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Interpretations of the Gravity Anomaly Map of Finland

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Abstract

An introduction to the Bouguer anomaly map of Finland is presented. Anomaly sources from surface to upper mantle are discussed both qualitatively and quantitatively. Related density variations are reviewed. The article is based on gravity data from the Finnish Geodetic Institute supplemented by petrophysical and gravity data from the Geological Survey of Finland. The Finnish Geodetic Institute has established a national network with an average station interval of 5 km. In the regional gravity measurements of the Geological Survey of Finland average station intervals have varied from 0.4 to 1.0 km. A number of details have been ascertained by gravity profiles with station spacings from 5 to 40 m.

The overall range of Bouguer anomaly in Finland is 120 mGal. The distribution of rock densities within the exposed crust is not unimodal. The main component is the granodioritic one, 2685 - 2720 kg/m^3, consisting of mainly plutonic rocks as in Central Finland Granitoid area, or metamorphosed rocks in schist and migmatite belts. The granitic component, 2600 - 2660 kg/m^3, is common. The high density mode from 2740 to 2920 kg/m^3 comprises high-grade metamorphic, mafic intrusive and volcanic rocks and has the largest standard deviation of the components.

Distinct Bouguer anomaly maxima in Finland are caused by greenstones and other mafic volcanic rocks, ultramafic and mafic intrusions, diabase dykes and sills, carbonatite intrusions, granulite belts and upthrusted blocks from middle crust in general, and schist and migmatite belts when surrounded by a gneissgranite basement or granites.

Distinct Bouguer anomaly minima are caused by granite intrusions, rapakivi granites in particular and potassium rich granitoids in general, thick layers of quartzites in metamorphosed schist belts, granitegneiss basement domes or blocks in metamorphosed volcanosedimentary surroundings, preserved basins of Mesoproterozoic and younger sedimentary rocks, meteorite craters, strongly weathered bedrock and occasionally by thick overburden.

Mass anomalies due to crustal thickness variations are more or less compensated for by mass distributions within the crust. The gravity minimum associated with the land uplift has its source in the lithosphere - upper mantle transition zone.

Key words: gravity anomaly, density, overburden, bedrock, crust, lithosphere

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1. Introduction

The Finnish Geodetic Institute has established a national gravity net with an average station separation of 5 km (*Kiviniemi*, 1980). This net provides valuable information on major crustal structures and geoidal undulations, but it is not detailed enough for geological mapping and raw material exploration. Consequently, the national net is currently being made more dense by the Finnish Geodetic Institute, the Geological Survey of Finland, and the mining companies using various regional, reconnaissance profile and local systematic measurement schemes.

The national gravity net of the Finnish Geodetic Institute has made it possible to publish various gravity anomaly maps covering the entire Finland excluding only some sea areas. The Geological Survey of Finland has cooperated with the Finnish Geodetic Institute in putting gravity data into the hands of geophysicists and geologists and in publishing Bouguer anomaly and its derivative maps with discussions on interpretation principles and geological meaning. Such general reviews have been published by e.g. *Puranen et al.* (1978), *Elo* (1992a & b), and *Elo* (1993a). The data of the Finnish Geodetic Institute has also been an essential element in many international projects such as Nordkalott (*Korhonen*, 1989), Midnorden (*Ruotoistenmäki*, 1993 and 1996), European GeoTraverse (*Freeman et al.*, 1989) and Global GeoTransect (*Elo* and *Virtanen*, 1994e).

In this paper, several major features of the Bouguer anomaly map of Finland are shortly summarized. Selected topics are discussed in some detail. Although this paper follows several previous reviews and bears an interim character, it also incorporates several features that are published for the first time. Because regional gravity surveys in Finland deal mainly with the upper crust, it is natural to start from the surface and proceed as deep into the lithosphere as possible.

2. On the nature of gravity anomaly maps

Gravity measurements are used in studying the figure, composition, and structure of the Earth. Density variations of bedrock and soil in the immediate vicinity of measuring points influence the force of gravity in a discernible way. Quantities describing position, shape, and structure of geological formations can be interpreted from this local variation of gravity. Measuring gravity has thus become an important method in geological mapping and exploration for mineral resources.

Total gravity is influenced primarily by the mass and figure of the Earth, local and regional topography, and centrifugal force due to the Earth's rotation. Measured gravity values are reduced into gravity anomalies in such a way that the features under study stand out as clearly and correctly as possible. As a result, there are several different ways to represent the measured gravity values such as free-air anomaly, Bouguer anomaly, isostatic anomaly and deviation of the geoid from the reference ellipsoid.

Bouguer anomaly maps, which are the most common ones in geological applications, display best the sub-surface density variations of soil and bedrock. The variations of the Bouguer anomaly can be enhanced by calculating 2nd vertical derivative, horizontal gradient or shaded relief maps. These help in perceiving, locating, delineating, and classifying geological formations and structures affecting gravity.

A gravity anomaly is always a sum of several components, which are often difficult to isolate individually. The deeper below the Earth's surface a formation or structure affecting gravity is situated, the smaller in amplitude and broader its contribution to the total anomaly is. Although one can calculate the gravity effect of any mass distribution, one usually cannot unambiguously deduce mass distribution from measured gravity values.

The precision of features in maps depends on the number of observations; maps become more detailed with increasing number of observations. However, contrary to shallow formations and structures, the deeply buried ones can be adequately mapped with relatively few observations, and their contribution cannot be made much clearer by increasing the number of observations beyond a certain limit. To cause discernible gravity variations, a deeply buried formation or structure must be relatively massive.

Interpretation of gravity maps requires understanding of the character of different gravity anomaly types; knowledge of the properties of the Earth, bedrock and soil; quantitative methods to calculate gravity effects of models depicting geological structures and formations; and a modelling framework in which other geological and geophysical data can be accomodated.

Small-scale versions of coloured Bouguer anomaly or its derivative maps have been printed e.g. in *Elo* (1992a) and *Korhonen* (1991). In Figs. 1a through 1c, a different selection of gravity maps, based on the data of the Finnish Geodetic Institute, is presented. The average station interval of the original data is 5 km. Fig. 1d is an index map specifying the locations of some items discussed in the following text. Figs. 1e and 1f provide general geological background. Paper copies of the coloured Bouguer anomaly and its derivative maps on scale 1: 2 000 000 produced by the Finnish Geodetic Institute and the Geological Survey of Finland in 1989 are available from GSF. The 1: 1 000 000 gravity anomaly maps published by *Kiviniemi* (1980) have been sold out. The Finnish Geodetic Institute is preparing a new Bouguer anomaly map on the same scale.

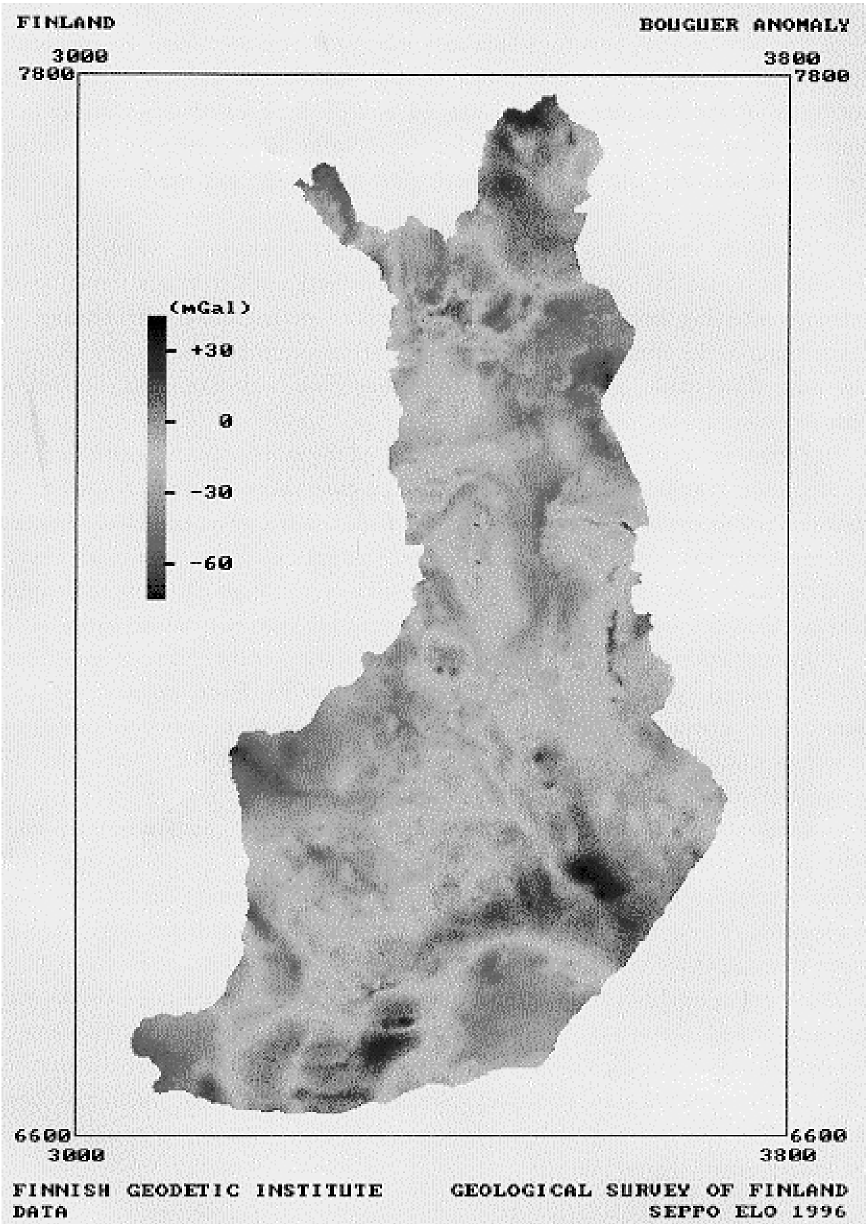

Fig. 1a. A Bouguer anomaly map of Finland. The map was prepared at the Geological Survey of Finland from the data of the Finnish Geodetic Institute.

