suggest that the NE-trending Oslo Rift is connected with a NW-trending branch along the Danish–Polish Depression (Ramberg 1971, fig. 7).

The intersection between the Oslo Rift and the Danish–Polish Depression is close to the inferred volcanic centers in the Skagerrak and could represent a triple junction (McKenzie & Morgan 1969) tying in with the systems of fault-bounded throughs or grabens in the North Sea and neighboring areas (Sorgenfrei 1969, Naylor et al. 1974). Alternative connections have been suggested (van Bemmelen 1969, Burke & Dewey 1973) but until now no direct connection between the Oslo Rift system and the North Sea troughs has been proven. Nevertheless, the Oslo Rift would appear to be part of a discontinuous, regional fracture system, forming an extensive zig-zag pattern in western Europe (Fig. 87). Although the crustal rupture and subsidence largely took place in the Permian, the actual directions of the fracture system are remarkably dependent on the pre-existing and predominantly NNW- and NNE-trending tectonic lineaments criss-crossing marginal parts of the Fennoscandian Shield. In the terminology of Burke and co-workers (Burke & Dewey 1973, Burke & Whiteman 1974) the Oslo Graben (like the other arms of the Skagerrak triple junction – the 'Jutland junction' of Burke & Dewey (1973)) may essentially be interpreted as a Permian 'failed arm'. However, the rather voluminous axial intrusives testify that the crustal break-up was accompanied by at least some lateral separation before the process stopped.

The Oslo Graben, as well as other graben areas in Fennoscandia, seems confined to narrow belts along which most of the known carbonatite and alkaline-ultrabasic complexes also occur (Ramberg 1973). The age of some of these complexes such as Fen (Ramberg & Barth 1966) and Alnø (von Eckermann & Wickman 1956), has suggested a correlation with carbonatites of similar age (approx. 565 m.y.) in Canada and Greenland (Doig 1970). The Canadian occurrences are closely associated with the largely Mesozoic St. Lawrence rift system (Kumarapeli & Saull 1966), and Doig (1970, p. 22) suggested that all these rock complexes "belong to a single alkaline rock province, defined by a rift system extending at least from central Canada to eastern Sweden that was active throughout its length about 565 m.y. ago." The positions of the alkaline complexes are shown on a pre-drift reconstruction of the North American and Eurasian continents (Bullard et al. 1965) together with some major fracture or rift zones (Fig. 87). One can see the rather similar position of the Oslo Graben, the Midland Valley, and the (Carboniferous-Triassic) Gulf of Maine rift zone (Ballard & Uchupi 1972) in relation to the eugeosynclinal belt of the Acadian-Caledonian orogeny (as taken from Kay 1969) and the consistent parallelism of many of the Upper Paleozoic grabens with the later North Atlantic spreading axis as well as with the older trends of rifts as indicated by alkaline complexes and Precambrian fracture zones. The evidence leaves little doubt that the Oslo Graben is related to a widespread system of fractures or rifts that was periodically rejuvenated.

The indicated relationship of rifts and older tectonic lineaments is by no means unique for this area and it seems to be generally true that once major fault or fracture zones have been formed they are frequently reactivated