The geographical coordinates of the corner points are ase follows:

NW corner:	X=7800 km,	Y=3000 km,	LAT=69047′41″N,	LON= 13056′53″E
NE corner:	X=7800 km,	Y=3800 km,	LAT=70006′04″N,	LON= 34054′48″E
SW corner:	X=6600 km,	Y=3000 km,	LAT=59013′02″N,	LON= 18013′46″E
SE corner:	X=6600 km,	Y=3800 km,	LAT=59024′22″N,	LON= 32017′07″E.

Fig. 1b. Bouguer anomaly illuminated from northwest. The map was prepared at the Geological Survey of Finland from the data of the Finnish Geodetic Institute. The geographical coordinates of the corner points are given in the caption of Fig. 1a.

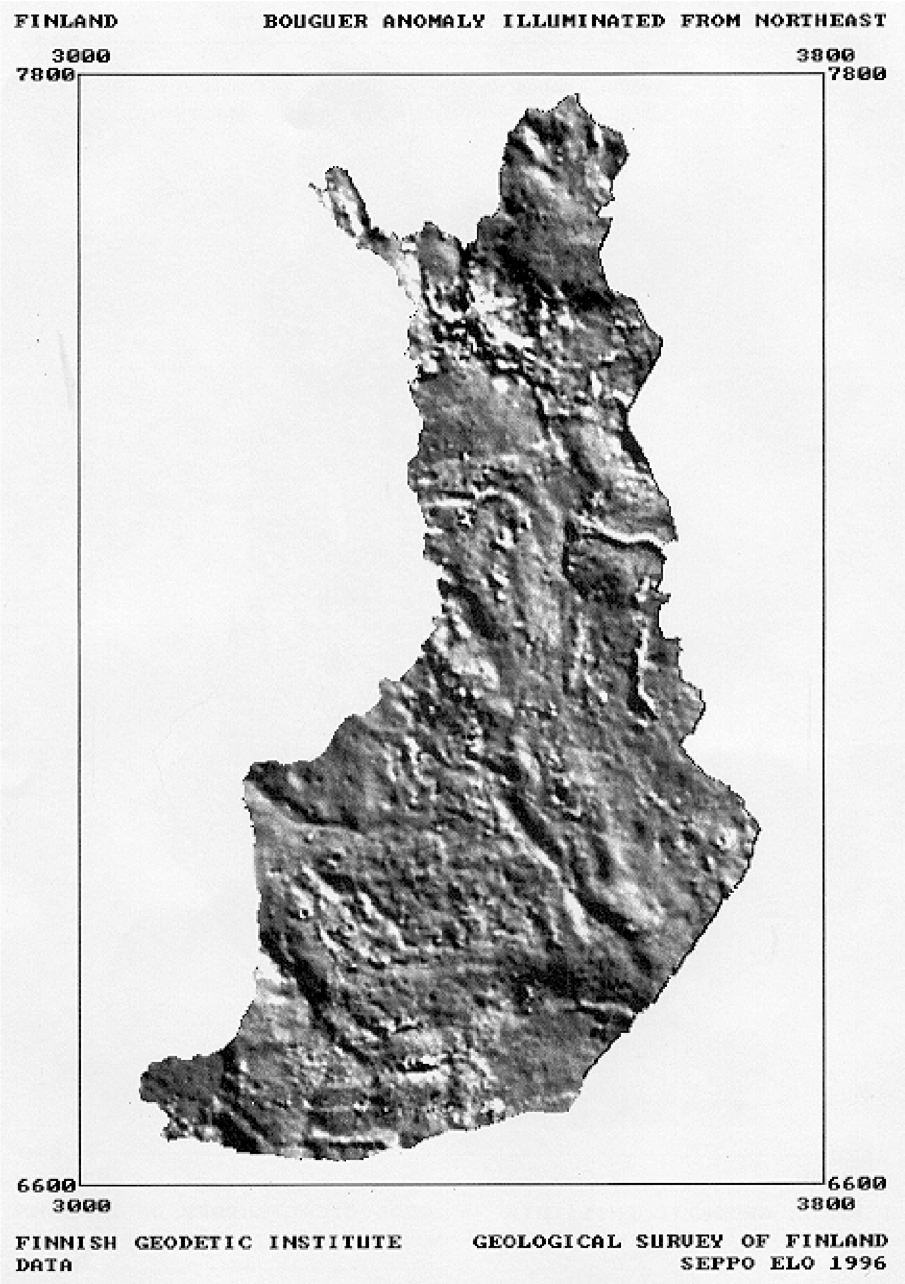

Fig. 1c. Bouguer anomaly illuminated from northeast. The map was prepared at the Geological Survey of Finland from the data of the Finnish Geodetic Institute. The geographical coordinates of the corner points are given in the caption of Fig. 1a.

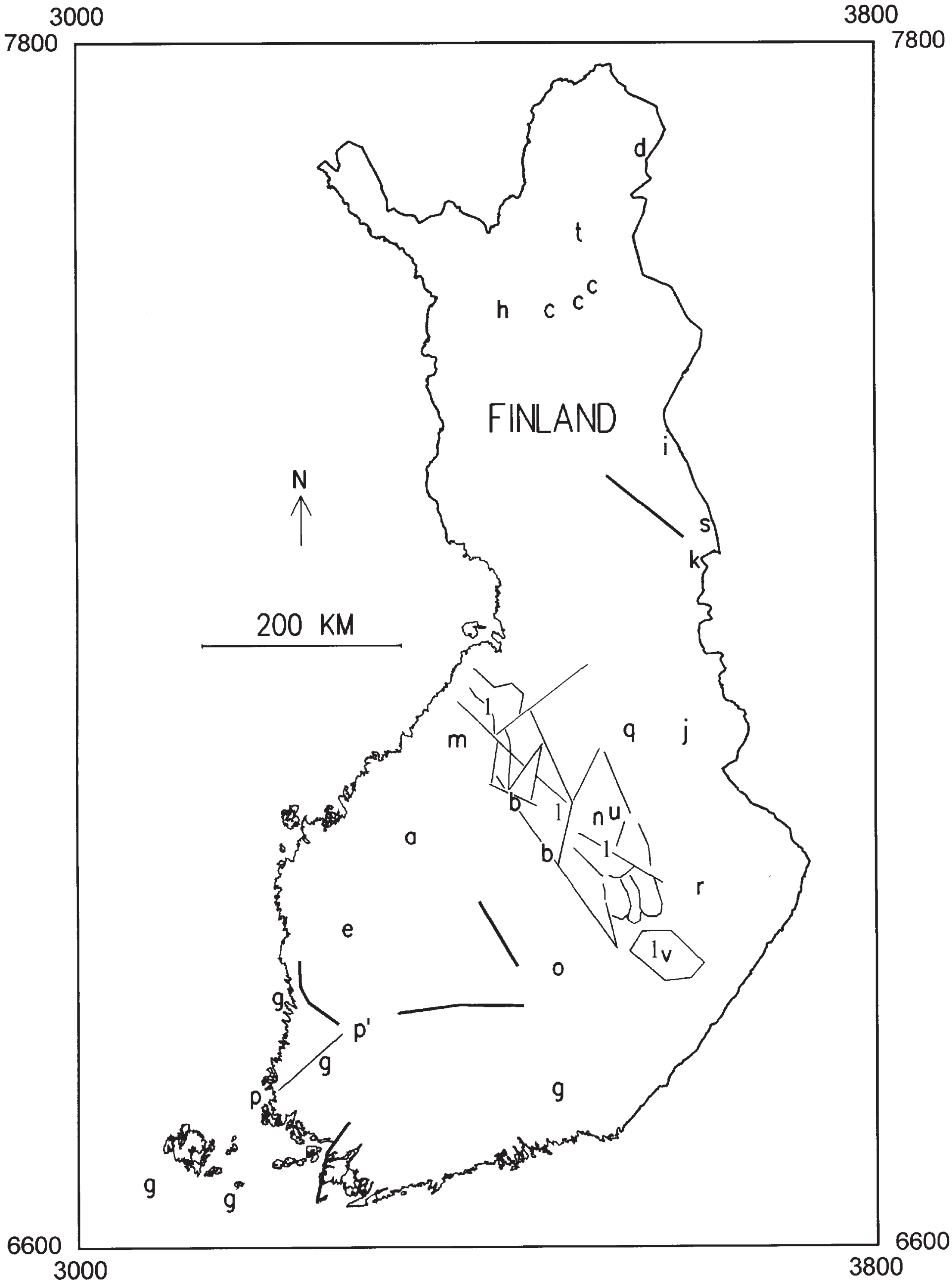

Fig. 1d. Symbols identifying some items referred to in the text. a - Lappajärvi meteorite impact crater. b - Pyhäjärvi granite and Tervo porphyritic granodiorite. c - Pomovaara, Riestovaara and Nattanen granites. d - Vainospää granite. e - Karvia-Jalasjärvi granite. g - rapakivi granites. h - Kittilä greenstone. i - Salla greenstone. j - Kuhmo greenstone. k - Näränkäärä mafic intrusion. l - gravity maxima along the Laatokka-Perämeri zone. m - Ylivieska gabbro. n - Lapinlahti gabbro. o - Kangasniemi gabbro. p-p' interpretation profile along the Global GeoTransect. q - Kainuu schist belt. r - Sotkuma basement block. s - Kuusamo gneiss. t - Lapland granulite. u - Varpaisjärvi granulite. Thin lines outline boundaries, fault zones and shear zones in the Lake Ladoga - Gulf of Bothnia zone. Thick lines represent some very prominent examples of fault and shear zones. The geographical coordinates of the corner points are given in the caption of Fig. 1a.

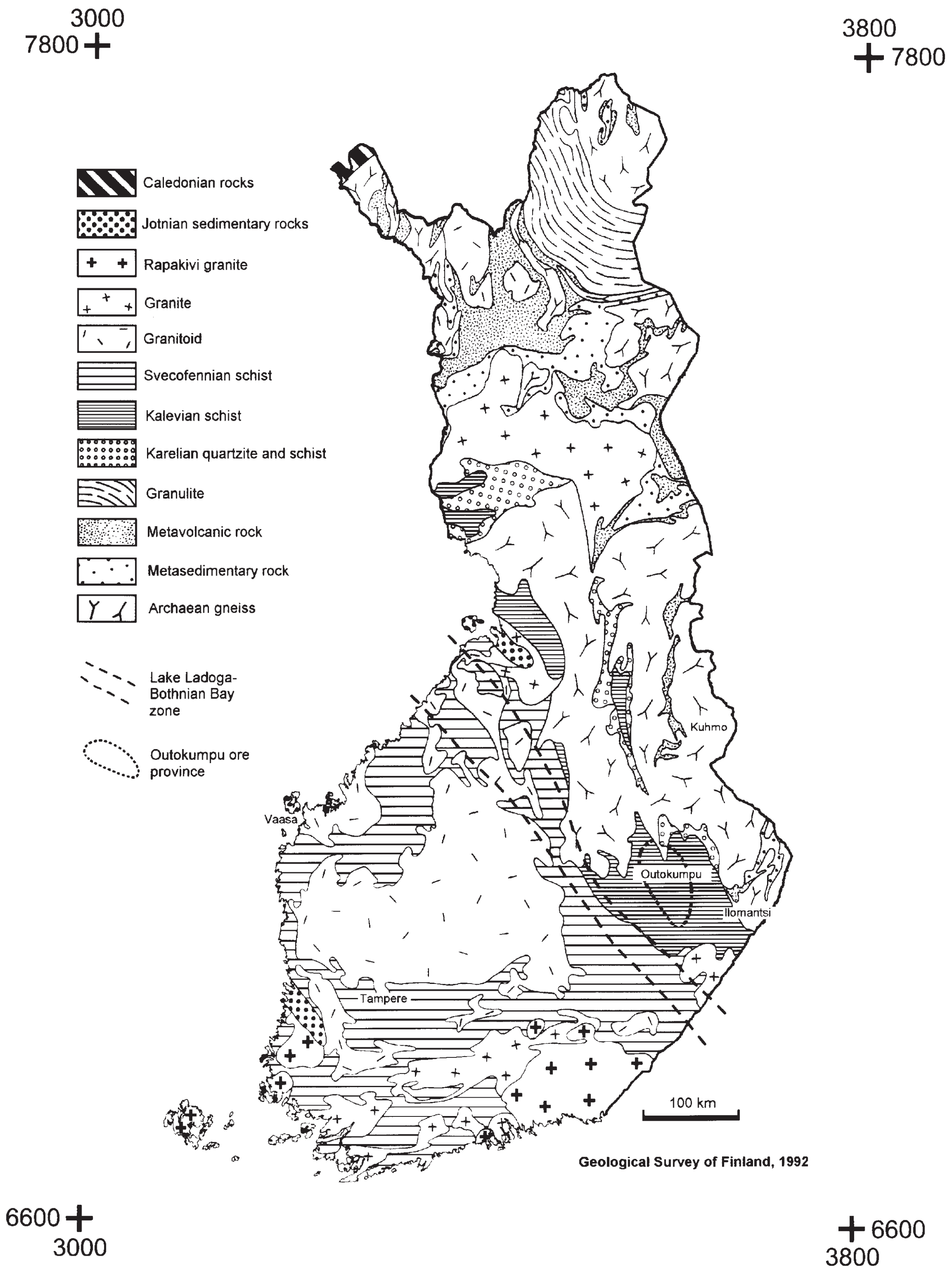

Fig. 1e. Prequaternary rocks of Finland. (*Koljonen*, 1992. The Geochemical Atlas of Finland, Part 2: Till, Appendix 2). The geographical coordinates of the corner points are given in the caption of Fig. 1a.

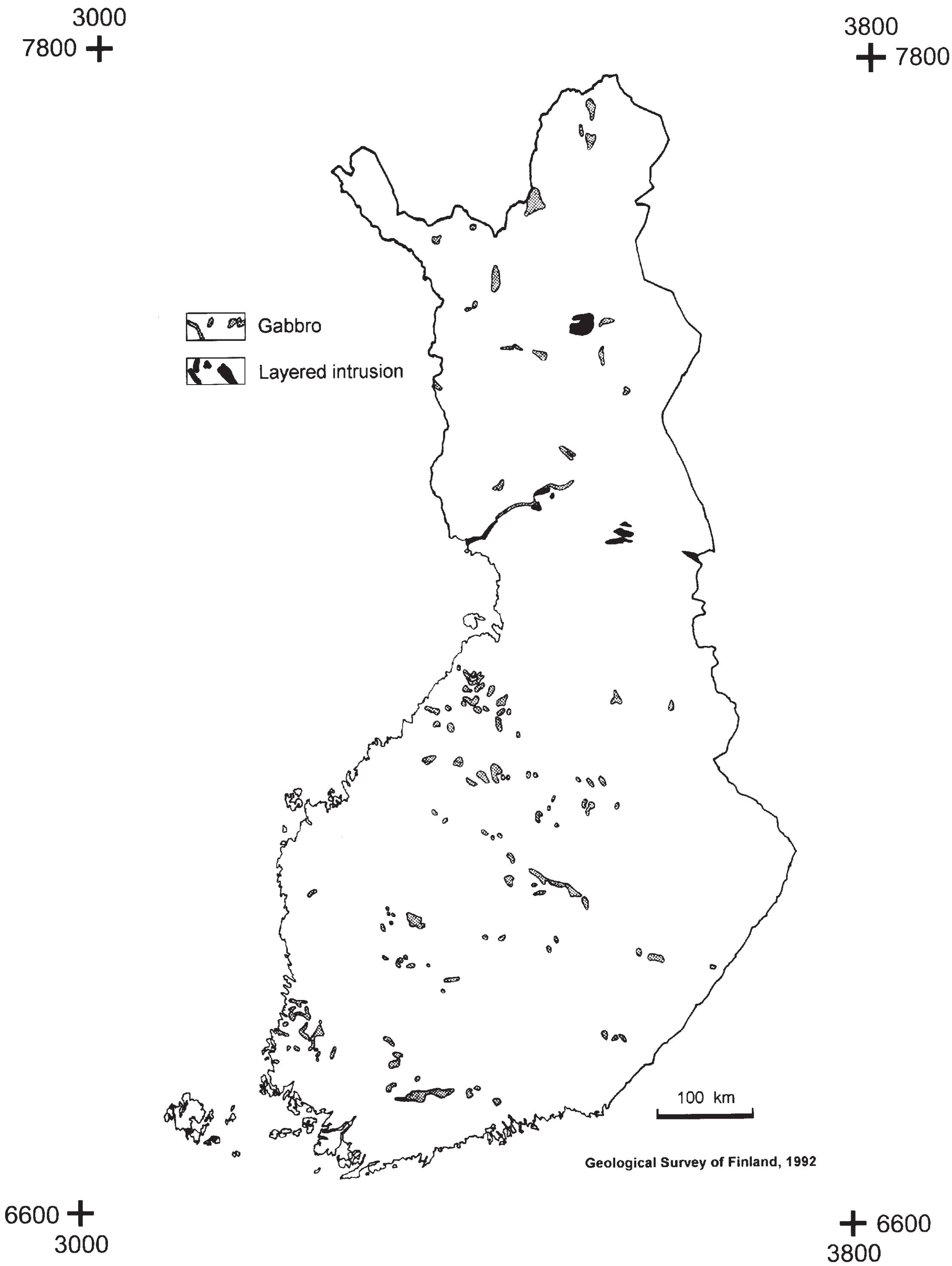

Fig. 1f. Gabbros and layered intrusions (in Finland). (*Koljonen*, 1992. The Geochemical Atlas of Finland, Part 2: Till, Appendix 4). The geographical coordinates of the corner points are given in the caption of Fig. 1a.

3. Density distribution from surface to upper mantle and related gravity anomalies

There are two rather extensive summaries of rock densities in Finland (*Puranen et al.*, 1978; *Lähde*, 1985), and others are being prepared (*Korhonen et al.*, 1991 and 1993). Density in relation to magnetic mass susceptibility was studied by *Puranen* (1989). The following text does not focus on individual rock types, rather a general complementary discussion is attempted based on the data given in Tables 1, 2, 3, 4, and in Figs 2 and 3.

As Isaac Newton already reasoned, the density of the Earth increases towards the centre. However, there are important inhomogeneities, even inversions. The increase certainly is not monotonous. There are transition zones such as the base of volcanosedimentary formations, the base of granitic upper crust, the Moho, mantle-astenosphere-mantle boundaries, and "olivine-spinel" and "spinel-perovskite" transitions. These represent changes in chemical or in mineralogical compositions or in both of them.

Table 1. Selected generalised mineral densities. Unit is kg/m^3.