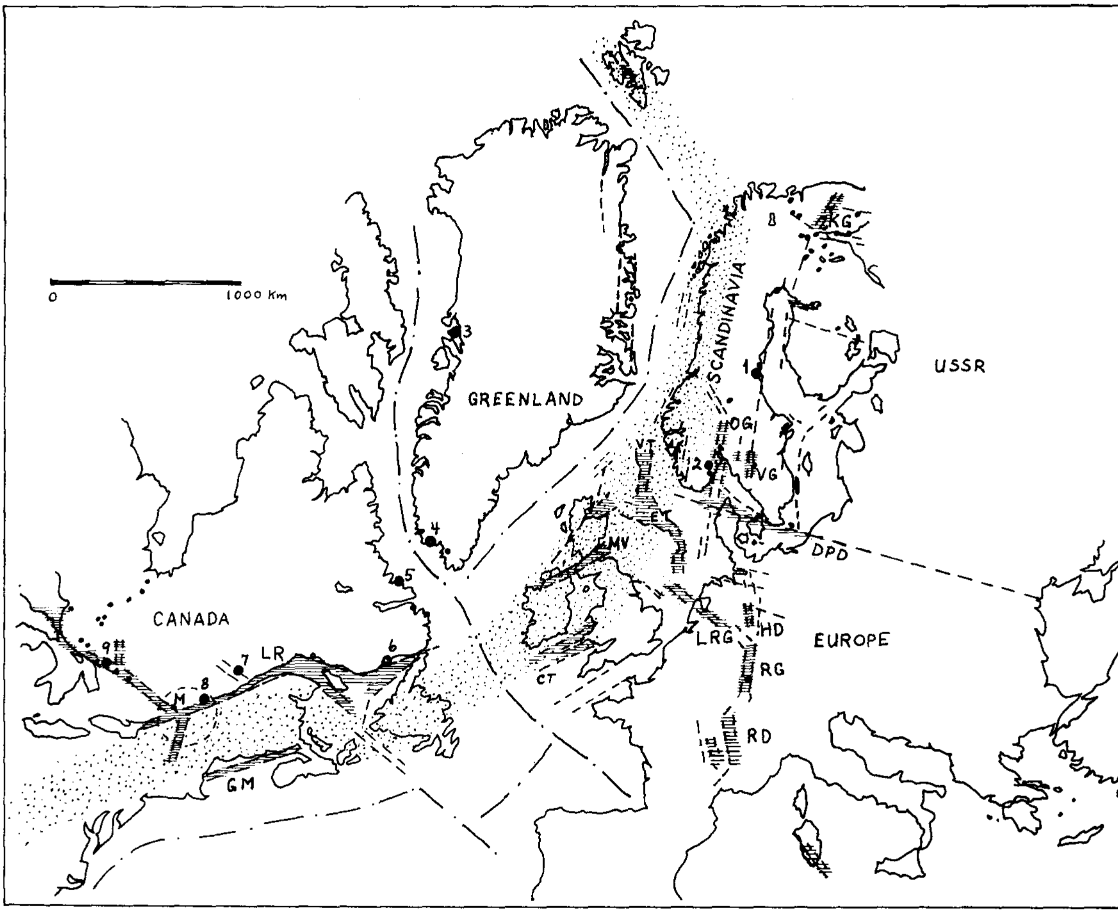


Fig. 87. The Oslo Rift and related features plotted on a pre-drift reconstruction of the North American and Eurasian continents (Bullard et al. 1965). Also shown is the position of the eugeosynclinal belt of the Acadian-Caledonian orogeny (dotted) (Kay 1969) and nine carbonatite and alkaline-ultrabasic complexes of similar (ca. 565 m.y.) ages (large black dots denoted 1 to 9) and other alkaline complexes (small black dots). 1 = Alnø, 2 = Fen, 3 = Umanak, 4 = Ivigtut, 5 = Aillik Bay, 6 = Mutton Bay, 7 = Chicoutimi, 8 = Lac Noire, 9 = Manitou Island, M (in circle) = several complexes in the Monteregian. OG = Oslo Graben, CT = Celtic Sea Trough, DPD = Danish-Polish Depression, ET = Ekofisk Trough, HD = Hessen Depression, KG = Kola 'Graben', LR = St. Lawrence Rift, LRG = Lower Rhine Graben, Netherland Trough, MV = Midland Valley Graben, RD = Rhône Depression/Graben, VG = Vättern Graben, VT = Viking Trough.

(LeBas 1971, Baker et al. 1972, Burke & Dewey 1973, Roberts 1974). The fact that major rift structures in many cases apparently have a very long structural history, does not imply that the hypothesis of plate tectonic is improbable, as maintained for example by Beloussov (1969), but that older rifts, suture lines, etc. constitute zones of weakness that are more vulnerable to repeated rupture than normal continental lithosphere.

The occurrence of several rifts having approximately the same age of foundering as the Oslo Graben (e.g. Midland Valley, Ekofisk Trough, Gulf of Maine) suggests that the Upper Paleozoic represents an important period of crustal stretching and deformation. While the initial opening of the present North Atlantic Ocean has been suggested by some authors to have occurred in Permian (Creer 1965, Heirtzler & Hayes 1967) or Triassic time (Funnel & Smith 1968, Dietz & Holden 1970), most recent studies seem to agree upon a Jurassic opening for the southern North Atlantic, propagating northward and reaching the Norwegian Sea in the beginning of the Tertiary (Hallam 1971, Talwani & Eldholm 1972). Whether or not one considers the Oslo Graben and related features as being related to an early phase of the break-up of the continents, it is evident that the Mesozoic to Cenozoic spreading in the North Atlantic was preceded by an extended period of episodic tensional faulting in the neighboring continental masses.

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Appendix

Characteristics of the rock-types of the main Kjelsåsite-Granite plutonic series (based on Oftedahl (1960) and other sources).

Kjelsåsite is a group name for rocks in the range monzodiorite-syenite. The comprehensive term syenodiorite (Streckeisen 1967) is rightly applied to this compositionally rather variable, coarse-grained rock group. The mineral content of a typical kjelsåsite is plagioclase and alkali feldspar (ratio about 2:1, but highly variable), sporadic quartz, no nepheline, and about 20% of dark minerals like diopsidic augite and possibly hypersthene and olivine (Brøgger 1933a, Barth 1945, Oftedahl 1953, Raade 1973). The areal extent is somewhat uncertain (about 200 km^2) since kjelsåsite frequently grades into more basic rocks (such as the sørkedalite or ordinary gabbros [Sæther 1962, Bose 1969]) or into larvikite from which it is separated by having a plagioclase more basic than An_{30}.

Larvikite is an augite-monzonite with plagioclase/alkali feldspar ratio from 40:60 to 60:40 (Barth 1945). Its mostly bluish schiller effect is caused by cryptoperthitic exsolution lamellae in the original high-temperature ternary feldspars (Oftedahl 1948, 1960, Rosenqvist 1965). Descriptions of characteristic macro- to crypto-perthitic textures in the larvikite and associated pegmatite feldspars have been given by Brøgger (1890), Smith & Muir (1958), Ramberg (1972c) and Brown et al. (1972). Exsolution lamellae of nepheline within the larvikite feldspars have also been reported (Widenfalk 1972, Raade 1973).