	Range	Optimum	
Quartz	2500 ... 2700	2650	(*Puranen et al.*, 1978)
Plagioclase	2625 ... 2750	-	"
Alkali feldspar	2560 ... 2630	2580	"
Biotite	2700 ... 3300	2860	"
Muscovite	2770 ... 2880	2800	"
Amphibole	3000 ... 3330	3080	"
Pyroxene	3200 ... 3500	3280	"
Chlorite	2600 ... 3300	2780	"
Epidote	3380 ... 3490	3410	"
Sericite	2760 ... 3100	2850	"
Garnet	3510 ... 4250	3700	"
Calcite	2600 ... 2800	2720	"
Cordierite	2530 ... 2780	2600	"
Olivine	3270 ... 3480	3300	"
Sillimanite	3230 ... 3270	3240	"
Opaques	4600 ... 5200	4760	"
Serpentine	2100 ... 2300	-	(*Suominen*, 1973)
Graphite	2100 ... 2300	-	"

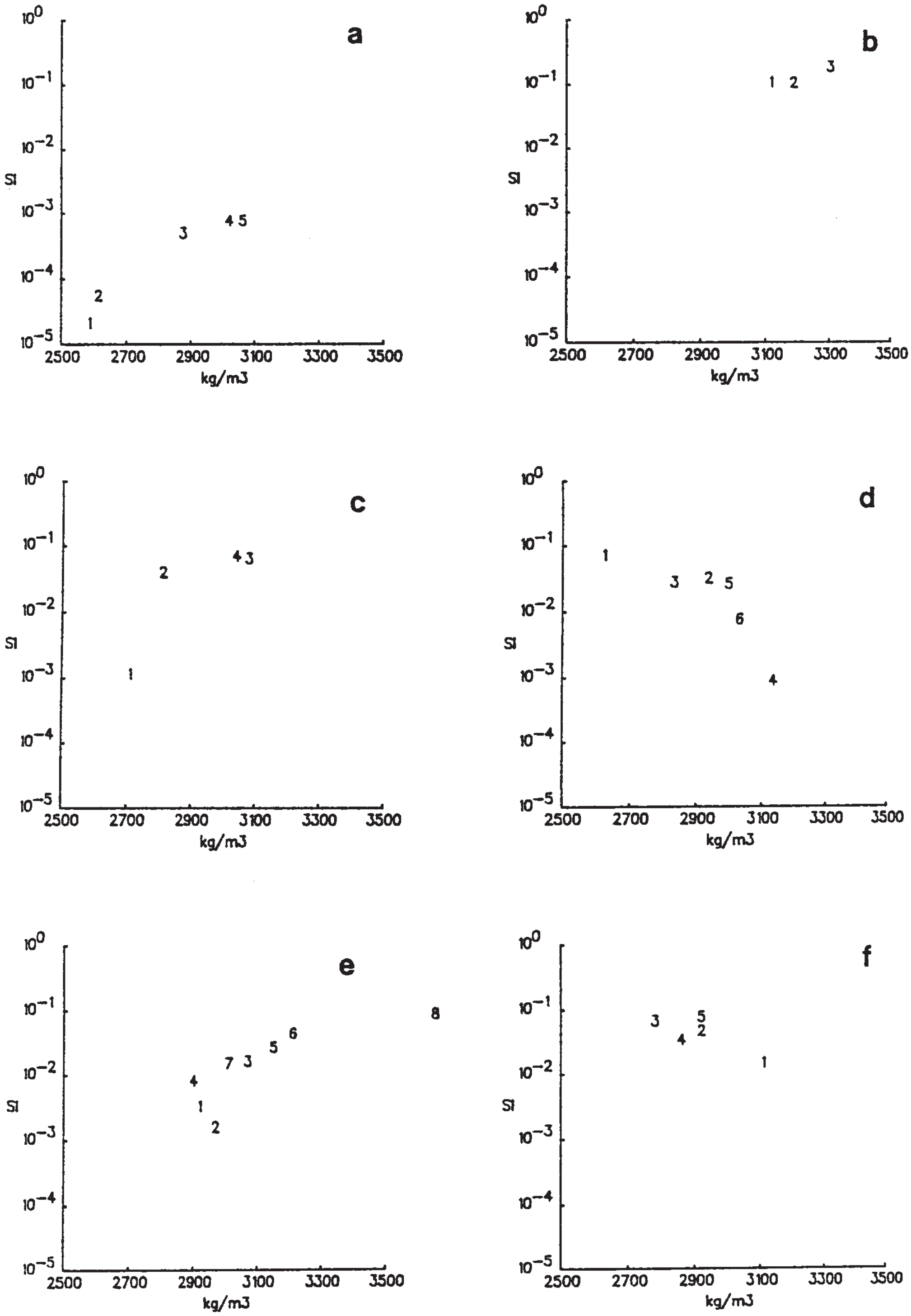

Fig. 2. Density versus magnetic susceptibility diagrams for some mafic/utramafic intrusions. **a)** Tynki suite (*Suominen et al.*, 1991). 1 granite, 2 trondhjemite, 3 diorite, 4 porphyritic diorite, 5 gabbro. **b)** Honkajoki magnetite-apatite gabbro (*Pakarinen*, 1984). 1 all samples, 2 apatite ore, type A, 3 apatite ore, type B. **c)** Lapinlahti gabbro, sampling by geologist *J. Paavola*. 1 anorthosite, 2 anorthosite gabbro, 3 pyroxene gabbro, 4 hornblende gabbro. **d)** Ylivieska gabbro, sampling by geologist *T. Mutanen*. 1 serpentinite, 2 lherzolite, 3 harrisite, 4 - pyroxenite, 5 - pyroxene gabbro, 6 olivine gabbro. **e)** Laukunkangas mafic intrusion (*Niemi*, 1989). 1 - hornblende gabbro, 2 pyroxene gabbro, 3 norite, 4 metaperidotite, 8 Ni-ore. **f)** Stormi mafic intrusion (*Niemi*, 1989). 1 cortlandite, 2 peridotite, 3 metaperidotite, 4 serpentinite, 5 Ni-ore.

Table 2. Selected generalised surficial densities. Unit is kg/m^3.

Air	1.2	*CINA*
New snow	50 ... 70	(*Paterson*, 1981)
Damp new snow	100 ... 200	"
Settled snow	200 ... 300	"
Depth hoar	100 ... 300	"
Wind packed snow	350 ... 400	"
Firn	400 ... 830	"
Very wet snow and firn	700 ... 800	"
Glacier ice	830 ... 920	(*Paterson*, 1981)
Standard water	1000	
Ocean water	1030	(*Stacey*, 1977)
Quaternary overburden in Finland:		
Peat	1000 ... 1100	(*Soveri et al.*, 1972)
Mud	1100 ... 1400	"
Humus	1100 ... 1500	"
Clay	1400 ... 1700	"
Silt	1700 ... 2000	"
Sand	1500 ... 1900	"
Gravel	1600 ... 2000	"
Till	1800 ... 2200	"

Quaternary deposits

The surficial layers in Finland are shallow. The maximum overburden thickness is just over 100 metres, although Quaternary deposits are most frequently only 3 ... 4 metres thick (*Rankama*, 1964). The average density of Quaternary overburden distinctly differs from that of crystalline bedrock. Because overburden is porous the wet and dry densities differ considerably from each other, too. The gravity anomalies associated with 50 to 100 m thick overburden range from -2 to -4 mGal. Combining gravity methods with seismic soundings and drilling for mapping overburden thickness and density variations is recommended. See *(Mattson*, 1991), (*Elo et al.*, 1992e and 1994d), (*Elo*, 1994a and 1995a).

Late Proterozoic and early Phanerozoic sedimentary rocks

The mean densities of non-metamorphic, late Proterozoic and early Phanerozoic sedimentary rocks in Finland range from 2160 to 2630 kg/m^3. There are several locations where preserved sedimentary basins of this type are more than 100 m deep. The maximum thickness is at least 900 metres. In southwestern Finland, the density range of Jotnian sandstones overlaps that of the underlying rapakivi granites. This is one reason why reliable estimates of the thickness of these sandstone formations is difficult to obtain from gravity data alone. What is their maximum thickness in Finland

Table 3. Selected generalised crustal densities. Unit is kg/m3.

Sandstone	1600 ... 2680	(*Jakosky*, 1950)
Limestone, Dolomite	1740 ... 2800	" , (*Puranen et al.*,1978)
Sedimentary rocks in Finland:		
Cambrian	2160 ... 2290	(*Lehtovaara*, 1982)
Neoproterozoic	2280 ... 2420	(*Elo et al.*,1983), (*Elo et al.*, 1993)
Mesoproterozoic	2470 ... 2630	(*Elo*, 1976), (*Kivekäs*, 1994)
Meteorite impact rocks in Finland:		
Impact melts	2530 ... 2600	(*Kukkonen et al.*, 1992)
Suevites	2170 ... 2300	(*Kukkonen et al.*, 1992), (*Kivekäs*, 1993)
Impact breccias	2200 ... 2400	(*Elo*, 1976), (*Kivekäs*, 1993)
Weathered bedrock in Finland		
Kaolinites	1800 ... 2000	(*Järvimäki et al.*, 1980), (*Sarapää*, 1996, (*Kivekäs*, 1993)
Nattanen granite, weathered	2600	(*Kivekäs*, 1993)
Nattanen granite, unweathered	2625	"
Crustal modes in Finland:		
Granitic crust	2610 ... 2660	(*Puranen et al.*, 1978), (*Lähde*, 1985), (*Elo*, 1991), (*this paper*)
Granodioritic crust	2690 ... 2720	(*Lähde*, 1985), (*Elo*, 1991), (*this paper*)
Dioritic to mafic crust	2790 ... 2940	(*Lähde*, 1985), (*Elo et al.*, 1989), (*this paper*)
Metapelites, Sulkava, Finland:		
measured:		
lowest metamorphic grade	2745	(*Korsman*, 1976), (*this paper*)
K-feldspar+sillimanite	2715	"
K-feldspar+cordierite	2730	"
garnet+cordierite+sillimanite	2785	"
calculated from modal compositions:		
lowest metamorphic grade	2795	(*Korsman*, 1976), (*this paper*)
K-feldspar+sillimanite	2820	"
K-feldspar+cordierite	2700	"
garnet+cordierite+sillimanite	2920	"
Archean rocks, Varpaisjärvi, Finland		
High-grade enderbites		
Component 1	2695	(*Paavola*, 1988 & 1991), (*this paper*)
Component 2	2820	"
Low-grade trondhjemitic-tonalitic		
Component 1	2685	(*Paavola*, 1988 & 1991), (*this paper*)
Component 2	2765	"
Lapland, Finland:		
Granulite belt	2745 ... 2750	(*Puranen et al.*, 1978), (*Elo et al.*, 1989)
Greenstone belts	2860 ... 2940	(*Elo et al.*, 1989), (*Elo*, 1991)
Average Finnish bedrock	2695 ... 2720	(*Puranen et al.*, 1978), (*Korhonen*, 1995b)
Upper continental crust	2720	(*Stacey*, 1977)
Lower continental crust	2900	"
Oceanic crust	2850	"

Table 4. Selected average subcrustal densities. Unit is kg/m^3.