The dark to bluish-grey larvikites occur in two separate semicircular areas in the southern part of the Oslo Region (the Larvik and the Skrim plutons), and in somewhat scattered occurrences partly surroundede by younger, acidic intrusives in the northern part (in the Nordmarka and the Hurdalen areas). Apart from orthoclase and albite, larvikite may carry small amounts of quartz or nepheline (especially in the western part of the Larvik pluton, Raade (1973)), and augite, biotite, and sometimes olivine. A reddish variety of larvikite, *tønsbergite* (the red coloration due to finely dispersed iron oxides in the feldspars) is found in the vicinity of Tønsberg in the eastern part of the Larvik pluton. Biotite larvikite occurs NW of Sandefjord, and transitional types to syenites, nepheline-syenites and syenodiorites are known. Larvikite and closely related rocks cover an area of about 1670 km^2 of the Oslo Province.

Within the Larvik plutons occur syenite- and nepheline-syenite pegmatites which have been described in detail by Brøgger (1890). The nepheline-syenite pegmatites occur in the western border zone of the larvikite mass. This zone contains a number of structural features which indicate that the adjacent supracrustals have been tilted, dropped and dragged into the larvikite masses.

Lardalite (nepheline-monzonite to plagifoyaite) and *foyaite-hedrumite* (foyaite to alkali syenite) are normally very coarse-grained rocks that occur with a variety of nepheline-bearing rocks in a small area north of Larvik within the larger larvikite pluton (Brøgger 1890, 1898). Oftedahl (1960) subdivided the nepheline-bearing rocks into four varieties based on feldspar morphology and on dark mineral constituents which may be augite, biotite, hornblende, aegirite-diopside or aegirite. The nepheline content is normally 15–20% but may amount to about 30%.

Akerite has been applied as a name for a number of different rocks of monzonitic composition, some of which are of hybrid origin. The variable use of the term akerite probably reflects the transitional textural and compositional characters between the rock members. Akerite also occurs genetically associated with the Oslo-essexites and a survey of the various modes of occurrence of akerite in the Oslo province, and of contrasted definitions of akerite in earlier literature, has been presented by Oftedahl (1946). Oftedahl proposed to restrict the term akerite to the more medium-grained varieties characterized by rectangular plagioclase crystals, often selvaged by alkali feldspar. Both alkali feldspar and plagioclase occur in considerable amount, quarts up to 10%, and normally two or three of the minerals augite, hornblende, biotite and chlorite.

Pulaskite is another term introduced by Brøgger (1933a), later abandoned by Barth (1945), but reintroduced by Sæther (1962) on the basis of its apparently wide occurrence in the northwestern part of Nordmarka. It is actually a basic syenite and was grouped with the nordmarkites by Barth (1945). Sæther (1962) states that quartz is virtually absent, but that the pulaskite grades into nordmarkite (with 1–10% quartz) or into 'nepheline syenite' when nepheline is present. Sæther regards the pulaskite as a regular member of the deep-seated rock series, possibly modified by assimilation of bedrock fragments. *Hedrumite* is a relatively fine-grained variety of pulaskite, mainly forming a broad ring-dike associated with the Øyangen cauldron. Both pulaskite and hedrumite carry micro-perthitic alkali feldspar; the dark minerals are pyroxenes, alkali amphibole and biotite.

Nordmarkite was originally introduced as a name for the red quartz syenites from Nordmarka north of Oslo (Brøgger 1890). It has, however, a variety of local and transitional types, and various subdivisions have been suggested (Barth 1945, Sæther 1945, 1962, Oftedahl 1948, 1953, McCulloch 1952, Raade 1973). Nordmarkite (*sensu stricto)* is an alkali syenite with rectangular micro- to cryptoperthitic alkali feldspars, aegirite and/or alkali amphibole (Oftedahl 1948). Plagioclase may appear with the alkali amphibole types. Nordmarkite (*sensu lato*), which may also contain ordinary hornblende or biotite, grades into syenitic types with increased plagioclase content (grefsen-syenite), into granitic types with increased quartz content (nordmarkite-ekerite series), and more rarely into true pulaskite or umptekite with a little nepheline instead of quartz. Also the amount of dark minerals can increase with the plagioclase, leading to basic syenites transitional into monzonites. Porphyric types occur both as individual intrusive bodies and as border facies. The typical nordmarkite (*s. s.*), and some of the syenites and granites, characteristically contain miarolitic cavities rimmed by quartz and albite crystals.

Detailed mapping of the 'nordmarkite' subtypes has been carried out only in restricted areas (Sæther 1962, Nystuen 1975a, b). Currently, the term nordmarkite is, therefore, used in a broad sense for the alkali-syenitic to syenitic rocks that predominate in the northern part of the region, and also occur in the southern part in between the larvikite plutons and in some of the cauldron subsidences, altogether covering an area of about 1425 km^2.

Ekerite is a soda granite with mesoperthitic alkali feldspar as the only feldspar, abundant quartz, and aegirite and/or arfvedsonite as the major dark constituents (Dietrich et al. 1965). The largest ekerite massif is in the Eikern area of the southernmost composite batholith. It has clearly cross-cutting contacts relative to the larvikite. In the Sande cauldron subsidence, a peripheral ring of ekerite grades into nordmarkitic and monzonitic rocks in the central part. Ekerite also occurs in a larger massif in the Hurdal area, and as smaller occurrences in Nordmarka. Ekerite differs from nordmarkite in having a higher quartz content (> 10%) but principally no sharp divide exists between the two types.

Biotite granites occur in great quantities in the central part of the region: around the town of Drammen where it forms the subcircular Drammen granite and Finnemarka alkali granite (Brøgger 1933a, Oftedahl 1960, Czamanske 1965, Raade 1973). Also, in the Glitrevann cauldron subsidence (Oftedahl 1953) the central pluton consists of aplitic and porphyritic biotite granite varieties, very similar to some of the border facies found in the Finnemarka and Drammen plutons. Typical biotite granite of the 'Drammen granite type' occurs south of Grua. Several smaller occurrences in the northern part of the province consist of ordinary biotite granites or alkali granites. These varieties are often fine-to medium-grained, frequently showing transitions into the surrounding nordmarkites and ekerites. Albite granites have been reported from the Nordmarka region (Raade 1973). Some of the biotite granites are situated as irregular bodies at the borders of or even outside the main batholiths.

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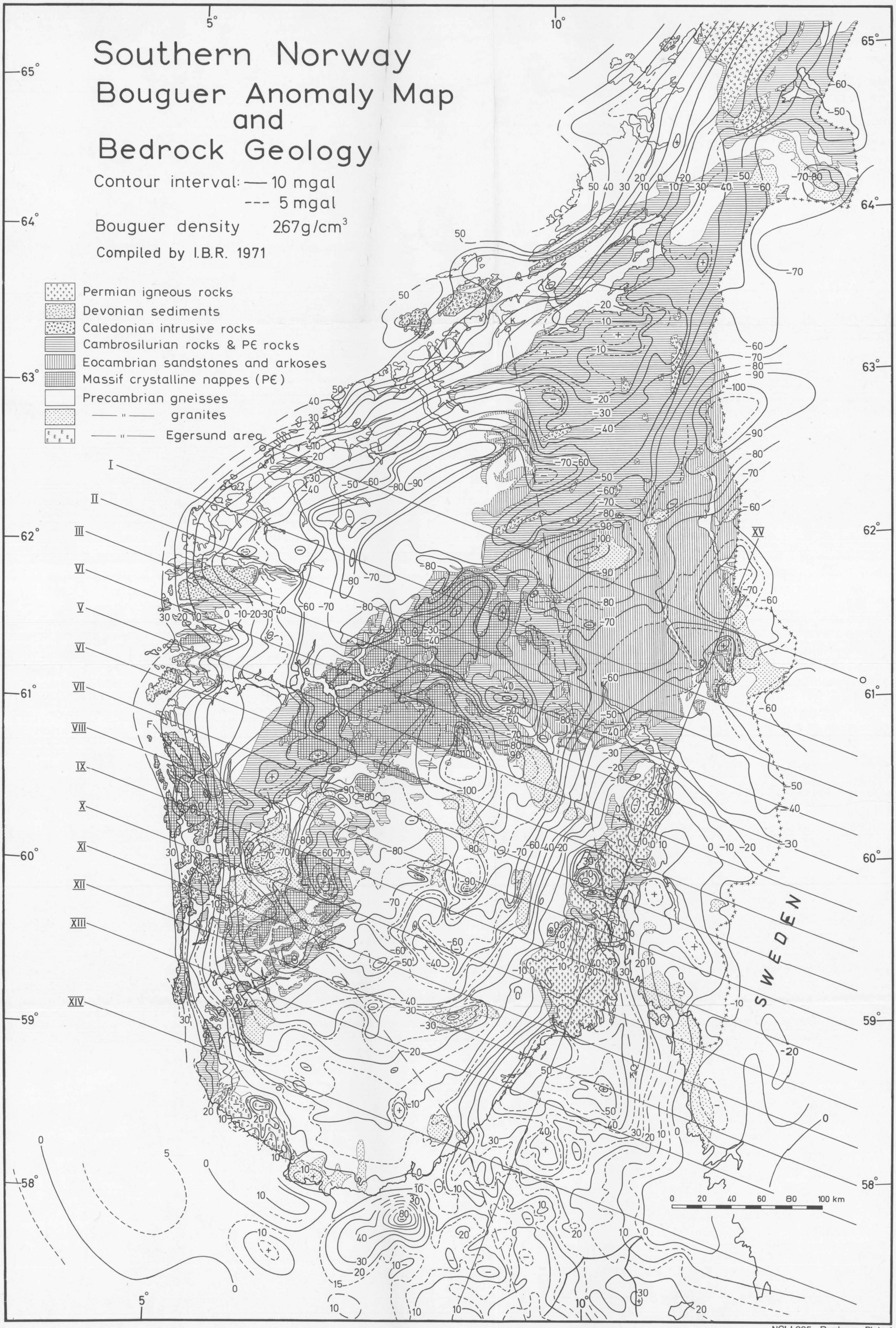
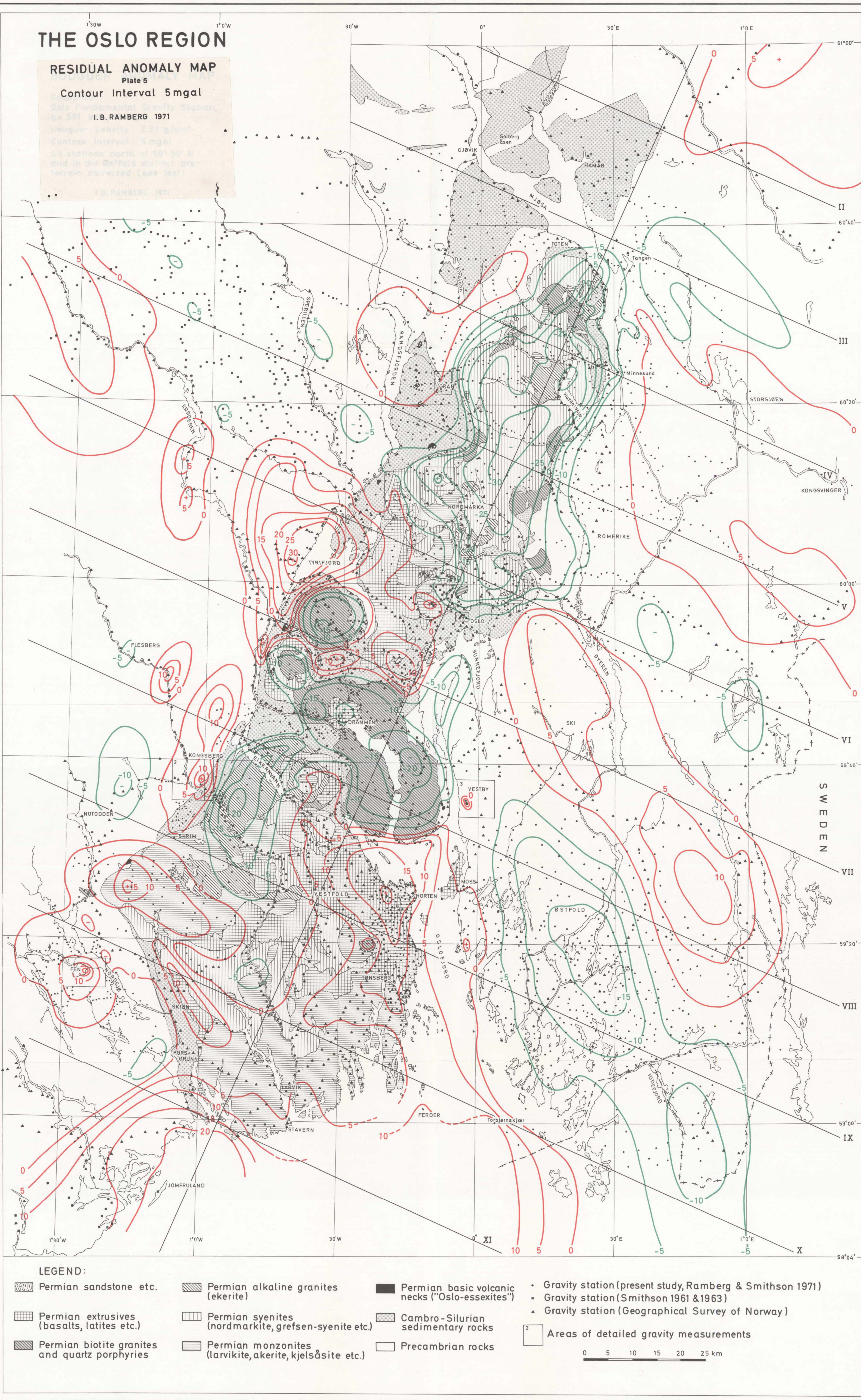
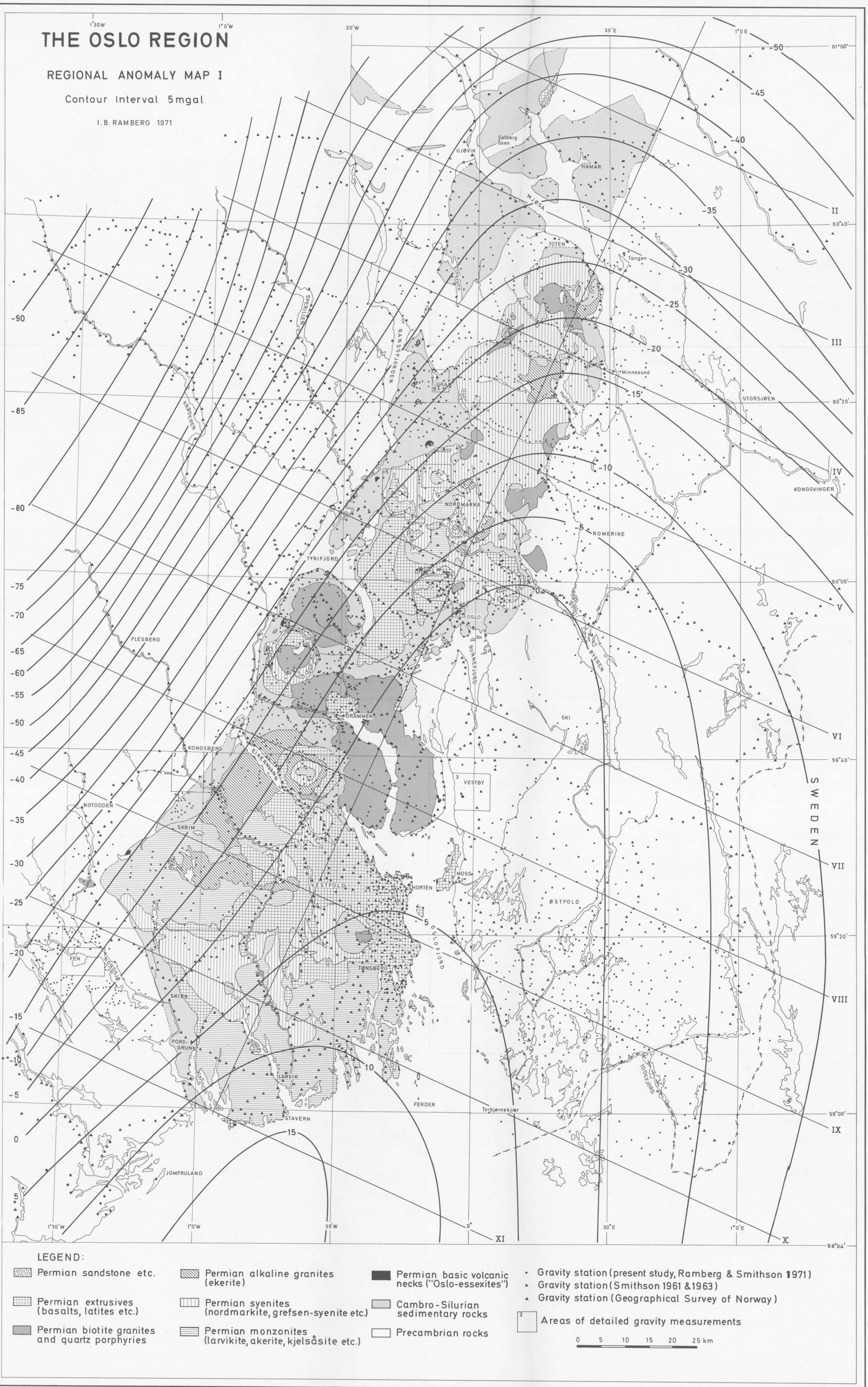