Uppermost mantle	3320	(*Stacey*, 1977)
LVZ mantle	3260	"
"Olivine" mantle	3330 ... 3590	"
"Spinel" mantle	3790 ... 4110	"
Uppermost "perovskite" mantle	4460	"
Lowermost mantle	5530	"
Liquid core	10000 ... 12200	"
Solid core	12700 ... 13000	"

is not exactly known, probably well over 1 km. The basins themselves are preserved as down-faulted blocks, pull-apart basins along strike-slip faults, or as eroded valleys or impact craters filled by sedimentation. The gravity anomaly component associated with non-metamorphic sedimentary basins in Finland ranges at least to -6 mGal. Gravity measurements are a major tool in searching for and studying such young sedimentary basins. See (*Kalla*, 1960), (*Puranen*, 1963), *(Elo*, 1976a), (*Lauren et al.*, 1978), (*Elo et al.*, 1983), (*Lehtovaara*, 1992), (*Elo et al.*, 1993c), and (*Kohonen et al.*, 1993).

Meteorite impact structures

Gravity measurements are also an important method in locating and investigating impact craters. Impacting meteorites break, excavate a crater into, and melt existing bedrock and hurl material into air. The impact process is an explosion. In a very short time, a crater is formed and hurled material falls down forming layers that resemble sediments in physical properties. In energetic impacts, the walls of the unstable transient cavity collapse and simultaneously the central region is uplifted. In large multiring craters, the structure is even more complicated. Impact melts solidify into impact lavas. In-situ suevites and impact breccias have been found in several locations in Finland, where they cause gravity minima as low as -6.5 to -10 mGal. The bedrock outside the impact crater is shocked and fractured with a corresponding decrease in density. Because, the structure of a fresh simple meteorite crater is well known, intepretations of present structures make it possible to estimate, for example, how much erosion has taken place since the formation of such craters. Data on the Sääksjärvi crater, southwestern Finland, imply that erosion has been more than 1 km in the last 500-600 million years. See *Elo et al.* 1992c and 1992d; *Kukkonen, et al.*, 1992; *Elo et al.*; 1993c; *Kivekäs*, 1993; *Elo*, 1994b. One of the proven craters, Lappajärvi, (index a in Fig. 1d), causes a gravity mimimum -10 mGal in amplitude and 17 km in diameter. Korhonen (1994) has suggested that two circular structures with outer diameters of 400 km and 190 km, 'Fennia' in southern and 'Marras' in northern Finland may be manifestations of large eroded meteorite craters. No shock effects have yet been found. The verification of such ideas probably takes decades rather than years. Whatever the final verdict will be, it is a fundamental scientific problem to resolve which observations must be explained by impact and which observations by other processes, in other words, what traces, if any, there still remains of ancient, huge meteorite impacts.

Kimberlites

The surface layers of kimberlite diatremes also form in a kind of an explosion, which reduces rock densities. According to *Korhonen* (1995a) half of kimberlite samples he has obtained from Africa, Australia, Siberia, and Eastern Fennoscandia fall out of the density-susceptibility range of ordinary Archaean rocks because of small bulk density.

Weathered bedrock

The exposed bedrock, planed by the last glaciation, is relatively unweathered. Strongly weathered rocks are likely to be found under sedimentary basins, bedrock valleys, or in areas of gentle glacier movement. Kaolinites are examples of very weathered rocks. Good kaolinite deposits in Finland typically have in-situ densities of 1800 - 2000 kg/m^3, are up to 100 meters thick and cause gravity anomalies ranging from -1 to -3 mGal. See (*Järvimäki et al.*, 1980), (*Sarapää*, 1996), and (*Kivekäs*, 1993). Many rocks are moderately weathered, especially in Lapland, where ice movement during the last glaciation was not as forceful as in southern Finland. The average wet bulk densities of weathered and fresh Nattanen granites in central Lapland are respectively 2601 and 2626 kg/m^3 (*Kivekäs*, 1993). The gravity effect due to a 50 m thick weathered layer of such granite would be as small as 0.05 mGal.

Metamorphosed limestones and dolomites

The mean density of metamorphosed limestones and dolomites is about 2800 kg/m^3. These rocks themselves are not likely sources of gravity lows unless surrounded by mafic rocks. In this matter, some erroneous interpretations have been made in the past.

Mafic magmatism

The density of mafic melts depends primarily on their composition, temperature and water content. According to *Best* (1982), the density of basaltic melts ranges from 2600 kg/m^3 (hot, wet, Fe-poor) to 2780 kg/m^3 (cold, dry, Fe-rich). Since the fundamental force driving magmas upwards is buoyancy (e.g. *Walker*, 1989), the high density of mafic magmas means that they tend to be emplaced in denser surroundings than do sialic magmas. These surroundings may develop into a moderate to high metamorphic grade. Stratiform mafic intrusions comprise of layers which have been explained in terms of crystal-liquid fractionation and crystal settling in a convective magma chamber and several magma pulses separated in time (*Best*, 1982). Mafic magmas tend to spread horizontally when they reach crust of sufficiently low density, for example, granitic or young sedimentary rocks, where they form sills or sheet-like bodies which are much larger horizontally than vertically. Obviously solidified mafic bodies have a higher density than their surroundings. In an old metamorphic crust, mafic rocks often occur as fragments. The fragments are tectonically displaced parts of larger bodies. Magmas may also fractionate on their way up from their source region towards the surface. Serpentinization of mafic intrusions reduces density sometimes even to the level or even below the density of their host rocks.

The Kittilä greenstone belt (h in Fig. 1d), whose mean density of 2 940 kg/m^3 is 260 kg/m^3 higher than that of the surrounding bedrock, increases the Bouguer anomaly by 30 mGal. According to a gravity interpretation (*Elo et al.*, 1989), the thickness of the greenstone belt is 6 km. The anomaly maximum due to greenstones is situated within a larger anomaly minimum, which is caused by relatively thick Earth's crust. The situation is similar in the case of the Salla greenstone (i in Fig. 1d), whose interpreted thickness is about 3 km (*Elo*, 1992a). The southern part (j in Fig. 1d) of the Kuhmo greenstone belt causes a 20 mGal positive residual Bouguer anomaly and, according to an interpretation, the belt extends to an average depth of 2.5 km (*Puranen et al.*, 1978).

The Näränkävaara mafic intrusion (k in Fig. 1d) causes a 45 mGal anomaly maximum and, according to an interpretation, extends to a depth of at least 10 km. Such a vertical extent is exceptional for a large mafic body. The southern part of the intrusion contains abundantly magnetite bearing serpentinites, which are clearly visible in aeromagnetic maps but not in gravity anomaly maps because of their low density. Gravity and magnetic anomaly maxima connect the Näränkävaara intrusion to layered intrusions in the west. Näränkäväärä has been intepreted to be their feeder. See (*Ruotsalainen*, 1977) and (*Elo*, 1992b).

An unusually large number of outcropping mafic intrusions occur in the Lake Ladoga - Bothnian Bay zone. In their surroundings, blocks of high-grade metamorphic blocks are also found. The areal extent of the gravity anomaly maxima (l in Fig. 1d) along the zone indicates that more mafic rocks and/or high-grade rocks are hidden below than are found at the surface. The blocks are delimited by fault and shear zones (thin lines in Fig. 1d) in the Proterozoic-Archaean transition zone. The relative contribution of mafic magmatism increases and the relative contribution of high-grade rocks decreases from southeast to northwest.

Petrophysical data given in Fig. 2 is in keeping with the fact that combined interpretation of gravity and magnetic anomalies is a powerful tool in classifying mafic intrusions. For example, magnetite-apatite-ilmenite gabbros are exceptionally dense and magnetic. Some gabbro varieties are dense and weakly ferrimagnetic, some ferrimagnetic but less dense, some moderately dense and moderately ferrimagnetic. Quantitative interpretation of various combinations is an important and unfinished task.

Understanding that it is reasonable to refer to a few examples but not too many, amplitudes of gravity effects and interpreted horizontal cross-sectional areas and vertical thicknesses, are given for the following three intrusions: Ylivieska gabbro (m in Fig. 1d) 17 mGal, 26 ... 29 km^2, 4 km (*Elo*, 1979), Lapinlahti gabbro (n in Fig. 1d) 20 mGal, 33 ... 35 km^2, 2.9 km (*Kukkonen*, 1981), and Kangasniemi gabbro (o in Fig. 1d) 6.5 mGal, 8 ... 11 km^2, and 1.1 km (*Elo*, 1981). The Ylivieska and Kangasniemi gabbros were among the test cases to demonstrate how 2nd vertical derivative maps of Bouguer anomaly can be used to deliniate outcropping intrusive bodies. In the Lapinlahti case, a concentric layered structure of pyroxene gabbros, anorthosites and hornblende gabbros were interpreted from the detailed gravity profiles.

The diabase sills in southwestern Finland constitute a prime example of a situation where mafic magmas, close to the base of sandstones, lost their buoyancy and were forced to propagate laterally to end up as a fragmented sheet of several hundred square kilometres in area. Along the profile p-p' (in Fig. 1d), the SW leg of the Finnish GGT, the sills are 36 km wide, cause wide and flat Bouguer anomaly highs of about 4 to 6 mGal, and according to a gravity interpretation are about 200-400 meters thick on the average (*Elo*, 1994c). In the Archaean crust in Finland, there are a number of areas where mafic dykes clearly contribute to gravity highs.

Granitic magmatism

Granitic magmas ascend easily into and through predominantly mafic crust. Mantle melts, which help in heating Earth's crust are often followed by crustal melts. Alternatively, magma generation is related to crustal thickening followed by orogenic collapse. Deformation often enhances magma segregation. In the long run, the uppermost crust tends to become granitic-granodioritic. *Brown* (1994) has explained how shear zones control migration and emplacement of granitic magmas. Magmas enter shear zones at points of local extension and are expelled from zones of compression. According to Brown, many granites appear to have been emplaced in transient dilatational sites along transpressional strike-slip fault systems undergoing net contractional deformation.