Southern Norway
Bouguer Anomaly Map
and
Bedrock Geology
Contour interval: — 10 mgal
--- 5 mgal
Bouguer density 2.67 g/cm³
Compiled by I.B.R. 1971
Permian igneous rocks
Devonian sediments
Caledonian intrusive rocks
Cambrosilurian rocks & PЄ rocks
Eocambrian sandstones and arkoses
Massif crystalline nappes (PЄ)
Precambrian gneisses
—— " —— granites
—— " —— Egersund area
SWEDEN
0 20 40 60 80 100 km


THE OSLO REGION
RESIDUAL ANOMALY MAP
Plate 5
Contour Interval 5 mgal
I. B. RAMBERG 1971
LEGEND:
Permian sandstone etc.
Permian extrusives (basalts, latites etc.)
Permian biotite granites and quartz porphyries
Permian alkaline granites (ekerite)
Permian syenites (nordmarkite, grefsen-syenite etc.)
Permian monzonites (larvikite, akerite, kjelsåsite etc.)
Permian basic volcanic necks ("Oslo-essexites")
Cambro-Silurian sedimentary rocks
Precambrian rocks
Gravity station (present study, Ramberg & Smithson 1971)
Gravity station (Smithson 1961 & 1963)
Gravity station (Geographical Survey of Norway)
Areas of detailed gravity measurements
0 5 10 15 20 25 km
SWEDEN
GJØVIK
HAMAR
MJØSA
TOTEN
Tangen
Minnesund
STORSJØEN
KONGSVINGER
ROMERIKE
NORDMARKA
TYRIFJORD
OSLO
ØYEREN
SKI
DRAMMEN
KONGSBERG
FLESBERG
NOTODDEN
VESTBY
MOSS
HORTEN
VESTFOLD
ØSTFOLD
TØNSBERG
OSLOFJORD
SKIEN
PORSGRUNN
LARVIK
STAVERN
FERDER
Torbjørnskjær
JOMFRULAND
FEN
IDDEFJORD
RANDSFJORDEN
SPERILLEN
KRØDEREN
EIKEREN
SKRIM
BUNNEFJORD
GRAN
Solberg åsen
1°30'W
1°0'W
30'W
0°
30'E
1°0'E
61°00'
60°40'
60°20'
60°00'
59°40'
59°20'
59°00'
58°04'
II
III
IV
V
VI
VII
VIII
IX
X
XI


THE OSLO REGION
REGIONAL ANOMALY MAP I
Contour Interval 5mgal
I. B. RAMBERG 1971
GJØVIK
HAMAR
MJØSA
TOTEN
Tangen
RANDSFJORDEN
SPERILLEN
KRØDEREN
Minnesund
STORSJØEN
KONGSVINGER
NORDMARKA
ROMERIKE
TYRIFJORD
OSLO
BUNNEFJORD
ØYEREN
FLESBERG
DRAMMEN
SKI
KONGSBERG
EIKEREN
VESTBY
NOTODDEN
SKRIM
MOSS
HORTEN
VESTFOLD
ØSTFOLD
OSLOFJORD
FEN
NORSJØ
TØNSBERG
SKIEN
PORSGRUNN
LARVIK
FERDER
Torbjørnskjær
IDDEFJORD
STAVERN
JOMFRULAND
SWEDEN
LEGEND:
Permian sandstone etc.
Permian extrusives (basalts, latites etc.)
Permian biotite granites and quartz porphyries
Permian alkaline granites (ekerite)
Permian syenites (nordmarkite, grefsen-syenite etc.)
Permian monzonites (larvikite, akerite, kjelsåsite etc.)
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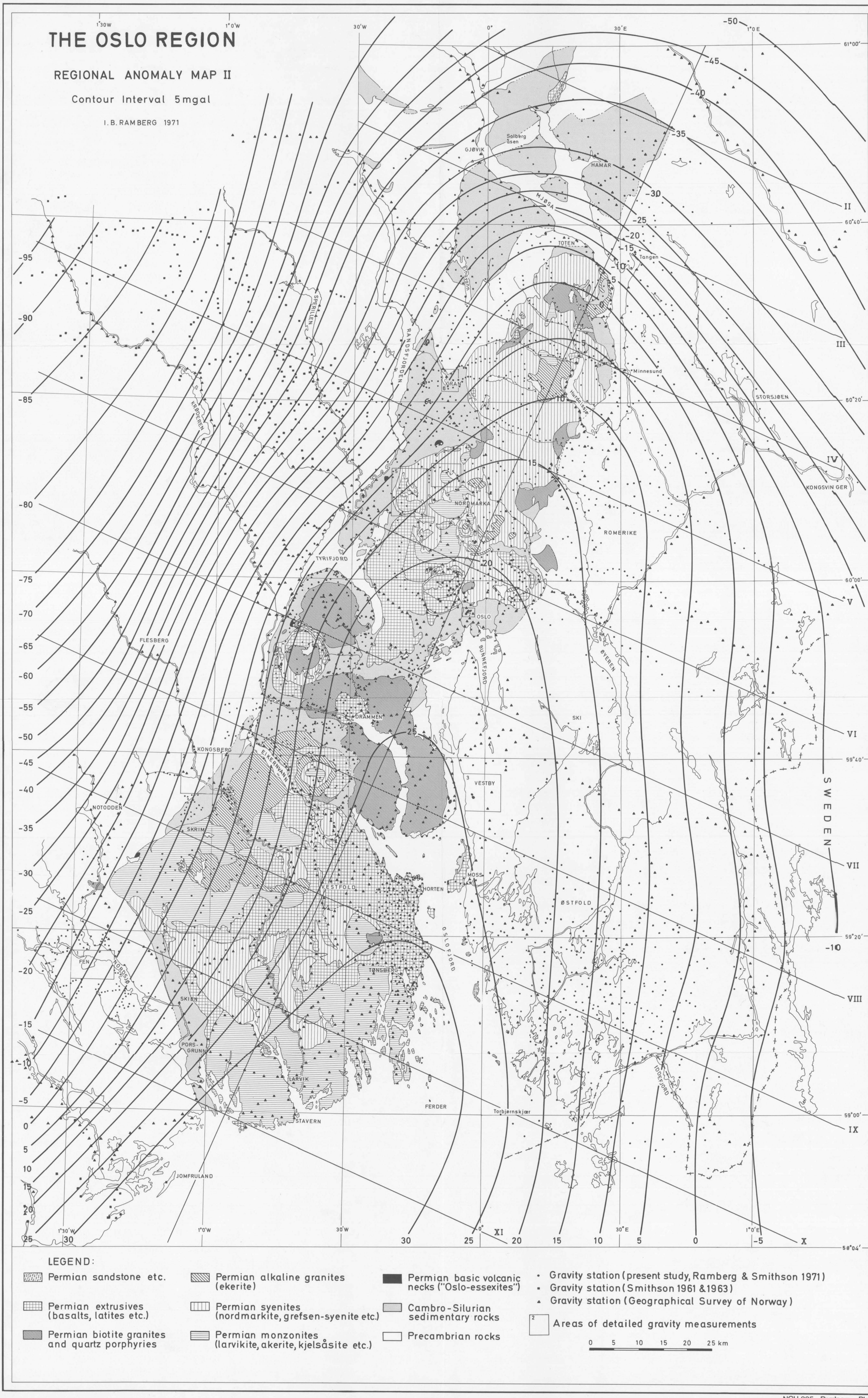
Cambro-Silurian sedimentary rocks
Precambrian rocks
Gravity station (present study, Ramberg & Smithson 1971)
Gravity station (Smithson 1961 & 1963)
Gravity station (Geographical Survey of Norway)
Areas of detailed gravity measurements
0 5 10 15 20 25 km