Large gravity mimima associated with late-orogenic, post-orogenic and anorogenic granitic intrusions indicate that they have systematically lower densities by about 70 to 130 kg/m^3 than their surroundings. According to gravity modeling, the vertical thickness of these intrusive bodies often is 4 to 10 km or even more.

Granites or potassium feldspar rich, porphyritic granodiorites cause prominent anomaly minima. Examples (b in Fig. 1d) are associated with about -10 to -15 mGal residual anomalies and controlled by the fault and shear zones in the Lake Ladoga - Gulf of Bothnia zone.

In Lapland, there are several granite intrusions (c and d in Fig. 1d) bearing magnetite and abundantly potassium feldspar, which are clearly visible in both aeromagnetic and gravity maps. The mean density of these intrusions is about

2600 kg/m^3, and they cause anomaly minima of -10 to -20 mGal. For example, according to the gravity modelling, the thickness of the Vainospää granite (d in Fig. 1d) is approximately 6 km (*Elo et al*., 1989).

In western Finland, at the southwestern margin of the Central Finland granitoid area, there is a prominent gravity minimum of -25 mGal associated with alaskitic, coarse-grained, F-rich granite (e in Fig. 1d). According to gravity modelling the thickness of the intrusion is at least 6 to 8 km (*Elo*, 1976b). The intrusion is situated in a long, elongate, NW-SE trending zone of gravity minima.

The extensive rapakivi granite batholiths in southern Finland cause marked Bouguer anomaly minima (f and g in Fig. 1d). The anomaly pattern associated with the Wiborg batholith (f in Fig. 1d) in southeastern Finland is a large minimum superposed on a still larger maximum. The pattern indicates that the crust is relatively thin under the rapakivi granites and that the mean density of the upper crust is abnormally low. The low density is mainly due to great amounts of rapakivi granites, whose mean density at the surface is about 2 630 kg/m^3. Actually, the density distribution is bimodal. The major component (96% of samples) has a mean density of 2625 kg/m^3 and the minor component (4%) a mean density of 2690 kg/m^3. The amplitude of the gravity anomaly pattern is nearly 60 mGal (*Elo* and *Korja*, 1992).

Schist belts and basement gneisses

The average thickness of the Kainuu schist belt (q in Fig. 1d) surrounded by gneissgranites has been interpreted using a linear regression between mean gravity and density values for 100 km^2 areas. To estimate the mean densities, data on individual rock types were weighted by their areal proportions given by geologist M. Havola (personal communication). The average density values range from 2645 to 2725 kg/m^3 and mean Bouguer anomalies from -19.9 to -15.1 mGal. The slopes for two sets of subareas are 53.3 ± 6.5 and 56.7 ± 12.6 leading to depth estimates of 1.8 ± 0.3 and 2.0 ± 0.6 km. The non-linearity between the gravity anomaly and the depth was taken into account.

Archaean basement gneisses outcropping through younger Svecokarelian schists and gneisses are associated with Bouguer anomaly minima. For instance, the Sotkuma basement window (r in Fig. 1d) is associated with a -14 mGal residual anomaly from which the thickness of the surrounding schists has been deduced to be about 6 to 8 km (*Kohonen et al.*, 1991).

Estimation of crustal density modes

The distribution of rock densities within upper crust is not unimodal. *Puranen et al*. (1978) and *Lähde* (1985) made the first comprehensive attempts in Finland to define crustal density modes. They calculated geologically weighted average densities for areas of 10 km times 10 km, and decomposed the histograms of such mean densities into normally distributed components. Fig. 3 is an example of such a decomposition (*Elo*, 1995b) for 18645 rock samples available from Central Finland Granitoid area.

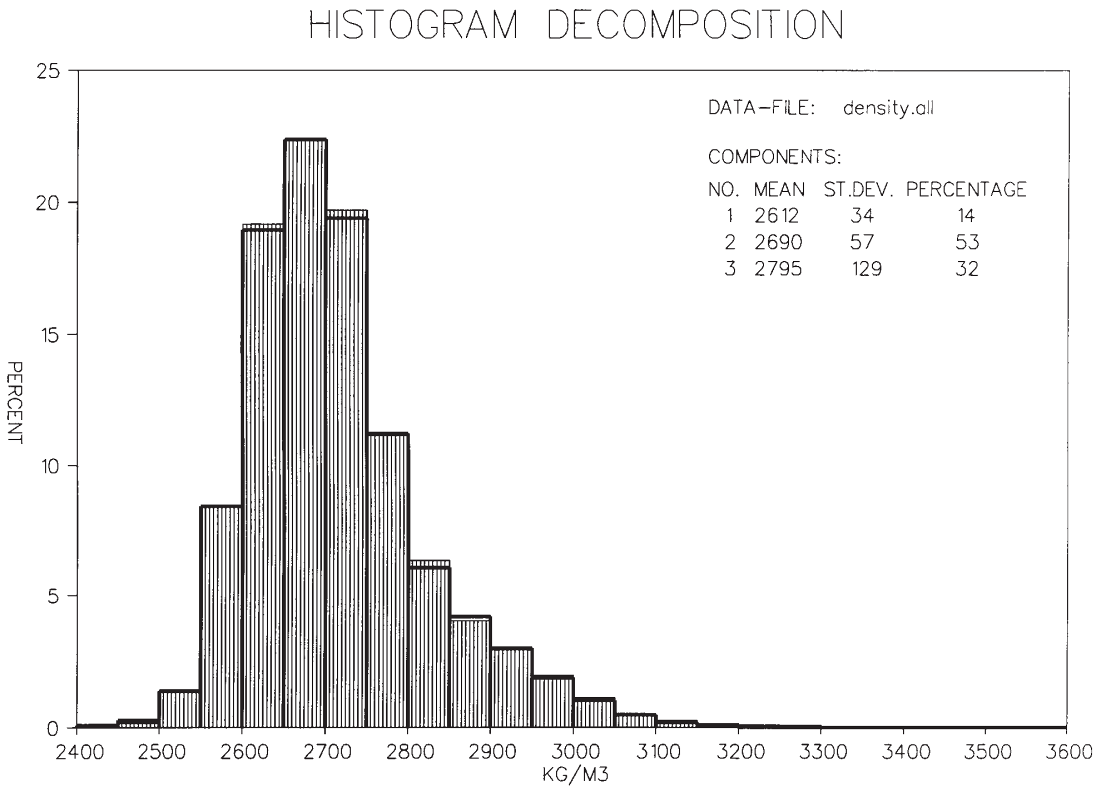

Fig. 3. A decomposition of density histogram into normally distributed components for 18645 rock samples from Central Finland Granitoid area.

The main component in Finland, obtained by various methods, is the granodioritic one, 2685 - 2720 kg/m^3, consisting of mainly plutonic rocks as in Central Finland Granitoid area, or metasedimentary rocks in schist and migmatite belts, or granitegneisses. The granitic and gneissgranite components, 2600 - 2660 kg/m^3, are common. The density distribution of Presvecokarelian, Archaean gneisses is bimodal, which is reflected by Bouguer anomaly maps e.g. in Kuusamo (s in Fig. 1d), where a basement block having a mean density of 2715 kg/m^3 is surrounded by another basement block having a mean density of 2660 kg/m^3 as estimated by combining density data and gravity modeling (*Elo*, 1992a). The high density crustal mode from 2740 to 2920 kg/m^3 comprises high-grade metamorphic, mafic intrusive and volcanic rocks and has the largest standard deviation of the crustal components.

The Lapland granulite belt, whose mean density is 2750 kg/m^3, causes a 30 mGal positive gravity anomaly residual (t in Fig. 1d). The base of the belt dips gently to the northeast and reaches a depth of about 16 km at the steeply dipping northeast contact (*Elo et al.*, 1989). The Archean Varpaisjärvi rocks (u in Fig. 1d) which have experienced high grade metamorphism are partly exposed (*Paavola*, 1988 and 1991). Granulite-facies enderbitic and pyroxene-amphibolitic rocks are tectonically juxtaposed to amphibolite-banded trondhjemitic-tonalitic migmatites. The average magnetic susceptibility of former is about 0.02560 SI and that of the latter is 0.00057 SI. The density distributions of the both are bimodal, the former having normally distributed components 2685 and 2765 kg/m^3 and the latter 2695 and 2820 kg/m^3. The density of amphibolites ranges from 2880 to 2920 kg/m^3. The mean density of abundant diabase dykes is around 2990 kg/m^3. The average crustal density depends on the relative amounts of different rock types and the metamorphic grade and is somewhat higher for the high-grade block than for the low-grade block. Also the Varpaisjärvi low-grade block has a markedly higher mean density than gneissgranite terraines. The density difference between high-grade hypersthene-enderbites and low-grade tonalites is as small as 50-60 kg/m^3. Gravity highs due to uplifted Varpaisjärvi low- and high-grade blocks together are relatively smooth but substantial in amplitude because large volumes of rocks are involved.

Differences in metamorphic grade arise not only due to depth of burial but also due to differences in heat flow. *Korsman* (1976) has studied the influence of a thermal dome in Sulkava metapelites. The progressive metamorphism at first decreases density when K-feldspar and cordierite are introduced. With increasing metamorphism garnet begins to crystallize and density begins to increase. If potassium is introduced from outside some discrepancies may occur as in Table 3 between the measured and calculated densities for K-feldspar+sillimanite metapelites. Potassium feldspars and cordierite decrease, garnet and sillimanite increase density.

Faults and fracture zones

Faults and fractures manifest themselves on Bouguer anomaly maps as: (1) a consistent system of gradients or step-like anomaly patterns forming boundaries between areas of different Bouguer anomaly level, relief or texture; (2) disruptions and/or deflections of Bouguer anomaly trends indicated for example by e.g. abrupt changes in the amplitude of horizontal gradient; (3) linear or lace-like Bouguer anomaly lows caused by fractured and/or weathered fault zones, by abnormally thick soils, by granites intruded into pressure minima or granitization all controlled by deep fault, fracture or shear zones; (4) narrow, linear Bouguer anomaly lows or highs due to low density (e.g. graphites, serpentinites, or pegmatites) or high density (e.g. diabase) rocks intruded into or born in fault and fracture zones. Gravity anomaly patterns often delineate blocks or shear lenses which have experienced differential horizontal and vertical movements. To what degree fracture and shear zones decrease in-situ bedrock density has been difficult to determine exactly. The question deserves further study as do new emerging ideas of complicated upper crustal fragmentation. The more detailed data, the clearer the result. For example, five fault and shear zones are pinpointed by thick lines in Fig. 1d.

Effects of crustal thickness variations on gravity

According to deep seismic soundings, the Proterozoic crust in southern Finland is very thick, while topographic elevations are very low, everywhere below 300 m. The Archean crust in eastern Finland is distinctly thinner than the Proterozoic crust. The crustal thickness in southern Finland ranges from 58-62 to 44-46 km. The thickness of the lower crustal high-velocity layer accounts for most of the variation in total crustal thickness. See (Korja et al., 1994).

The thick crust alone should reduce gravity by 100 to 200 mGal depending on what density contrast is assumed. However, striking Bouguer anomaly highs (v in Fig. 1d), approximately 40-50 mGal in amplitude, coincide with the thickest crust in Proterozoic southern Finland. According to gravity modelling, the lower crust has a mean density somewhere between densities of a normal lower crust and an upper mantle and a considerable mass surplus must exist in the upper and middle crust. It seems that a mass transfer from the upper mantle and lower crust to what is now the upper and middle crust took place in association with crustal thickening or later on. The mass adjustment was not only due to mafic magmatism and thrusting of mid-crustal blocks towards surface but also to the creation of the anomalous lower crust.

Under the Archean Kuhmo greenstone belt (j in Fig. 1d) the crustal thickness changes so that it is about 10 km thinner on the eastern than on the western side (*Yliniemi*, 1987). The crustal thickening to the west is associated with about 17 km thick high-velocity lower crust, which disappears in the east. The 10 km step in the Moho should cause a Bouguer anomaly step of nearly 167 mGal if a density contrast of 400 kg/m^3 is assumed. The Bouguer anomaly on the eastern side of the Kuhmo belt is only about 12 mGal higher than in the western side. Based on the amplitude-gradient ratio most of the observed 12 mGal step has a source in the upper crust. Most probable reason is that the base of gneissgranite rocks is some kilometres closer to the surface on the eastern side than on the western side. The following explanation is offered (Fig. 4) based on the seismic modelling presented by Urmas Luosto at a Finnish GGT workshop in Espoo in 1996. The lower crust has a mean density between densities of a normal lower crust and an upper mantle. The upper part of the high-velocity lower crust in the southwest, being juxtaposed to the normal lower crust in the northeast, represents a mass surplus. The lower part of the high-velocity lower crust, being juxtaposed to the upper mantle, represents a mass deficit. Taken together they more or less preserve isostatic equilibrium and counterbalance each other's gravity effect. The reader should note that geometrically simple, symmetric equal positive and negative masses, one above the other, cause side effects, as shown by the thin curve in Fig. 4. There are various ways to counterbalance side effects. In Fig. 4, the side effect due to the anomalous masses near the Moho is more or less masked by the inclined dip of the upper crustal step. From the point of view of counterbalancing, the dip could as well be the other way. Here, the dip has been selected so that the base of the granitic uppermost crust has been upthrusted from the northeast as suggested by *Luukkonen* (1992).

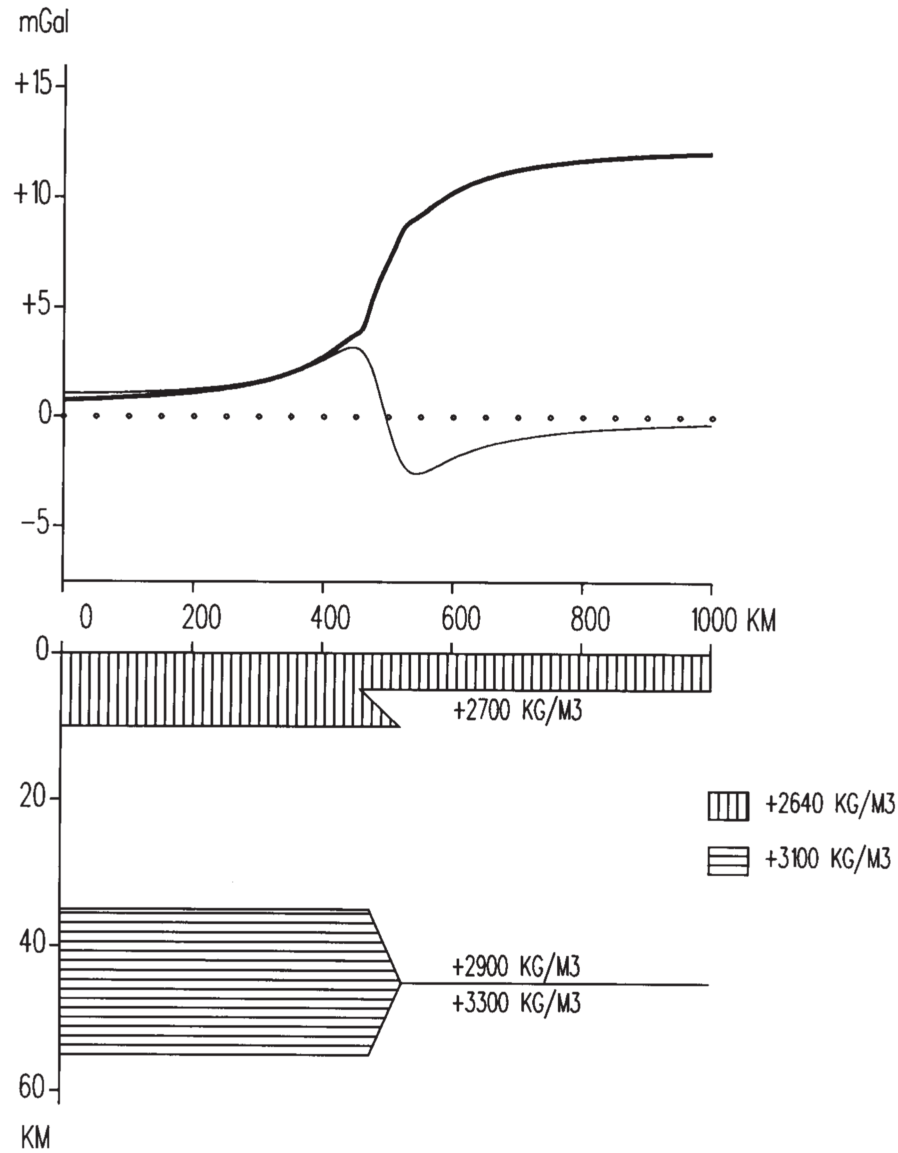

Fig. 4. A schematic 2-D gravity model across the Kuhmo schist belt based on the deep seismic model of U. Luosto. On the left side of the Kuhmo schist belt, there exists an anomalous lower crust with the upper part having a positive and the lower part having a negative density contrast. Under the Kuhmo greenstone belt, the base of the uppermost granitic crust has been upthrusted from the right some kilometers corresponding to a rather sharp increase of gravity anomaly by about 12 mGal from the left to the right. The thin line represents the gravity effect of the anomalous lower crust only, the thick line represents the total gravity effect of the upper and lower crust.

Isostatic equilibrium of Scandinavian mountains

The Bouguer anomaly over the Scandinavian Caledonides are negative (-70 mGal at minimum), while free-air anomalies remain positive (+40 mGal at maximum) as they should when in nearly isostatic equilibrium (*Balling*, 1980). The tail of the Bouguer anomaly and the side minimum of the free-air anomaly extend to Finland. A model calculation demonstrates in Fig. 5 this state of affairs.