THE OSLO REGION
REGIONAL ANOMALY MAP II
Contour Interval 5mgal
I.B. RAMBERG 1971
GJØVIK
HAMAR
MJØSA
TOTEN
Tangen
SPERILLEN
RANDSFJORDEN
Minnesund
Hurdalssjø
STORSJØEN
KONGSVINGER
KRØDEREN
NORDMARKA
ROMERIKE
TYRIFJORD
OSLO
BUNNEFJORD
ØYEREN
FLESBERG
SKI
DRAMMEN
KONGSBERG
EIKEREN
VESTBY
SWEDEN
NOTODDEN
SKRIM
MOSS
VESTFOLD
HORTEN
ØSTFOLD
OSLOFJORD
FEN
NORDSJØ
TØNSBERG
SKIEN
PORS-GRUNN
LARVIK
IDDEFJORD
FERDER
Torbjørnskjær
STAVERN
JOMFRULAND
LEGEND:
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Permian alkaline granites (ekerite)
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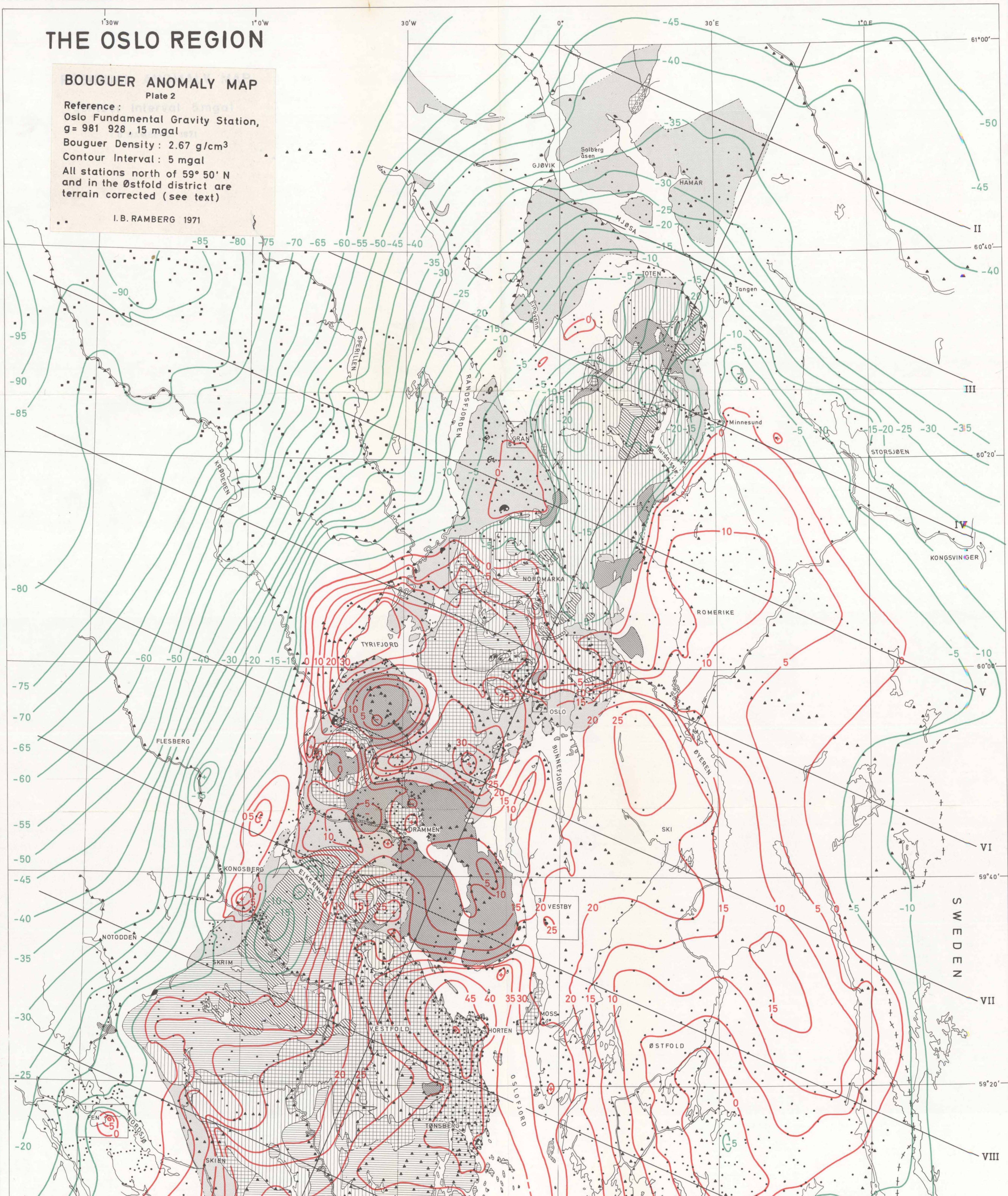
Gravity station (present study, Ramberg & Smithson 1971)
Gravity station (Smithson 1961 & 1963)
Gravity station (Geographical Survey of Norway)
Areas of detailed gravity measurements
0 5 10 15 20 25 km



LEGEND:

- Permian sandstone etc.
- Permian extrusives (basalts, latites etc.)
- Permian biotite granites and quartz porphyries
- Permian alkaline granites (ekerite)
- Permian syenites (nordmarkite, grefsen-syenite etc.)
- Permian monzonites (larvikite, akerite, kjelsåsite etc.)
- Permian basic volcanic necks ("Oslo-essexites")
- Cambro-Silurian sedimentary rocks
- Precambrian rocks
- • Gravity station (present study, Ramberg & Smithson 1971)
- ▪ Gravity station (Smithson 1961 & 1963)
- ▴ Gravity station (Geographical Survey of Norway)
- 2 Areas of detailed gravity measurements

0 5 10 15 20 25 km

Nr. 325

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