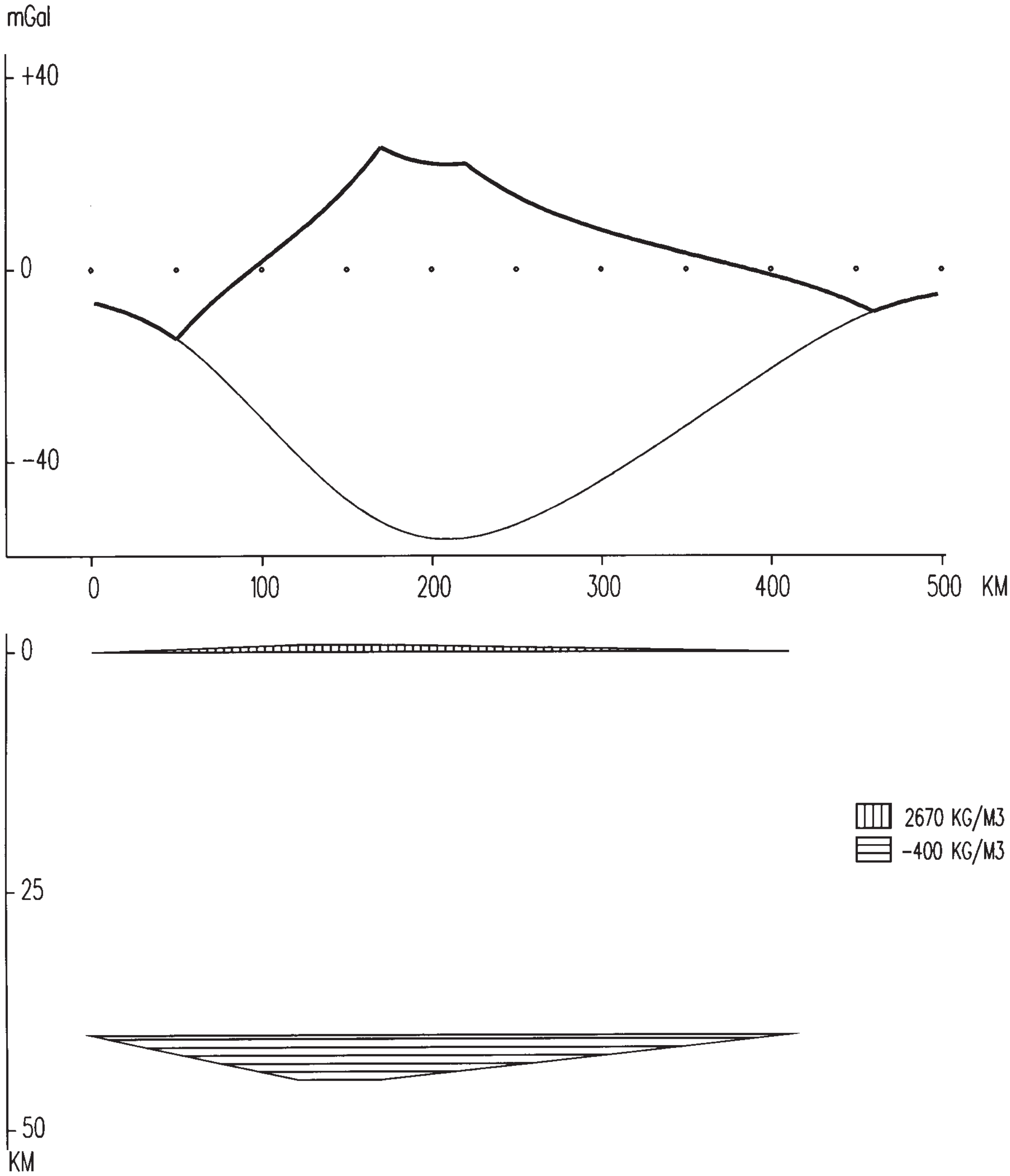

Fig. 5. A generalised gravity model across Scandinavian mountains from the northwestern coast of Norway to the east coast of Bothnian Bay in Finland with Sweden in the middle. The top part represents the topographic mass of the Caledonides with the following cornerpoints (0, 0.0) (120, 0.7) (170, 0.7) (410, 0.0) in km. The bottom part represents the compensating mass near the Moho with the following cornerpoints (0, -40), (120, -44.7) (170, -44.7) (410, -40.0). The thick curve is Free Air anomaly, and the thin curve is Bouguer anomaly. Under the mountains Bouguer anomaly is negative, almost -60 mGal, and Free Air anomaly positive, slightly over +20 mGal. Free Air anomaly exhibits a side minimum of about -10 mGal coinciding with the Bothnian Bay (at x=450 km). The side minimum runs alongside the Caledonian mountains in direction perpendicular to the modeling profile.

Density contrast in the upper mantle

Finland belongs to the Fennoscandian land uplift area. The mean values of Bouguer anomaly at the centre of land uplift are smaller than values on the outskirts of land uplift, although variations smaller in area but greater in intensity mask this areally large component of about 20 mGal in amplitude (*Balling*, 1980) and (*Kakkuri*, 1986). The half amplitude width of the component is about 600 km along the short axis. Obviously, the depth variations of the Moho as discussed above cannot explain this anomaly. Three other alternatives are analyzed by means of 3-D modelling along the short axis (roughly NW-SE) of the land uplift anomaly. The lowermost model represents an undulation of the "olivine-spinel" transition, which takes place at the depth of about 400 km. A density increase of about 300 kg/m^3 is associated with this boundary. The boundary is controlled by temperature so that a rise in temperature of 4.5 °C would cause the transition to migrate downwards by 1 km (*Bott*, 1971). In the model, the undulation is 2000 km long, 200 km wide and 14 km deep. There is no feasible reason for such an undulation, which corresponds to a temperature increase of 63 °C. Another difficulty is that the model does not produce the required amplitude with a small enough half amplitude width (see the thick curve in Fig. 6) due to its depth. It follows that the "spinel-perovskite" transition at the depth of about 680 km lies also too far from the surface. On the other hand, the lower lithosphere or tectosphere should have a slightly lower density than the mantle below it. The two uppermost models in Fig. 6, representing end members among possible solutions, produce the required gravity effect. The available literature (e.g. *Mueller et al.*, 1983) predicts that the lithosphere under Fennoscandia is about 180 km thick, while in surrounding regions its thickness is less than 100 km. The uppermost model in Fig. 6 assumes that there is no low density astenosphere. The middle model assumes that some kind of tectosphere immersed in the upper mantle is the prime cause. In conclusion, the most likely source for the gravity anomaly associated with the land uplift area has its source within the depth range of 100 to 300 km.

4. Conclusions

The Bouguer anomaly map of Finland has components which reflect the isostatic equilibrium and the earth's composition and structure from surface to the base of the lithosphere.

Distinct Bouguer anomaly maxima in Finland are caused by greenstones and other mafic volcanic rocks, ultramafic and mafic intrusions, diabase dykes and sills, carbonatite intrusions, granulites and upthrusted blocks from middle crust in general, schist and migmatite belts when surrounded by a gneissgranite basement or granites. It should not be forgotten, however, that locally some massive ore and industrial mineral deposits cause striking anomalies, as well.

Distinct Bouguer anomaly minima are caused by granite intrusions, rapakivi granites in particular and potassium rich granitoids in general, thick layers of quartzites in metamorphosed schist belts, granitegneiss basement domes or blocks in metamorphosed volcanosedimentary surroundings, preserved basins of Mesoproterozoic and younger sedimentary rocks, meteorite craters, strongly weathered bedrock and thick overburden. It should not be forgotten, however, that some kimberlite diatremes, graphite deposits and pegmatites cause discernible gravity minima, as well.

Mass anomalies due to crustal thickness variations are more or less compensated for by mass distributions within the crust. The gravity minimum associated with the land uplift has its source in the lithosphere - upper mantle transition zone.

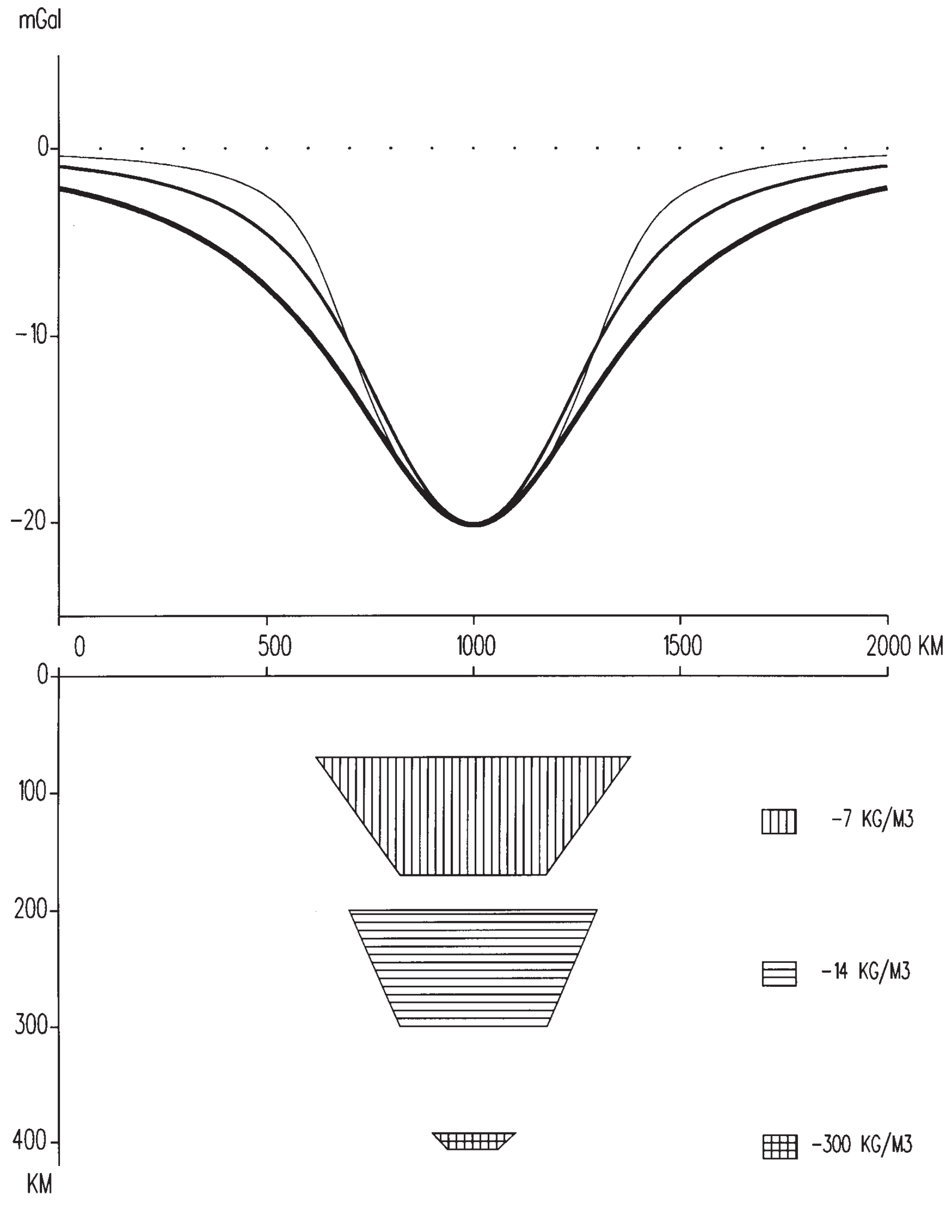

Fig. 6. Three-dimensional modelling of the gravity anomaly associated with the Fennoscandian land-uplift area along its short axis. The lowermost model and the thick curve represents an undulation in the "olivine-spinel" transition zone. This undulation is deemed unreasonable (for further discussion see the text). The uppermost two models are more likely candidates. They represent the density difference between the lower lithosphere or tectosphere and the upper mantle below. The length of the models in perpendicular direction to the profile is 2000 km. The end faces are inclined in the same manner as the sides. The models are fitted to a residual gravity anomaly that is -20 mGal in amplitude and 600 km in half amplitude width. To arrive at this residual anomaly, the effect due to the Caledonian mountains and the masses compensating them must be subtracted from the total anomaly.

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